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TECTONIC EVOLUTION OF SOUTH AMERICA

Edited by

U.G. Cordani
E.J. Milani
A. Thomaz Filho
D.A. Campos

Tectonic Evolution of South America

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Had I been present at the Creation,
I would have given some useful hints for
the better ordering of the universe.

Alfonso X

... no vestige of a beginning – no prospect of an end.

James Hutton

Prof. Fernando Flavio Marques de Almeida

This book is written in honor of
Fernando Flavio Marques de Almeida,
a man whose life has been devoted
to the Geology of Brazil.

Geoscientist and educator, throughout his career he traveled all over Brazil and produced a series of scientific publications on his research, the most outstanding of which is on the Origin and Evolution of the Brazilian Platform, an essential reference for Brazilian geoscientists on the understanding of the geotectonic framework of the Country. Furthermore, he was a pioneer in the application of mobilistic ideas to South America and to the South Atlantic Ocean, thus contributing to a better understanding of the processes of continental drift and plate tectonics.

Born in Rio de Janeiro in 1916, he graduated as a Civil Engineer from the Polytechnic School of São Paulo in 1938. It was there that Prof. Luis Flores de Moraes Rego, Senior Researcher in Mineralogy, Petrography and Geology deeply influenced his career. Following his graduation, he became Assistant Professor at the USP. In 1957 Dr. Almeida was appointed lecturer for his thesis on the Geology and Petrology of the Archipelago of Fernando de Noronha. Subsequently, in 1962, he was appointed to the chair of Petrology for his thesis on the Island of Trindade. These two theses represent a milestone in the field of Petrology and Geochemistry of oceanic volcanic islands. He remained at the USP for 35 years.

A part time professor, he joined the National Department of Mineral Production (DNPM) in 1956 where he remained until 1969. He set up the DNPM office in Goiânia, State of Goiás, and started a huge mapping programme in Central Brazil, with emphasis on the geology of the ultramafic complexes of Cana Brava, Niquelândia and Barro Alto. However, his strong desire to be fully involved with Education and Research caused him move back to the Geoscience Institute of the University of São Paulo in 1974, where he remained until 1978. He subsequently left the University and joined the Institute of Technological Research of the State of São Paulo (IPT). Between 1985 and 1990 he was granted a temporary leave of absence from the IPT, and was contracted by the Geoscience Institute of the State University of Campinas from which he retired in 1994.

Prof. Almeida is a member of the Brazilian Academy of Sciences (1951), and Vice-President of the Societé Géologique de France (1971). He was awarded the Admiral Alvaro Alberto Prize from the Brazilian Research Council (1986), and was nominated *Doctor honoris causa* by the State University of Campinas (1991). He received the Medal for Merit from the Federal Council of Engineering and Architecture (1995), and the Grand Cross of the National Order of Scientific Achievements from the Brazilian Government (1995).

Fernando de Almeida can be considered the most outstanding Brazilian geoscientist of the 20th century.

The Brazilian Federal Mines Department, the Departamento Nacional de Produção Mineral – DNPM has the honor to co-sponsor the publication of this volume on the Tectonic Evolution of South America. The publication of this volume coincides with the *31st International Geological Congress*, when for the first time, the largest event of the world's geological community takes place in a South American country.

Historically, International Geological congresses act as a catalyst for geological exploration and mining in those countries in which they are held, and this is a direct proof of the beneficial effects that stimulate both the scientific community and industry.

Presently, the DNPM plays a major role in the development of geological exploration, a mission inherited from the Serviço Geológico e Mineralógico do Brasil (Geological and Mineralogical Survey of Brazil). Since 1907, the DNPM has been responsible for the subsoil, and for the granting of exploration permits and mining concessions.

Knowledge of the subsoil has been the main objective of the DNPM Results have been very well presented in compendiums and in maps of geological synthesis, published under the auspices of this agency. These works include the Geology of Brazil by Avelino Ignacio de Oliveira and Othon Henry Leonardos, and the Geology of Brazil by Djalma Guimarães; in addition to a number of geological maps at scales less than 1: 1 000 000. This tradition extends in time from the map drawn up by John Casper Branner in 1919 to the Geological Map of South America, presently in the final stages of drafting by our Exploration Division.

Today, the DNPM is undergoing structural changes to become a National Mining Agency. This new federal body is to perform a role similar to that of a bureau of mines in many other countries. Another branch of the DNPM, the Museu de Ciências da Terra (Earth Sciences Museum) that houses a precious collection of mineral, rock, meteorite and fossil specimens will also be transformed into a more active institution in order to promote research and scientific studies on mineralogy, petrology and paleontology.

Ninety-three years of basic geological studies have made available an immense amount of information to the mining sector. Now, on the eve of the XXI century, the DNPM has the required knowledge and the skill to manage the Brazilian subsoil, utilizing the mineral endowment that Nature took billions of years to create, to the benefit of our society and future generations.

João R. Pimentel
Director-General
DNPM/Brazil

We have to go back to 1953 to find a comprehensive volume on the Geology of South America, when Jenks edited his *Handbook on South American Geology*, with the aid of several distinguished regional geologists of that time. Since then, the Earth Sciences have undergone a complete change, with major developments in such disciplines as Solid Earth Geophysics, Rock Geochemistry, Structural Geology, and above all the mobilistic "revolution" of the late 1960's, which led to the Global Tectonics theory.

In all South American countries, an immense set of new geological information has become available in recent times. Many good geological and tectonic maps of national character have been prepared in most countries, as well as a few syntheses at the regional or national level, such as that by Schobbenhaus et al.,[1] and that on the Andean tectonics by Gansser[2]. A few general continental geological maps were also published through the work of the Commission for the Geological Map of the World, but their written comments were usually very short and barely descriptive of the represented units. Moreover, many research articles were produced on the various aspects of the Geology of South America, but most of these were written in Spanish or Portuguese, and published in local journals, or as internal reports for geological organizations of the South American countries. As a result, much of that work was not easily available to the international geological community, and a comprehensive work synthesizing the updated geological information for South America remained lacking.

In order to overcome this situation, one of us (UGC), in the early eighties, envisaged the production of a volume similar to this book. It was to be prepared with the help of many co-authors, the great majority of them being distinguished and active geoscientists from different parts of South America; people that could easily make, based on their personal and direct experience, a good review of the sections under their responsibility.

This work was to be called *The Geotectonics of South America*. It would have stressed the regional geological evolution of the main tectonic units of the continent, and would have followed the mobilistic trends of that time. It would not include complete descriptions of geological, lithological or stratigraphic units, and was to be based on the evidence from all geological fields, supported by the most reliable geochronological information available.

Because of the emphasis to be placed on geochronological information and interpretation, the coordination of the volume was given to colleagues from the Geochronological Research Center of the University of São Paulo. This was done taking advantage of the special situation of that center, which was for several years the only age dating laboratory in Brazil. About one hundred co-authors from most South American countries, as well as some French and British colleagues, committed themselves to participate in the different chapters of the book. In general, they were active and participated jointly in some international programmes, such as the Geodynamics Project, the International Lithosphere Project and some projects related to the International Geological Correlation Programme (IGCP), such as the 120 (Andean Magmatic Evolution) and the 204 (Precambrian Evolution of the Amazonian Region).

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[1] Schobbenhaus, C. *et al.* (1984). *Geologia do Brasil*. DNPM, Brasília.

[2] Gansser, A. (1973). Facts and theories on the Andes. *J. Geol. Soc. London* 129, 93-131.

The work received the sponsorship of the Academia Brasileira de Ciências, for the main editorial part, and of the Departamento Nacional de Produção Mineral (DNPM), for the preparation of the geological illustrations, including a map of the tectonic framework for South America at the scale of 1:10 000 000. A few meetings of the regional coordinators were held at Rio de Janeiro, at the Academia Brasileira de Ciências, and several articles were in fact written, covering almost half of the planned content of the book. However, very regrettably, the editors had to abandon the project in the early 1990's, because it was not possible to assure that the remaining articles required to complete the volume would be received. The authors that had written the regional papers as initially planned were set free to submit their work to different journals, as discret publications.

Subsequently, an opportunity to resume the preparation of a synthesis on South American tectonics presented itself in 1996. The General Assembly of the 30th International Geological Congress (IGC) approved a proposal made by a consortium of South American countries to organize the next venue of the IGC, in Rio de Janeiro in the year 2000. A few months later, at Campos do Jordão, Brazil, the First South American Symposium on Geochronology and Isotope Geology was held, with the presence of many of those involved in the first attempt. At this meeting the delegates agreed enthusiastically to make a renewed effort to prepare a comprehensive volume on the tectonics of the continent, to be issued in time for the 31st IGC.

Timing could not have been better, because in recent years South America has seen the production of an enormous amount of work not only in Geotectonics, but also in all other fields of Geology, and especially in Geochronology. Radiometric methods that were not usually employed in the study of South American rocks, such as the Pb/Pb, the Sm/Nd, and the U/Pb method in multigrain zircon samples, but also in single crystals by SHRIMP, began to be used in several regional contexts. They provided excellent opportunities to improve the interpretations of the geological history of such regions, and therefore it was for the best a synthesis like the volume had to wait for a few years before publication.

This volume includes the work of about 70 authors, most of them from Brazil, but several from Argentina, Chile and France. It comprises 25 regional articles, all of them very up-to-date in terms of their geological context, and amounting to over 800 pages. The work of many of the authors was facilitated by the fact that they were committed to write a short version of their articles for *Episodes*, the official journal of the International Union of Geological Sciences. This was published in September 1999, as a special issue dedicated to the 31st International Geological Congress.

At the meeting in Campos do Jordão we were confirmed as the editors of the volume, which once more would received the sponsorship of the Academia Brasileira de Ciências, and the Departamento Nacional de Produção Mineral. Soon we realized that an endeavor of this magnitude should not be exposed to possibilities of failure, and therefore we decided to invite two additional co-editors, Edison J. Milani and Antonio Thomaz Filho, to act as the executive editors. We are absolutely sure that the cooperation of Milani and Thomaz Filho was crucial to the materialization of this volume, in time for presentation at the 31st IGC.

We wish to thank the support by the Academia Brasileira de Ciências and the Departamento Nacional de Produção Mineral.

Rio de Janeiro, August 2000

Umberto Giuseppe Cordani
Universidade de São Paulo
Academia Brasileira de Ciências

Diogenes de Almeida Campos
Departamento Nacional de Produção Mineral
Academia Brasileira de Ciências

First of all, we would like to extend our sincere thanks to Professor Umberto Giuseppe Cordani and Professor Diogenes de Almeida Campos of the Brazilian Academy of Sciences for their kind invitation to participate in this important project. The unconditional confidence placed in us and the complete autonomy over editorial policy that we had during this period were fundamental to the success of this major undertaking.

Between February 1999 when a formal invitation was made to the contributors to this volume and the date this went to press in August 2000, there has been a year and a half of intense editing procedures. However, the effort dedicated to this monumental task has been amply rewarded by the quality of the final product.

We wish to express our thanks to all the contributors to this volume, who have given the very best of their work and dedication to the geological science. Likewise, thanks are extended to all those who have contributed indirectly and specifically those geoscientists and publishing houses who have kindly granted authorization to use figures published previously in scientific journals. We are much indebted to our families, and the ones of all those involved in the successful completion of this task for their patience and understanding during very many weekends of absence while working on this project.

The dedication of the editorial support staff is also gratefully acknowledged. Dr. Marcus Waring di Valderano performed a key role in the production of this volume by editing the language and grammar of the contributions with initiative and efficiency, bearing in mind that English is not the native language of most of the contributors. Cristina Barbosa prepared the final graphics project with unquestionable good taste. Paulo César de Souza Rocha, Murilo Paes Lins, and Bianca Maria Rego Martins undertook the arduous task of standardizing the presentation of the text figures. Finally, thanks are extended to Maira Nobre who organized the format in which the bibliographic references are cited. The Executive Editors are much indebted for the support of Petróleo Brasileiro S.A. – PETROBRAS.

Considering its wide scope this volume is certainly a pioneer contribution in the divulgence of the geology of our continent, being an up-to-date synthesis of the tectonic knowledge of South America. Each chapter was submitted by one or more invited authors that are directly involved in the research of specific aspects of the general subject. It is important to state that the responsibility for the technical and scientific content of the contributions lies with the authors, since the editorial policy was restricted to format and presentation. Naturally, with the authors remain the full merits of their work.

It is the sincere wish of the Executive Editors that this volume, published on the occasion of the *XXXI International Geological Congress – Rio 2000*, will provide a useful reference to the international community of geoscientists.

Rio de Janeiro, August 2000

Edison José Milani
Petróleo Brasileiro S.A.
Visiting Professor at Universidade do Estado do Rio de Janeiro

Antonio Thomaz Filho
Universidade do Estado do Rio de Janeiro

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Cratonic South America

"The South American Platform forms the nucleus or core of South America. It covers an area of about 15 million square kilometres, some 40% of which is exposed in three Precambrian shields: Guiana, Central Brazil (or Guaporé) and Atlantic shields. About 80% of the continental crust exposed in these shields was formed during the Paleoproterozoic and the Archean. At the end of the Neoproterozoic the South American Platform consisted of several plates or independent cratonic nuclei, most of which were still aggregated to their African counterparts... The most important cratons of the South American Platform are the Amazonian, São Francisco, and Rio de La Plata, in addition to smaller continental fragments. The basement of these cratons consists essentially of medium and high-grade metamorphic rocks, in addition to associations of the granite-greenstone belt type and numerous granitoid plutons. Fragments of Archean medium to high-grade metamorphites occur as inliers in the Proterozoic mobile belts".

(Dardenne and Schobbenhaus, this volume)

CORDANI, U.G. / MILANI, E.J. / THOMAZ FILHO, A./CAMPOS, D.A.
TECTONIC EVOLUTION OF SOUTH AMERICA, P. 19-40 / RIO DE JANEIRO, 2000

Crustal Evolution of the South American Platform

Umberto G. Cordani, Kei Sato, Wilson Teixeira, Colombo C. G. Tassinari, and Miguel A. S. Basei

This article aims to produce a synthesis of the geotectonic evolution of the South American Platform, putting together the available geochronological data and that obtained by isotope geochemistry on the tectonic provinces of that huge geotectonic unit. The Andean Belt is a major tectonic domain of the continental crust of South America and the authors have preferred not to include the Andean region in this synthesis on the continental growth of South America because this region is covered in a number of contributions included in this volume.

In this work, the authors give their views on how the crustal material of each of the major Precambrian tectonic provinces of the South American Platform was formed, either directly by mantle-crust differentiation processes, or by transformation of pre-existing continental crust. The time frame will be taken into account, as well as the major tectonic events that affected them. A mobilistic approach will be followed, throughout geological time, considering that the Earth's internal heat is responsible for the tectonic regime of the planet, which has evolved from a "permobile type" during primitive times in the Archean, to the present "platform type" governed by plate tectonics. The main ideas related to our interpretation can be synthesised in the paragraphs that follows:

1 - Primary "basaltic" magma is formed by partial fusion of the upper mantle, and its products make up the bulk of the oceanic lithosphere. After some tens of million years this material is recycled back to the mantle. The subducted slab may also suffer partial fusion, producing secondary magma of "granitoid" composition, which migrates towards the upper levels of the crust. Such magmatic arcs are incorporated to the crust, either as island arcs in oceanic domains, or within a pre-existing continent. In addition, some granitoid material may also be formed directly from the mantle, by heating of the mantle wedge overlying the subduction zone. All these processes produce continental growth by addition of material related to mantle-continental crust geochemical differentiation, and the products of it are calc-alkaline granitoid plutons (tonalite-trondhjemite-granodiorite) present throughout the Earth's history. Melting and reworking of the pre-existing crust can also form granitoid magma, if appropriate heat sources are available, and this process does not require continental growth.

2 - Juvenile continental crust, originating from the mantle in the way indicated above, was formed in great amounts during the Archean under the permobile tectonic regime, giving birth to small crustal units of continental character. By successive collisions, this primitive continental material formed increasingly larger masses, some of which with sizes of the order of the present continents. The fusion of such continental units, resulting from the disappearance of the oceanic floor originally occurring between them, and successive episodes of collision along orogenic sutures, produced supercontinents. After some time such extremely large units became unstable, and were disrupted during episodes of continental dispersion in the so-called supercontinent cycle.

3 - The composition of radiogenic isotopes produced by radioactive parents with very long half-lives, such as Sr, Nd, Pb, Hf, and Os, are mainly used to constrain the crustal evolution of geotectonic units. In the case of the Rb/Sr and the Sm/Nd systems, a rough comparison between them shows that both parent-daughter pairs are strongly fractionated during the complex processes, which form "juvenile" continental crust from primary mantle sources. However, in a clear contrast to the Rb/Sr system, the subsequent crustal processes normally produce only minor or even negligible changes in the Sm/Nd ratio, a property that has made the Nd isotope studies a very powerful tool in defining continental crustal provinces. Granitoid rocks, in the broader sense, are the main constituents of the continental crust. They may be formed through many different petrogenetic processes, and measurements of their isotope systematics can be employed as tracers for their origin and for defining the character of their source materials.

In this article, and dealing especially with granitoid rocks, we will describe the crustal evolution of the South American Continent, in terms of the formation of its crust, and its development through time. For our interpretation, magmatic suites ranging from granodiorite and tonalite to true granite were employed whenever possible, as well as granitoid rocks from medium to high-grade regional metamorphic terranes. These are the prevalent units of the gneiss-migmatite-granulite complexes so common in shield areas, and South America is no exception. Cordani *et al.* (1988) have already produced a synthesis on this subject, employing the Sr isotope composition of granitoid rocks, obtained from more than 10 000 individual measurements on samples collected all over the South American Platform. Such unit is the large and tectonically stable cratonic area, which served as the foreland for the development of the

Andean Belt in the Phanerozoic. The Nd isotope composition determinations, supported by Rb/Sr and U/Pb radiometric measurements, obtained recently in several hundred rock samples by Sato (1998), will be discussed in this paper. Cordani and Sato (1999) have already presented a preliminary article on this subject. More detailed descriptions on the geotectonic evolution for each of the tectonic provinces outlined here will be given by different authors in the chapters that follow in this volume.

Tectonic Evolution of the South American Platform

As widely known, some sort of mobilistic regimes similar to plate tectonics acted throughout the entire Earth's history, when different parts of South America were involved in the formation and break-up of supercontinents such as Rodinia, Gondwana and Pangea. Moreover, magmatism associated with mantle-continental crust differentiation processes was common during the Archean, when TTG's suites were prevalent. Two main petrogenetic processes can be used to explain the very large volumes of granitoid rocks occurring on the South American Platform as well as in other parts of the Earth:

1 - Formation of juvenile granitoid rocks within magmatic arcs, in association with subduction of oceanic lithosphere. These may have been derived from different types of magma source, including the mantle wedge above the subduction zone.

2 - Formation of granitoid rock from pre-existing crustal protoliths, by partial or complete melting within the continental crust. These may occur in association with both orogenic belts and intraplate magmatism; in this case by melting of the lower crust when heated by underplated primary basaltic liquids.

Figure 1 shows the main geotectonic provinces of South America. In this figure, the Andean Belt is indicated, and the Neoproterozoic geotectonic situation of the South American Platform is shown. The large Amazonian and São Francisco cratons are shown, as well as the small Rio de La Plata, São Luís and Luiz Alves cratonic fragments and the mobile belts associated with the Neoproterozoic Brasiliano-Pan African orogenic cycle. These correspond to the Paraguai-Araguaia-Tocantins belts, marginal to the Amazonian Craton; the Brasília and Araçuaí belts, marginal to the São Francisco Craton; the Dom Feliciano Belt (DF), marginal to the Rio de La Plata and Luiz Alves cratonic fragments; the Borborema Province (B) in northeastern Brazil; the Tocantins Province (T) in central Brazil; and the Mantiqueira Province (M) in eastern and southeastern Brazil. In Figure 1 the tectonic provinces of the Amazonian Craton were also displayed, corresponding to the Central Amazonian (CA), Maroni-Itacaiúnas (MI), Ventuari-Tapajós (VT), Rio Negro-Juruena (RNJ), Rondonian-San Ignacio (RO) and Sunsás (SS).

Only the broad outline of such tectonic units is included. It is beyond the scope of this article to describe their specific geological setting, and the reader is invited to consult the chapters that follow in this volume, where such matters are considered with more detail. However, if additional geological information is required, most of these units or their equivalents were adequately described in the reference work of Schobbenhaus *et al.* (1984), that accompanies the *Geological Map of Brazil* at 2 500 000 scale.

Figure 1 also shows the Andean Belt, and the approximate extent of the influence of the Phanerozoic orogenic events. Within the Andes, black dots or small irregular marks indicate the areas of outcrop of rocks with undisputed or suspected Precambrian age. The Andean Belt is a very large area, which has acted as a continental margin at least since the Neoproterozoic and Early Paleozoic when the Gondwana Supercontinent was formed.

Many Precambrian basement inliers are included in the Andean Belt, as shown in Figure 1. However, a direct age correlation between such inliers with the tectonic provinces occurring to the E is not possible because of the large horizontal displacement these underwent during the deformation associated with the Phanerozoic orogenies. These include the various tectonic pulses of the Meso-Cenozoic Andean Cycle; those of the Late Paleozoic Gondwanic (Hercynian) Cycle; and those of the Famatinian Cycle (pre-Hercynian deformation) observed in a few regions of the Andean Cordillera. In a general way, it seems that the Precambrian basement inliers within the Andean realm mainly show Neoproterozoic radiometric ages. Mesoproterozoic and Paleoproterozoic rocks are rare, and rocks of Archean age have not yet been found.

The contributions on the Andean domain (this volume) show the great complexity of the region and describe the active geological processes, which occur nowadays, due to the present interaction of the oceanic Nazca Plate with continental South America. Moreover, we know that the region has been orogenically active since the Early Paleozoic, and for more than 500 million years enormous volumes of magmatic rock were added to South America during several orogenic pulses forming calc-alkaline granitoid suites of the "Cordilleran"-type. Many of these rocks yield a juvenile signature. However, rocks formed by crustal reworking are also common. The Andean Belt has always been in a peripheral orogenic region, not connected with large continental collisions. Instead, collisions of volcanic island arcs and related microcontinental blocks have been more or less continuous throughout Phanerozoic times.

Some thought will now be given to the position of South America within the more recent supercontinent reconstructions, and the relative position of some of its cratonic fragments according to previous configurations.

Pangea is very well defined. The geological links between South America and Africa were established long ago (Wegener, 1912) as well as the age correlation of the provinces within both continents (West-African and São Luiz cratons, São Francisco and Congo cratons, Borborema and Nigerian provinces, etc.; Cordani and Torquato, 1981) which is well constrained. Pangea was formed by the fusion of Laurasia, Gondwana, Siberia, and other smaller continental masses, during the so-called Hercynian Orogenic Cycle, from about 360 to 270 Ma ago. This supercontinent, in existence throughout Permian and Triassic times, underwent break-up from the early Jurassic, about 180 Ma ago, when the Atlantic, Indian and

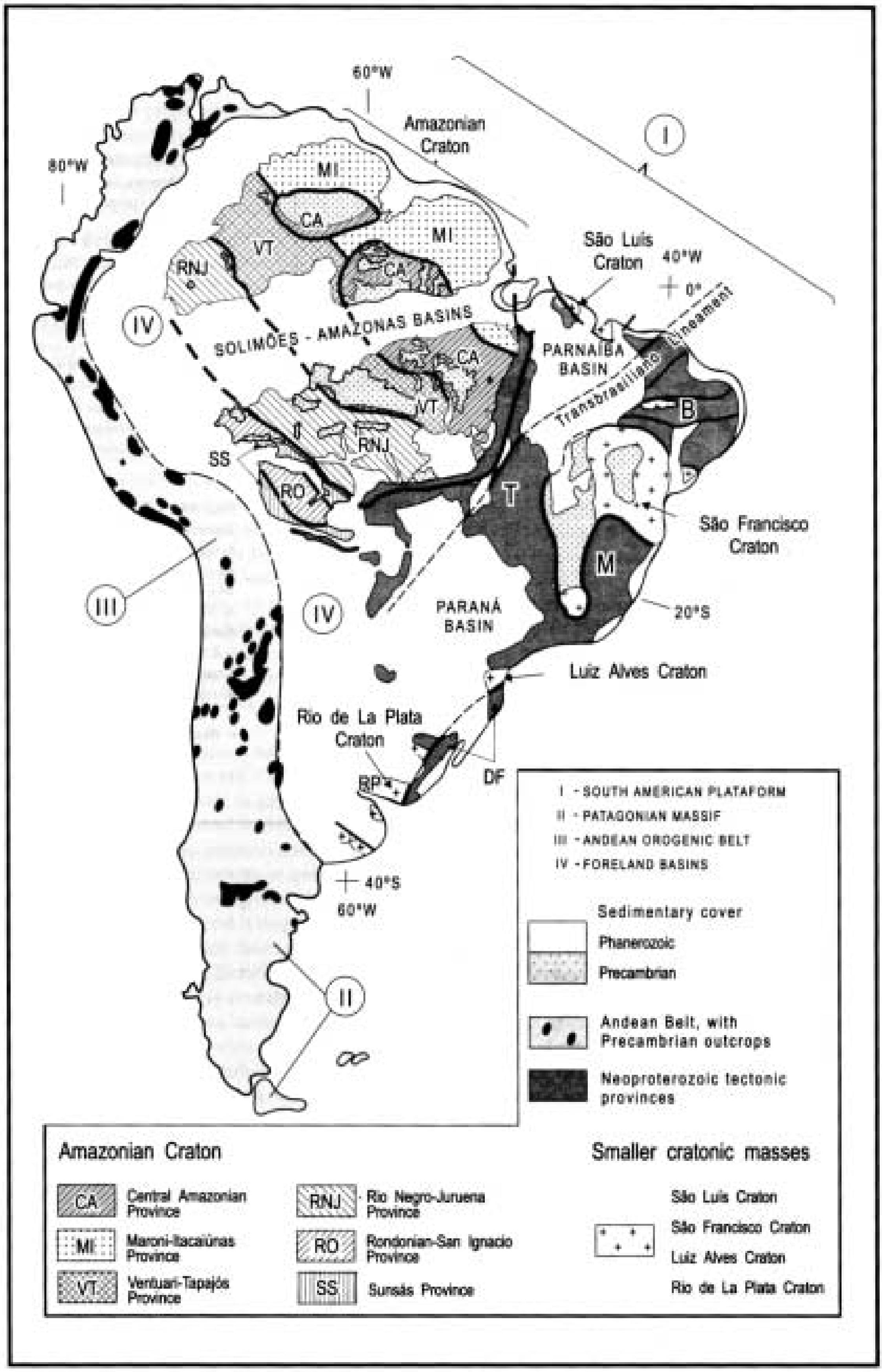


FIGURE 1 – Geotectonic provinces of South America.
SF = São Francisco Craton; CA = Central Amazonian Province; MI = Maroni-Itacaiúnas Province; VT = Ventuari-Tapajós Province; RNJ = Rio Negro-Juruena Province; RO = Rondonian-San Ignácio Province; SS = Sunsás Province; B = Borborema Province; T = Tocantins Province; M = Northern sector of Mantiqueira Province; DF = Dom Feliciano Belt; SL = São Luís Cratonic Fragment; LA = Luiz Alves Cratonic Fragment; RP = Rio de La Plata Craton.

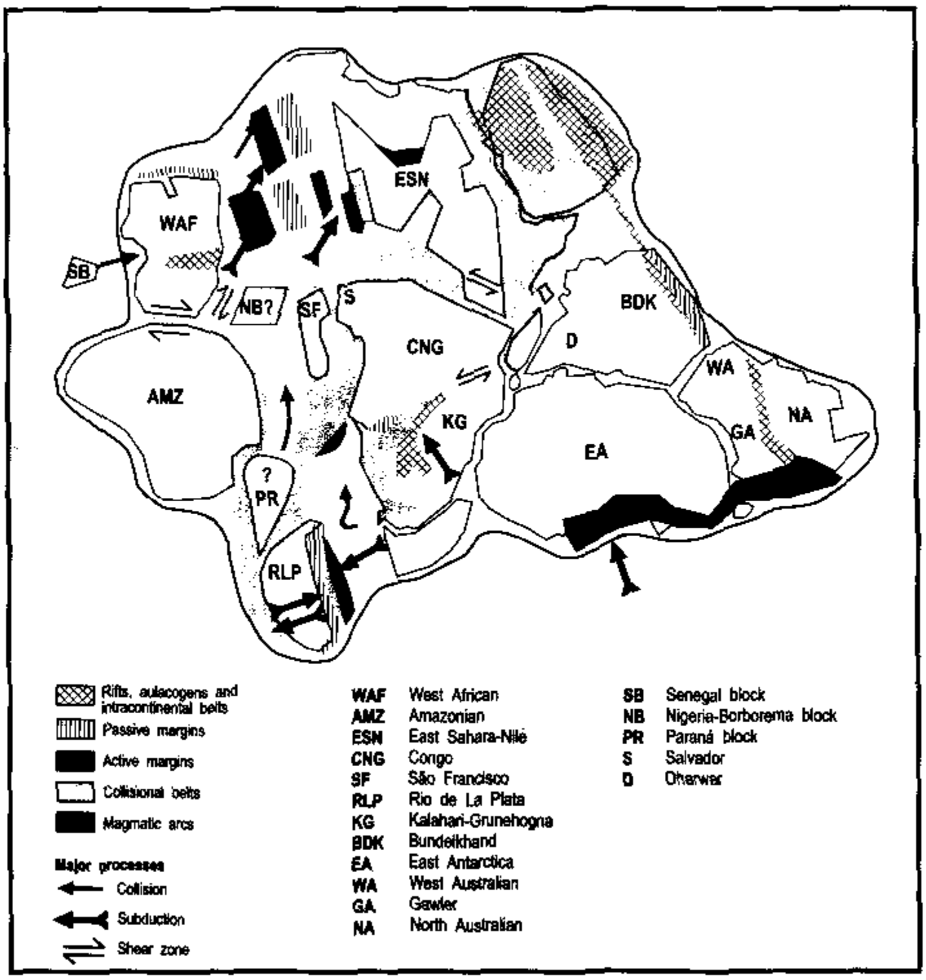
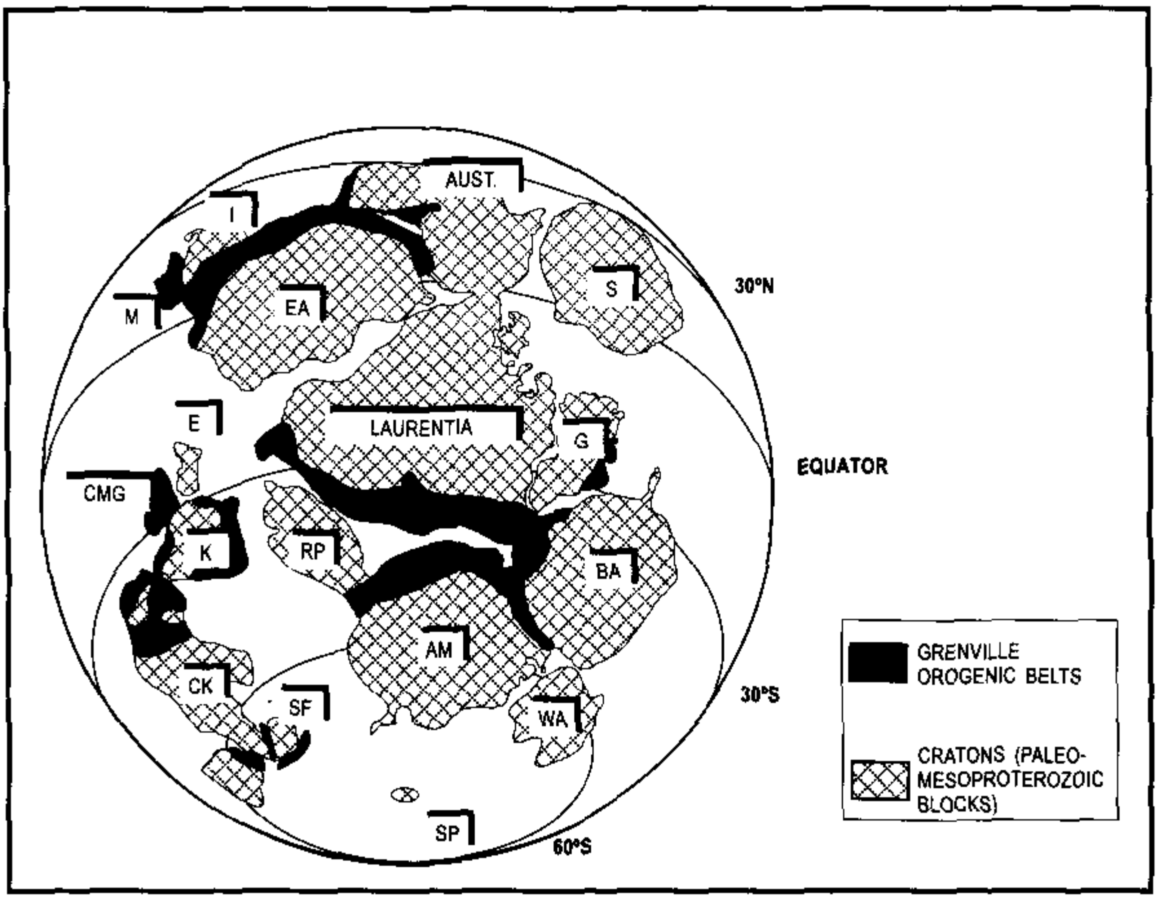


FIGURE 2 - The Brasiliano-Pan-African Orogeny and the assembly of Gondwana (adapted from Unrug, 1996).
AMZ = Amazonian; PR = Paraná Block; RLP = Rio de La Plata; WAF = West African; SB = Senegal Block; NB = Nigeria - Borborema Block; SF = São Francisco; ESN = East Sahara-Nile; CNG = Congo; KG = Kalahari-Grunehogna; BDK = Bundelkhand; EA = East Antarctica; WA = West Australian; NA = North Australian; GA = Gawler; S = Salvador; D = Dharwar

FIGURE 3 - The Rodinia Supercontinent, a reconstruction based on paleomagnetic measurements. (adapted from Weil et al., 1998).
AUST = Australia; EA = East Antarctica; S = Siberia; E = CMG = K = Kalahari CK = Congo-Kasai; SF = São Francisco; RP = Rio de La Plata; AM = Amazonian G = Greenland; BA = Baltica; WA = West African



Australian-Antarctic oceans were formed, and the resulting continental fragments dispersed and drifted to their present geographical positions. For South America, the Hercynian Cycle produced small amount of crustal accretion within the peripheral Andean Belt, and a fusion of a small continental mass, Patagonia, with the main part of the South American Platform.

Prior to the assembly of Pangea, during the early Paleozoic, some less important episodes of continental accretion are described, always within the realm of the Andes, such as the magmatic arcs related to the Famatinian Cycle in northern Argentina.

The position of South America within Gondwana is also well established, since this Neoproterozoic supercontinent was already accepted as being one of the main parts of Pangea. Gondwana was formed by the agglutination of smaller continental masses, which were involved in multiple and successive collisions, starting about 750 Ma ago, but developing mainly between 650 and 530 Ma, and coinciding with the main orogenic phase of the Brasiliano-Pan African Cycle. Brito Neves and Cordani (1991) described the fusion of the various cratonic nuclei involved in that cycle, making up most of western Gondwana. For South America, we are dealing with the Amazonian, São Francisco, São Luiz, Rio de La Plata cratons, and for Africa we are dealing with the West African, Congo and Kalahari cratons. Moreover, in this process of agglutination, several microcontinents were also involved, and Figure 2 shows a possible reconstruction of Gondwana, just after the termination of the Brasiliano-Pan African Orogenic Cycle.

The fusion of the above-mentioned cratonic nuclei to form Gondwana was obtained through successive continental collisions. Such large tectonic events dismembered many continental shelves in the colliding plates, and produced folded mountain belts and regional metamorphism and several interior orogenies in South America and Africa, where thrusting, crustal thickening and widespread deformation occurred in Neoproterozoic times. This is the geotectonic situation shown in both Figures 1 and 2, where the Borborema, Tocantins and Mantiqueira provinces are the main areas severely affected by the Brasiliano-Pan African Orogeny, indicating the sites of the major continental collisions within South America. In Africa, similar scars related to such interior orogenies are present within the Trans-Saharan, the Mauritanide-Rockelide, the West Congo and the Damaran-Katangan belts. By the time Gondwana appeared, the present western part of South America would have corresponded to a rifted border formed by the break-up of a former supercontinent, Rodinia, a passive margin ready to be affected by the marginal Pre-Andean and Andean orogenies during the Phanerozoic.

The position of the individual cratonic masses, which since the latest Neoproterozoic can be i dentified as the main constituents of Gondwana, Pangea, and present South America, is very dubious for earlier times. In the case of Rodinia, the supercontinent that is supposed to have existed at least between 1.0 Ga and 800 Ma, several reconstructions are given; for instance Hoffman (1991), Dalziel (1997), Evans (1998), and D' Agrella (1999).

It seems that for the reconstruction of Rodinia, there are enough correlation indicators to constrain the relative position of Laurentia, Baltica, Australia and eastern Antarctica (Kalstrom *et al.*, 1999), but the position of the other pertinent cratonic nuclei, which later on were critical to the make up of western Gondwana, is far from resolved. We indicate here in Figure 3, adapted from Weil *et al.* (1998), one possible Rodinia reconstruction, mainly based in paleomagnetic determinations, where the position of Amazonia, West Africa, São Francisco-Congo, Rio de La Plata and Kalahari is outlined. In our opinion this reconstruction of Rodinia suffers from an intrinsic diffect in that the cratonic masses that eventually need to came together to make up Gondwana at about 650 Ma are very far apart, and a great deal of oceanic lithosphere has to be destroyed along many subduction zones. However, magmatic arcs of that age within the Brasiliano-Pan African belts of Gondwana are presented only in minor amounts in some restricted regions.

Another difficulty with the concept of Rodinia is its time of break-up. Hoffman (1991) argues for the stability of that supercontinent until about 700 Ma when break-up would have occurred. Within the Brasiliano-Pan African belts in South America and Africa the main period of tectonomagmatism occurred between 650 and 530 Ma, but as will be seen later there were earlier precursors active along similar structural trends. These are related to continental collision in many of the Neoproterozoic tectonic provinces, dated with less precision at around 800-750 Ma, but some of them with indications of even earlier ages (Brito Neves *et al.*, 1999). In such cases, these collisional events related to some precocious events of the Brasiliano Orogeny must be considered as the first precursors of the agglutination of Gondwana. Furthermore, supposing that Rodinia indeed existed as a single individual mass, its break-up must have occurred much earlier than has been suggested. The Cariris Velhos Orogeny of about 1.0 Ga to 900 Ma within the Borborema Province of northeastern Brazil, and perhaps the orogenic event of about 1.0 Ga within the Mozambique Belt in East Africa (Pinna *et al.*, 1993) may be crucial in this respect. Were they the last collisional events in the formation of Rodinia, or were they the first orogenic events in the formation of Gondwana? The existence of very large ocean floors such as the Goianides Ocean (Pimentel *et al.*, 1997) with ages of more than 900 Ma, and the evidence of large dyke swarms over the eastern part of the São Francisco Craton with ages as old as 1.1 - 1.0 Ga (D' Agrella *et al.*, 1990) may indicate that the break-up of Rodinia could have started, at least in some places, not long after its formation in Grenville times.

The Grenville Orogenic Cycle, at about 1.1 - 1.0 Ga, is indicated as being responsible for the formation of Rodinia. For the continental masses of the present Southern Hemisphere, collisions are indicated at the southwestern end of the Amazonian Craton. Examples are the Sunsás Belt within the Rondonian Province (Sadowski and Bettencourt, 1996) and the Namaqua-Natal Belt, cutting across the Kalahari Craton (Hartnady *et al.*, 1985). However, it must be pointed out that collisional scars with similar tectonic trends but with much older ages are reported. These include the San Ignacio Orogenic Belt in Rondonia, and the Kibaran Belt in Central Africa, suggesting that the onset of the agglutination processes which eventually formed Rodinia may have started around 1.4 - 1.35 Ga, or well into the Mesoproterozoic.

For earlier times, only speculation is possible. There are indications that within the Amazonian and São Francisco cratons there occurred large rifts and disruptive structures associated with the development of cratonic volcano-sedimentary basins and aulacogens (Espinhaço, Roraima, Beneficiente, etc.), with ages roughly between 1.8 and 1.5 Ga (Teixeira *et al.*, 1989; Teixeira and Figueiredo, 1991). Moreover, in the same period, a great deal of continental crust was formed within the Amazonian Craton by a series of juvenile magmatic arcs which demonstrate the existence of a large region with oceanic lithosphere. In addition, numerous bodies with similar ages, formed by anorogenic magmatic processes, are also described, suggesting the possibility of their association with a major process of break-up of a Paleoproterozoic supercontinent.

The total amount of material forming continental-size masses at the surface of the Earth in Archean and Paleoproterozoic times was surely smaller than in later times. From the geochronological record (Cordani *et al.*, 1973; Cordani and Brito Neves, 1982; Cordani *et al.*, 1988; Teixeira and Figueiredo, 1991) it is apparent that the Paleoproterozoic was probably the most important period for the formation of continental crust. In South America, this is related to the widespread Transamazonian Orogenic Cycle (defined by Hurley *et al.*, 1967), when a large amount of juvenile continental crust was formed. It did not exist earlier, meaning that the cratonic masses were smaller, and their shapes different. The Transamazonian collisional belts, the age of which is roughly 2.1 - 1.9 Ga, may indicate a period of supercontinent formation (Atlantica?) the break-up of which produced the cratonic fragments that later agglutinated to form Rodinia.

During Archean times, only a smaller amount of continental crust is attributed to South America. Most of this material is juvenile and yields late Archean ages. Because of this, we suggest the action of several continental agglutination processes, possibly independent of each other, leading to a few granite-greenstone terranes, limited in size, in the Carajás District; in central Bahia; in the Iron Quadrangle; and in the Crixás region of Goiás. Some smaller areas with Archean age and medium to high grade metamorphism are also described as basement inliers or allochtonous terranes within the Proterozoic belts, reworked to varying degrees by the younger orogenic events.

Geochemical Significance of Sr and Nd Data

In the case of strontium, the initial $^{87}Sr/^{86}Sr$ ratio for a suite of cogenetic granitoid rocks is the main indicator for source material. In the work by Cordani *et al.* (1988), this indicator was used together with some other geological criteria (stratigraphic, petrogenetic, structural), in order to estimate the relative amount of juvenile continental crust accreted during each interval of geological time for the area of the Brazilian Shield. Curve 1 in Figure 17 shows the fraction of continental growth (or survival) with time, according to Cordani *et al.* (1988).

Sato (1998) carried out a comprehensive Sm/Nd study for the entire South American Platform. Nd isotope composition and Sm/Nd model ages were obtained by that author for several hundred granitoid rocks, many of which had already been analysed for Sr isotopes at the CPGeo-USP. The association of the Sm/Nd measurements with the already available geochronological data obtained by other methods permitted an improved interpretation for the origin of the granitoid rocks under examination.

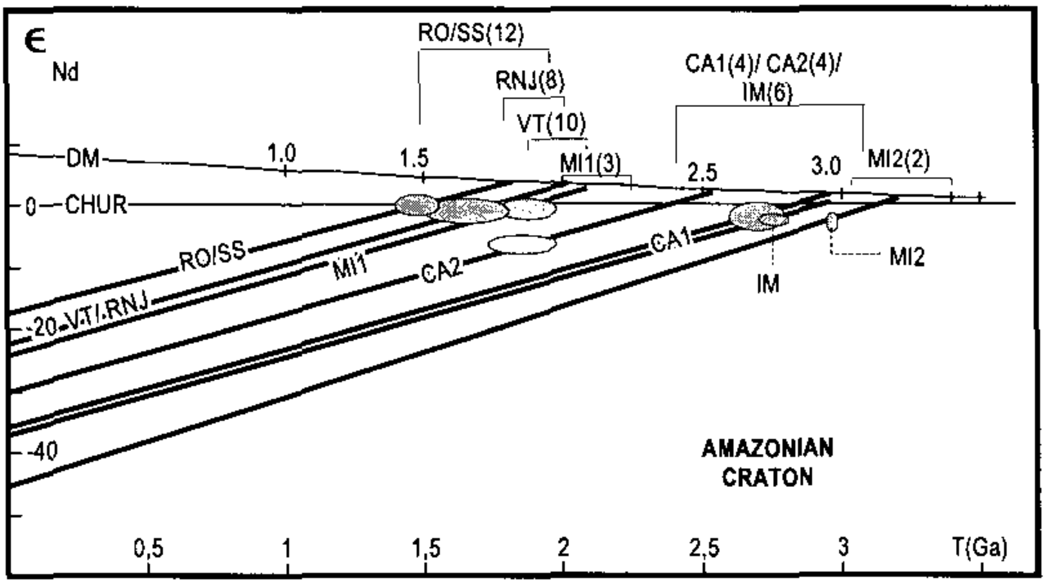
We are aware of the limitations of the Sm/Nd model ages for the interpretation of the crustal evolution of a given domain. Sm/Nd T_{DM} model ages are the most used in such interpretations. They are related to the evolution of the mantle during geological time admitting episodes of fractionation associated with the extraction of basaltic magma, with the residual material within the mantle sources becoming increasingly enriched in the Sm/Nd ratio, but depleted in incompatible elements (DM = depleted mantle). Taking into account both the natural variations in mantle material, demonstrated by the existence of geochemically distinct reservoirs, such as DMM, HIMU, EM1, EM2, and others, and also the possibilities of mixtures between such different reservoirs and other crustal sources, the Sm/Nd T_{DM} model ages must be interpreted very carefully. At any rate, the depleted mantle model is employed as a reasonable first approximation, because oceanic lithosphere was produced throughout geological time. It is known to be transient, recycling back to the mantle along b-subduction zones, but part of its material remains near the surface of the Earth as continental crust. For a comprehensive account on these complex matters, the authors recommend the work of De Paolo (1980, 1988).

Sm/Nd T_{DM} model ages had been considered to be "crust-formation ages", since they are obtained calculating the time when a given sample had an isotope composition identical to the depleted mantle, presumed to be its ultimate source. Their validity depends on at least two assumptions: 1 - That only a short time elapsed between the formation of the mantle-derived magma and the final emplacement of the differentiated material in the continental crust, with Sm to Nd fractionation and acquisition of a continental Sm/Nd ratio, and 2 - That the Sm/Nd ratio of the sample was not modified by subsequent geological events. Because such assumptions may be valid only in a limited number of cases, the Sm/Nd systematics usually provide an estimate of the average time that the material in each measured sample has been resident in the continental crust (Arndt and Goldstein, 1987).

The use of special interpretative diagrams and the association of the Sm/Nd measurements with other geochronological data allows one to consider whether the granitoid rocks under investigation could be considered "juvenile", and formed by mantle-continental crust differentiation processes as described in the previous paragraphs, or whether these have been "reworked" by melting of older crustal protoliths. Juvenile material is indicated by concordant (or slightly discordant) Rb/Sr (or U/Pb) and Sm/Nd apparent age values, as well as positive (or slightly negative) $\varepsilon_{Nd(T)}$ values and low $^{87}Sr/^{86}Sr$ initial ratios.

In this work, emphasis will be given to the Nd isotope data obtained by Sato (1998). The figures that follow (4, 6, 8, 10, 12 and 14) are the ones reported by Cordani and Sato (1999), slightly modified as to include some additional data

FIGURE 4 – Diagram of Nd isotope evolution for the Amazonian Craton. Central Amazonian (CA1 and CA2); Imataca (IM); Maroni-Itacaiúnas (MI1 and MI2); Ventuari-Tapajós (VT); Rio Negro-Juruena (RNJ); Rondonian-San Ignacio (RO); Sunsás (SS).



obtained during 1999. They show the evolution during geological time of the Nd isotope composition of different types of granitoid rock, representative of important units within the various geotectonic domains of the South American Platform. All diagrams include both the mantle evolution curve (CHUR, where ε_{Nd}= zero) and the depleted mantle (DM) evolution curve. Oblique lines plotted across the areas of the diagrams are related to different groups of granitoid rock, characterized by Nd isotope determinations which display a reasonably coherent pattern. Such lines show the average tendency of each group. In the figures, the average present-day $\varepsilon_{Nd(zero)}$ value, always negative, is plotted at the left end of each straight line.

The slopes of the oblique lines correspond to the general trends of the samples analysed. In all cases the $^{147}Sm/^{144}Nd$ ratios related to such trends are close to 0.11, typical of the products of the so-called single-stage differentiation from mantle to continental crust. Samples with $^{147}Sm/^{144}Nd$ outside the range 0.088 and 0.125 were not included, in order to avoid systems almost certainly not formed in a "single stage" event. In all of the lines, radiometric ages of each unit, obtained by the Rb/Sr or U/Pb methods, were indicated by small circles, or ellipses, drawn to represent roughly the spread in age and initial $\varepsilon_{Nd(T)}$ values obtained in the analytical work. Finally, for each of the units, the spread in the calculated Sm/Nd T_{DM} model ages was also indicated at the top of each diagram, close to the DM evolution line, together with the actual number of samples belonging to each group.

Summary of the Isotopic Constraints on Crustal Evolution

Hereafter, the crustal evolution of the main geotectonic provinces of the South American Platform will be reviewed with the aid of the geotectonic diagrams for each of the provinces, and the respective Nd isotope evolution diagrams.

Amazonian Craton

The Amazonian Craton is one of the largest cratonic areas of the world, situated in the northern part of South America. Teixeira *et al.* (1989), as well as Tassinari and Macambira (1999) reviewed its geotectonic evolution, and summarized the geochronological control. The craton can be subdivided into six major domains, very coherent in their geochronological patterns, the boundaries of which are shown in Figure 1. Geochronology has been crucial for the understanding of the crustal evolution of the entire region, because the general geological knowledge remains at reconnaissance level due to the abundant soil and vegetation cover.

The older parts of the craton are the Carajás-Iricoumé and the Roraima blocks, which belong to the Central Amazonian Province (CA in Figure 1). They correspond to regions with demonstrated or assumed Archean basement, covered by Proterozoic volcano-sedimentary cratonic sequences. Paleoproterozoic orogenies are typically developed at the border zones of this province, and their products are well represented in the northern and northeastern sides by the Maroni-Itacaiúnas Mobile Belt (MI in Figure 1). This is the site of amalgamation of at least three large continental masses at around 2.2 to 1.9 Ga: the Central Amazonian Province, the West African Craton and the Imataca Terrane. Part of the Maroni-Itacaiúnas Belt appears to have evolved from mantle-derived material, whereas other parts are likely to be recycled older crust. The Imataca Terrane is a possibly allochtonous crustal fragment formed by Archean high-grade metamorphic rocks, which suffered the effects of the Transamazonian Orogeny.

On the southwestern side of the Central Amazonian Province a Paleo-Mesoproterozoic continental crust began to be accreted to the continent through a series of successive magmatic arcs, producing the juvenile material of the Ventuari-Tapajós (2.0 to 1.9 Ga) and the Rio Negro-Juruena (1.8 to 1.5 Ga) tectonic provinces. (VT and RNJ in Figure 1). It is noteworthy that while continents collided in one side, during the so-called Transamazonian Orogeny, producing

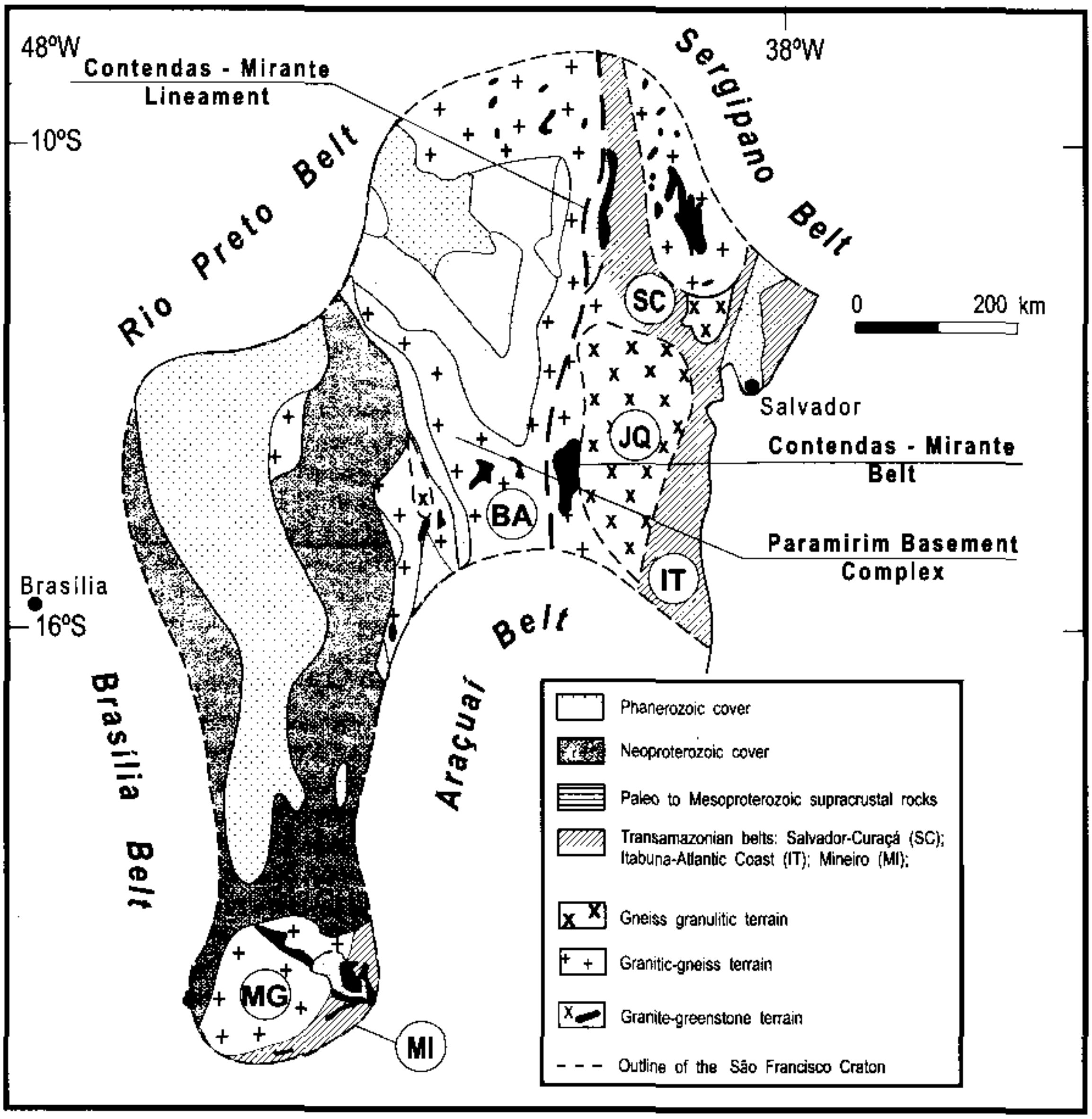


FIGURE 5 - The São Francisco Craton and its main tectonic provinces. BA = Gavião Block and related terranes; MG = Metamorphic complexes of Minas Gerais; MI = Mineiro Belt; SC = Salvador-Curaçá Belt; IT = Itabuna-Atlantic Coastal Belt; JQ = Jequié Block.

the Maroni-Itacaiúnas Belt, juvenile magmatic arcs started to be added at the opposite side of the Central Amazonian Province, in accretionary processes that produced a very large continental mass.

The southwestern part of the Amazonian Craton is relatively better known due to the work by Litherland and co-workers (1986) in the Precambrian of Bolivia, and by several projects in the states of Rondonia and Mato Grosso in Brazil (Tassinari *et al.*, 1996, Bettencourt *et al.*, 1999). This area is described as multi-orogenic, formed between 1.8 and 1.0 Ga. Its older rocks are the aforementioned granitoid plutons of the Rio Negro-Juruena Province, produced by successive subduction of oceanic lithosphere. Sadowski and Bettencourt (1996) proposed a continental collision occurring at about 1.4 -1.3 Ga, related to the San Ignacio Orogeny, typical of the Rondonian-San Ignacio Province (RO in Figure 1). This continental collision may have been between the Amazonian Craton and Laurentia, as part of the global Grenville Orogenic Cycle, in the formation of the Rodinia Supercontinent. The last important tectonomagmatic event was the Sunsás Orogeny (SS in Figure 1) at about 1.3 to 1.0 Ga, in the southwesternmost part of the region, finally establishing the Amazonian Craton in Neoproterozoic times. For additional information the reader is referred to Tassinari *et al.* (this volume).

Figure 4 shows the still very scanty Nd isotope data available for the Amazonian Craton. In this figure, most of the ellipses, representative of the geotectonic units, plot in the vicinity of the mantle evolution curve (CHUR), displaying positive or slightly negative initial ε_{Nd} values, indicating that juvenile material, formed through mantle differentiation processes, is widespread. This is valid for the Carajás granitoid rocks of Archean age (CA1); for the high-grade granulite rocks of the Imataca Complex (IM), as well as for the early to middle Proterozoic granitoid rocks belonging to the younger tectonic provinces, where most of the samples yielded Sm/Nd T_{CHUR} model ages practically concordant with the Rb/Sr or U/Pb age values in the same samples. These are the granitoid rocks belonging to the Maroni-Itacaiúnas (MI1) Mobile Belt, the Ventuari-Tapajós (VT), and Rio Negro-Juruena (RNJ) magmatic arc systems, as well as the Rondonian-San Ignacio (RO) and Sunsás (SS) collision belts.

Two important exceptions, clearly indicating reworking from pre-existent crustal material, are: 1) - the Cupixi high-grade gneiss, which are part of the basement for the Maroni-Itacaiúnas Province (MI2 in Figure 4), which shows an Archean age, and 2) - the early to mid-Proterozoic intrusive granite bodies in the Carajás region (CA2).

São Francisco Craton

Teixeira *et al.* (this volume) address the geology and dynamics of the São Francisco Craton, with the support of radiometric, petrogenetic, geochemical and structural constraints. This unit, occurring in the eastern part of South America (Figs. 1 and 5), covering large areas of the states of Bahia and Minas Gerais, was intimately involved in the Neoproterozoic collage of western Gondwana. In contrast to the Amazonian Craton, a great number of radiometric ages are available for this geotectonic unit, and Teixeira and Figueiredo (1991) summarized most of them.

In a general way, the crystalline basement of the São Francisco Craton comprises Archean medium to high-grade metamorphic terranes, and granite-greenstone associations. Paleoproterozoic supracrustal belts and plutonic rocks of varied composition are also present. A rift-thrust belt tectonically affected the central part of the unit with N-S trending elongated grabens and basins. These are filled with the sedimentary rocks of the Paleoproterozoic to Mesoproterozoic Espinhaço Supergroup, which in turn are intruded by plutonic rocks, such as those of the Lagoa Real Complex. Moreover, in Neoproterozoic times, the platform cover of the Bambui Group extensively overlay the area.

In the northern part of the craton, Archean blocks (Gavião, Jequié, Serrinha, Rio Capim, etc.) were welded together with Paleoproterozoic terranes by the events of the Transamazonian Orogeny (2.1-1.9 Ga), which is considered to be the main period of crust formation in the region. The Contendas-Jacobina Lineament as well as the Salvador-Curaçá and the Itabuna-Atlantic Coast belts (Fig. 5) controled the structural framework in those times.

The Gavião Block (GA in Figure 5) is composed of juvenile TTG plutonic rocks with some evidence of crustal contamination and recycling, as well as a few greenstone belt sequences. The regional evolution of that area took place through successive events from 3.4 to 2.7 Ga, when juvenile and recycled granitoid rocks were formed (Cordani *et al.*, 1985). These terranes include the oldest grey gneiss reported in South America, as determined by U/Pb SHRIMP zircon ages of more than 3.4 Ga (Nutman and Cordani, 1994), and Sm/Nd model ages of up to 3.7 Ga. Such rocks are the Sete Voltas and Boa Vista - Mata Verde granitoid plutons, the main units among the few basement bodies introduced as tectonic slices into the Contendas-Mirante Belt during the Transamazonian Orogeny.

The younger Jequié Block (JQ) of late Archean age includes high grade granulite, mainly enderbitic-charnockitic gneiss with calc-alkaline affinities, together with some high-grade supracrustal rocks and associated basic granulite (Barbosa, 1990). Some of the granulites yielded Sm/Nd model ages older than 3.0 Ga, suggesting heterogeneity in the Jequié Block. These rocks were also further affected by migmatization and penetrative deformation during the Transamazonian Orogeny, accompanied by significant isotope resetting at 2.1-2.0 Ga.

The high-grade metamorphic rocks of the Salvador-Curaçá (SC) and Itabuna-Atlantic Coast (IT) belts show successive collisional phases related to the Transamazonian Orogeny. The youngest of these phases (2.0 to 1.9 Ga) is related to the formation of an enormous amount of granitoid rock. They occur frequently in both belts, but also in other parts of the São Francisco Craton, such as in the Guanambi-Urandi Batholith, a very large area that originated as a magmatic arc in the Neoproterozoic.

In the southern part of the craton, in the State of Minas Gerais (MG in Figure 5), the geological evolution is compatible with that of the Gavião Block. Three metamorphic complexes are recognized (Bomfim, Belo Horizonte and Campo Belo); composed essentially of TTG gneissic rocks, as well as some mafic and ultramafic intrusions and greenstone belts (Teixeira *et al.*, 1996). These are mainly situated within the so called Iron Quadrangle, making up the Rio das Velhas Supergroup, affected by the late Archean Rio das Velhas Orogeny, which produced medium-grade metamorphism in the supracrustal rocks, induced many TTG intrusive bodies, and reworked some of the older basement granitoid plutons. The more important Archean-forming event in the Campo Belo and Belo Horizonte complexes was juvenile in character, and took place in the late Archean.

During the Paleoproterozoic, in the southeastern part of the craton, the sedimentary, magmatic and tectonic records seem to suggest that a complete Wilson Cycle occurred in the area of the Mineiro Belt (MI in Figure 5). There the Transamazonian Orogeny formed a string of peraluminous granitoid and alkaline plutons, and induced regional metamorphism in the supracrustal rocks of the Minas Supergroup as well as in parts of their Archean basement. A two-stage model (Alkmin and Marshak, 1998) is now considered for the evolution of this belt, starting with a fold-thrust system, and the formation of island arcs, which later collapsed during a collisional phase at the end of the Transamazonian Orogeny.

The low to medium-grade metamorphic rocks of the Espinhaço Supergroup represent the evolution of a sedimentary basin within an aulacogenic-type environment. During the initial phases of this evolution, the Archean basement rocks in the Paramirim region of central Bahia were intruded by a series of anorogenic granitoid plutons, later deformed by the Mesoproterozoic Espinhaço Orogeny (Cordani *et al.*, 1992). The geological history of the São Francisco Craton ends with the Bambui Group cratonic cover, formed mainly by pelite and carbonate beds deposited in a shallow marine platform in Neoproterozoic times.

Figure 6 shows the Nd isotope data available for the São Francisco Craton. It includes the Archean granitoid rocks of Sete Voltas and Boa Vista (SV), tectonically activated in the Contendas-Mirante Belt during the Transamazonian Orogeny. Their ages are slightly older than 3.4 Ga, and their Nd isotope signature suggests juvenile mantle-type protoliths. In contrast, the Lagoa do Morro Gneiss (LM), which is also basement to the Contendas-Mirante Group, yielded younger radiometric ages, and their Nd isotopes indicate reworking from older material.

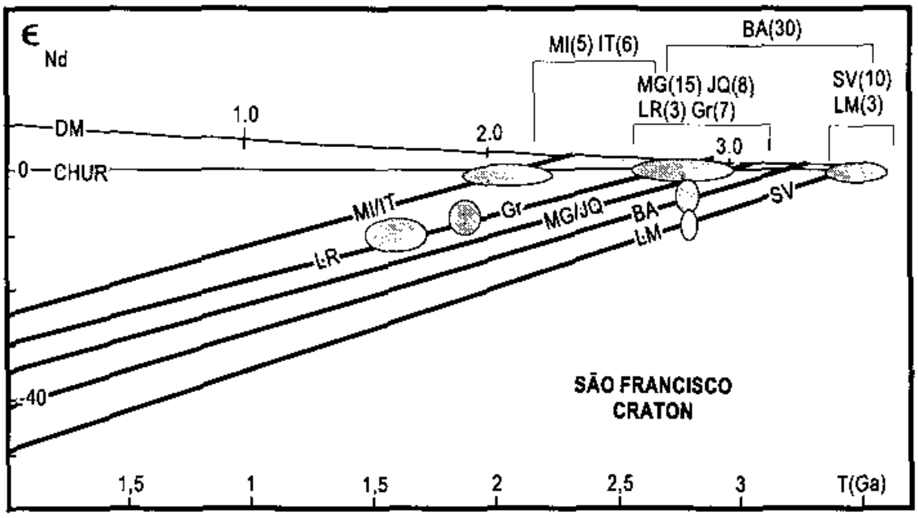


FIGURE 6 - Diagram of Nd isotope evolution for the São Francisco Craton.
SV = Sete Voltas and Boa Vista granitoid rocks; LM = Lagoa do Morro Gneiss; LR = Lagoa Real granitoid rocks; Gr = Granitoid rocks intrusive into the Contendas-Mirante Belt; BA = Gavião Block; MG = Metamorphic complexes of Minas Gerais; MI = Mineiro Belt; IT = Itabuna-Atlantic Coast Belt; JQ = Jequié Block.

Many granitoid rocks of the São Francisco Craton belong to large granite-greenstone or granite-gneiss-migmatite terranes, formed in the states of Bahia and Minas Gerais in the late Archean. In Bahia, within the Gavião Block and its vicinity, many granitoid rocks (BA) show strong negative initial ε_{Nd} values, indicating reworking from older crustal material. However, in Minas Gerais, the above-mentioned granitoid rocks of the Bonfim, Belo Horizonte, Campo Belo and similar old complexes (MG) yielded Nd parameters that plotted close to the mantle evolution curves, indicating predominance of juvenile components in their protoliths. The data from the late Archean high-grade rocks of the Jequié Complex in Bahia (JQ) also plotted near the mantle evolution curve, showing that they are not high-grade equivalents of the Gavião Block granitoid rocks.

A different episode of mantle-crust differentiation occurred in the Paleoproterozoic, in association with the Transamazonian Orogenic Cycle, when the Mineiro Belt formed in Minas Gerais and western Bahia (MI), and the Itabuna-Atlantic Coast Belt (IT) formed in eastern Bahia. Both of these geotectonic units seem to have originated from juvenile mantle-derived material. To this tectonomagmatic episode also corresponds the intrusion of granitoid rocks (Gr) into the Contendas-Mirante Belt, as well as towards the N along the Contendas-Mirante Lineament for which a contribution of crustal material will be accounted for.

Finally, the granitoid rocks of Lagoa Real (LR) are anorogenic, and linked to the middle Proterozoic extensional events of the Espinhaço Cycle (Cordani *et al.*, 1992). From their negative ε_{Nd} values, such intrusive rocks were also derived, at least partially, from pre-existing crustal material by reworking the Archean basement granitoid rocks of the Paramirim Complex.

Borborema Province

The Borborema Province consists of a series of crustal fragments, which took part in the process of agglutination of western Gondwana by the convergence of the large West African and Congo-São Francisco cratonic masses. It represents the westernmost extension of a huge Brasiliano-Pan African tectonic province, a "tectonic collage" within which several terranes of different character collided, during the Neoproterozoic. Almeida *et al.* (1977) defined the Borborema Province as a major structural province, a "complex mosaic-like folded region" where important tectonic, thermal and magmatic events occurred during the Neoproterozoic Brasiliano Orogenic Cycle. A detailed account of its geological evolution can be found in Brito Neves *et al.* (this volume).

The province includes a mosaic of Neoproterozoic fold belts, together with basement inliers in which many exposures of an older Transamazonian collage can be found. Structural trends show a fan-like distribution, and are cut by an important net of megashear zones or lineaments (Fig. 7). The main lineaments separate four tectonic domains, as follows:

1 – The Médio Coreau Domain (MC in Figure 7), to the W of the Transbrasiliano Lineament, which includes the reworked border of the São Luis-West African Craton and some fragments of fold belts, comprising supracrustal rocks that can be correlated with those of the Trans-Saharan Belt in western Africa. Few areas with volcano-sedimentary deposits, associated with undeformed granitic stocks of Cambrian age, are found within pull-apart basins developed along the Transbrasiliano Lineament.

2 – The Northern Domain, between the Transbrasiliano and the Patos lineaments, is underlain by an almost continuous Paleoproterozoic basement, which comprises the Tróia-Tauá (TT), Rio Piranhas (RP) and São José do Campestre (SJC) massifs, reworked and intruded by numerous granitoid bodies related to the Brasiliano-Pan African Orogeny. Within the São José do Campestre Massif, an Archean nucleus occurs. Several supracrustal belts are described within this domain, the Seridó Belt (SB) being the most investigated of these. It is composed of a quartzite-pelite assemblage, with minor volcanic components, submitted to a strong transpressional deformation during the Brasiliano Orogeny.

3 – The Transversal Domain, between the Patos and the Pernambuco lineaments, in the central part of the province, contains a well developed volcano-sedimentary fold belt, more than 750 km long and about 1.0 Ga to 950 Ma in age. This is the Cariris Velhos Fold Belt (CV in Figure 7), formed by meta-aluminous to peraluminous orthogneiss enclosed by felsic metavolcanic rocks and pelitic metasedimentary rocks, and affected by the Brasiliano-Pan African deformation. Some of

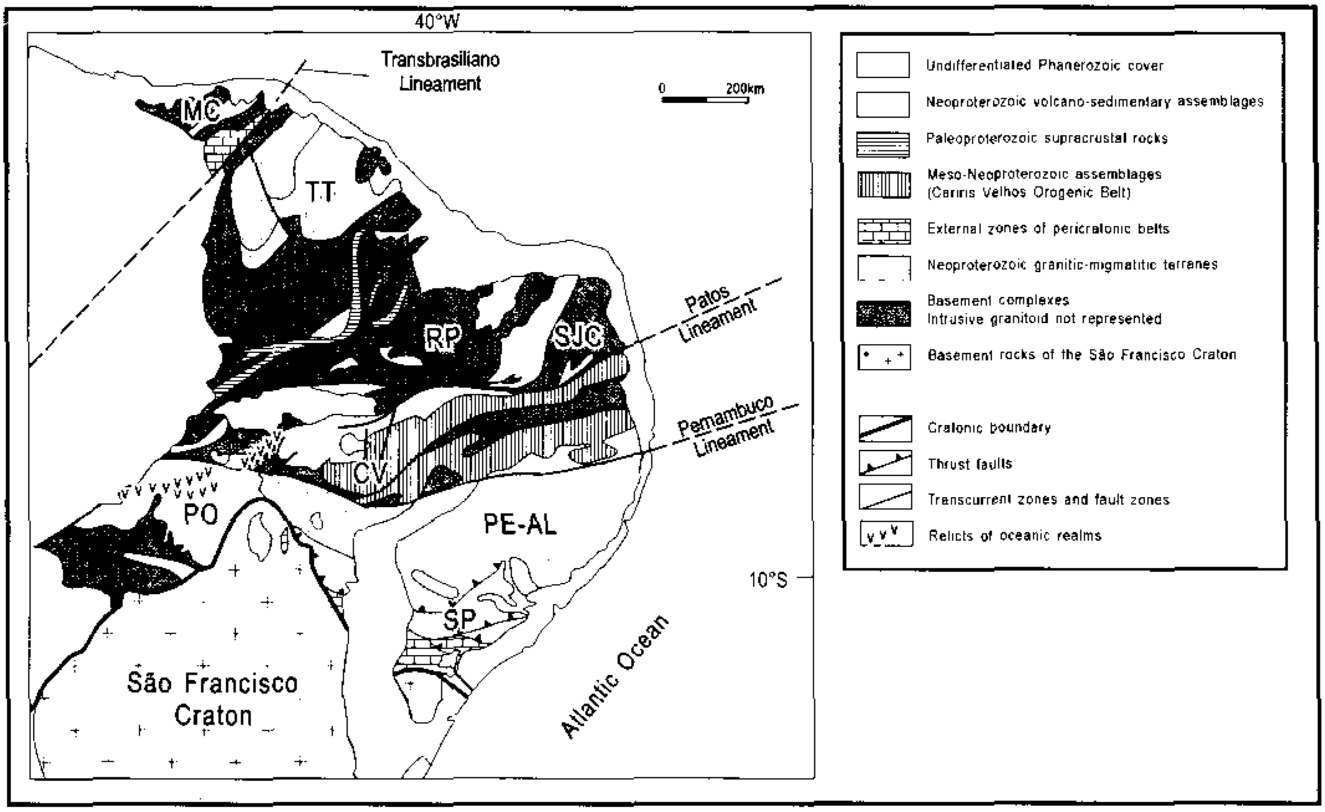


FIGURE 7 - The Borborema Province and its main tectonic domains. MC = Médio Coreau Domain; TT = Troia-Tauá Massif; RP = Rio Piranhas Massif; SJC = São João Campestre Massif; SB = Seridó Belt; CV = Cariris Velhos Fold Belt; PE-AL = Pernambuco-Alagoas Massif; PO = Riacho do Pontal Fold Belt; SP = Sergipano Fold Belt

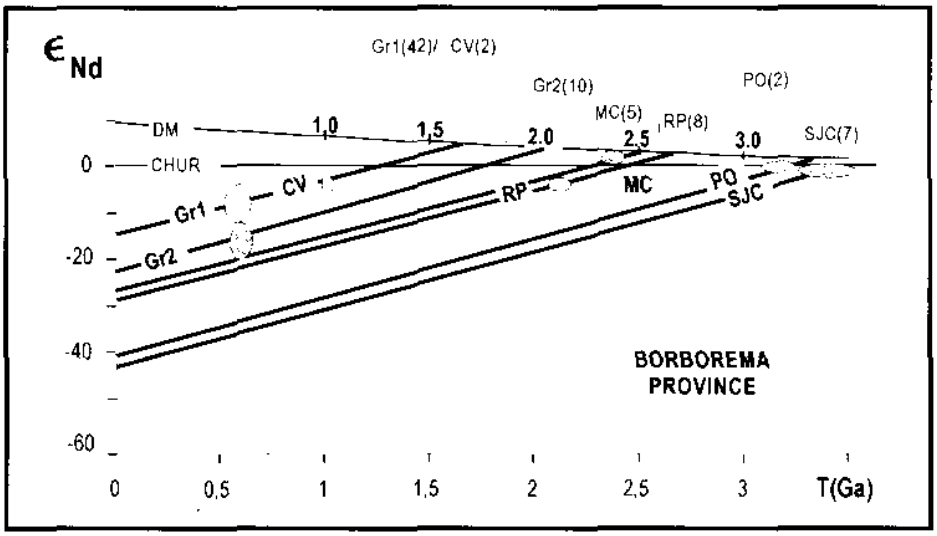


FIGURE 8 - Diagram of Nd isotope evolution for the Borborema Province. SJC = São João Campestre; PO = Riacho do Pontal; MC = Médio Coreau; RP = Rio Piranhas; CV = Cariris Velho granitoid rocks; Gr1 and Gr2 = intrusive granitoid rocks.

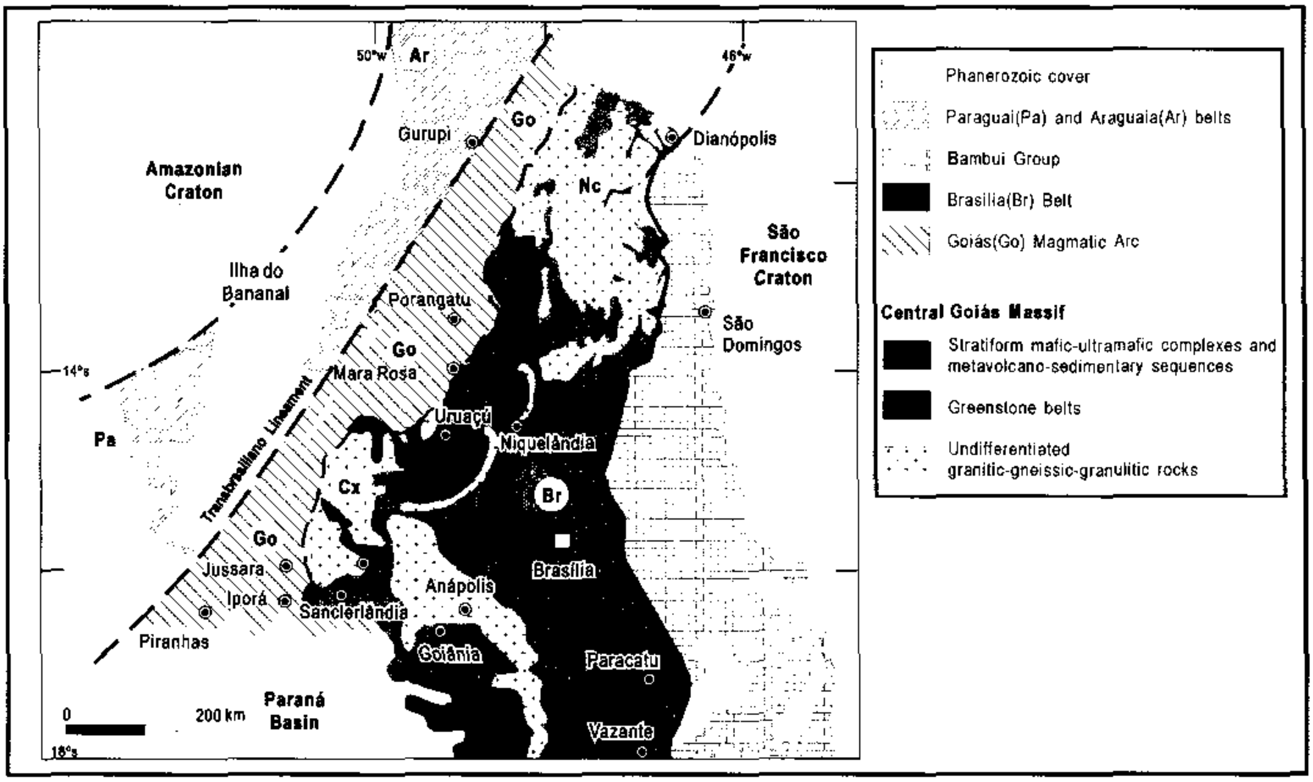


FIGURE 9 - The Tocantins Province and its main tectonic domains.
AR = Araguaia Belt; PA = Paraguai Belt; GO = Goiás Magmatic Arc; CX = Crixás granite-greenstone terrane; NC = Natividade-Cavalcante Block; BR = Brasília Belt

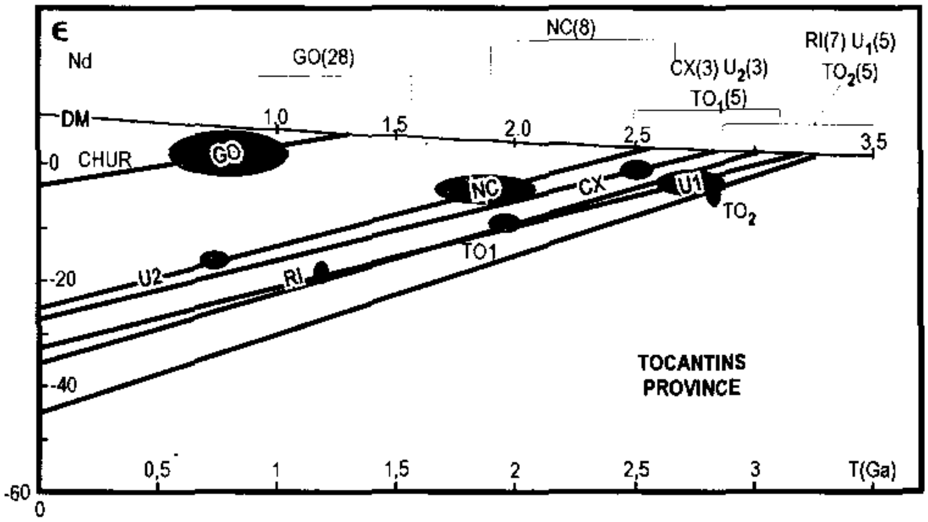


FIGURE 10 - Diagram of Nd isotopic evolution for the Tocantins Province TO_1 and TO_2 = Basement inliers into Araguaia Belt; U_1 and U_2 = Uvá basement inliers; CX = Granitoid rocks near Crixás; NC = Granitoid rocks in the Natividade-Cavalcante Block; GO = Granitoid rocks of the Goiás Magmatic Arc.

the clastic material from this unit yielded radiometric ages in the same interval, suggesting derivation from a juvenile magmatic arc near a continental margin. This unit is adjacent to a continuous Neoproterozoic belt, with low-grade schist interbedded with bimodal volcanic rocks with arc affinities, and intruded by granitoid rocks having an age of about 600 Ma. The Cariris Velhos Orogeny may indicate that the Borborema Province, or part of it, participated in the last stages of the agglutination of Rodinia. Alternatively, this orogeny may be the starting point for the process that led to the formation of western Gondwana.

4 - The Southern Domain, to the S of the Pernambuco Lineament, consists of marginal orogenic systems to the São Francisco Craton: the Sergipano (SP) and the Riacho do Pontal (PO) fold belts. These belts exhibit a clear tectonic polarity, with external and internal zones, and comprise continental margin assemblages (diamictite-quartzite-pelite-carbonate) gradually evolving toward volcano-sedimentary sequences and deep-water sediments. Relicts of oceanic lithosphere, or island arc material, have been described in some parts of the distal zones. The innermost zones consist of extensive granitoid rocks of the Pernambuco-Alagoas Massif (PE-AL in Figure 7), the radiometric ages of which are mainly Neoproterozoic, but with many indications of nuclei formed during previous magmatic events. The initial set up of the passive margins is considered to be around 750 Ma; the main phase of folding occurring at about 630 Ma; and late to post-orogenic plutons formed between 580 and 540 Ma (Brito Neves *et al.*, 1999). Basement inliers within the Riacho do Pontal Belt can be considered the extension of rocks of the São Francisco Craton, reworked by the Brasiliano Orogeny.

A large number of Nd isotope determinations are available for this region, but we will only include in Figure 8 those which seem to be the most appropriate to demonstrate its geochemical evolution. A reasonable grouping of isotope determinations belongs to the Rio Piranhas Massif (RP), which includes the adjacent basement of the Seridó Belt where ear. Proterozoic rocks yielded slightly negative ε_{Nd} values, indicative of at least partial reworking from older crustal material. Gneissic rocks of the Médio Coreaú Complex (MC) occurring to the W of the Transbrasiliano Lineament and representing probable extensions of the São Luís Craton, give similar values of T_{DM} but positive ε_{Nd} values, indicating juvenile protoliths of possible late Archean age. Clear Archean ages were obtained in a few samples from the São João do Campestre (SJC), and from the Riacho do Pontal (PO) basement rocks. A different type of basement rock is exposed in the transversal domain of the Borborema Province, where granitoid rocks related to the Cariris Velhos Orogeny (CV) yielded radiometric ages around 1.0 Ga, with moderately negative ε_{Nd} values, indicative of crustal protoliths.

Many granitoid rocks are widespread in the entire Borborema Province, intrusive either into its basement rocks, or into the various volcano-sedimentary supracrustal belts These granitoid intrusives yield ages usually between 650-530 Ma. Their Nd isotope composition is plotted in two ellipses (Gr1 and Gr2) in Figure 8, and their position in the Nd evolution diagram indicates that their material can not be a direct derivation by partial or total fusion of the basement rocks already described. Some juvenile component, differentiated from the mantle during the Mesoproterozoic or Neoproterozoic, shall be taken into account for their formation.

Tocantins Tectonic Province

The Tocantins Tectonic Province (Almeida *et al.*, 1981) underlies a large part of Central Brazil, and it consists of a series of cratonic fragments that resulted from the collision between the large continental masses of the Amazonian and São Francisco cratons. This collisional process took place in the Neoproterozoic, when the entire province was affected by and partially reworked in the Brasiliano Orogenic Cycle. The main megasuture related to this event is the Transbrasiliano Lineament that crosses the entire region (Fig. 9). To the NW of that lineament, the Araguaia and the Paraguay fold belts are marginal to the Amazonian Craton, and display a tectonic vergence towards it. To the SW of the Transbrasiliano Lineament, there occurs a mosaic of tectonic blocks of different age and evolution. More up to date information regarding the Tocantins Province can be found in Pimentel *et al.* (this volume).

The Araguaia Belt (AR in Figure 9) is a N-S trending thrust and fold belt, 1000 km long, composed of low to medium-grade metapelite with minor carbonate rocks and numerous interbedded mafic and ultramafic bodies which have been interpreted as ophiolite remnants (Koutschoubey and Hieronimus, 1996). This structural domain includes some mantle gneiss domes cored by orthogneiss and migmatitic rocks with Archean and Paleoproterozoic ages. The Ilha do Bananal Cenozoic sediments conceal the possible link between the Araguaia and the Paraguai belts (PA), a similar system marginal to the Amazonian Craton in its southwestern part. The latter includes glaciogenic rift-related sequences of Vendian age (Trompette, 1994) followed by pelitic and carbonate rocks of a passive margin type environment.

The complex collisional system occurring to the SE of the Transbrasiliano Megasuture, including the Brasília Belt, comprises very large areas of the so-called Goiás Magmatic Arc (Pimentel and Fuck, 1992; Pimentel *et al.*, 1997), GO in Figure 9. This unit contains metaplutonic rocks, mainly tonalite and granodiorite along and adjacent to the suture. The radiometric ages span a large interval, at least between 930 and 640 Ma. The isotope geochemistry of these granitoid rocks suggests their origin from mantle derived juvenile material, and their formation within a series of island arcs. They are the best witnesses to the existence of a large region with oceanic floor, the Goianides Ocean of Pimentel *et al.* (1997), developed during the Neoproterozoic, and separating the Amazonian and the São Francisco-Congo cratons.

The Brasília Belt (BR) comprises low-grade metasediments, interpreted as shelf and slope deposits characteristic of a passive margin, defining the western limit of the São Francisco Craton. They are covered by the pelite-carbonate sequence of the Bambuí Group, which also overlies large parts of the São Francisco Craton. Thomaz Filho *et al.* (1998) suggested that the Bambuí Group was deposited in a foreland basin, in response to the development of the Araxá thrust sheets to the W.

The Central Goiás Massif, situated between the Goiás Magmatic Arc and the Brasília Belt, is a very heterogeneous domain formed by different juxtaposed tectonic blocks. One of these is the Crixás granite-greenstone terrane (CX) that displays late Archean ages. Others are referred generally as basement terranes, and they are included in Figure 9 within a general unit of undifferentiated granitic-gneissic-granulitic rocks. They are usually granitoid in character with ages varying between Archean and Mesoproterozoic. High-grade metamorphites occur near Anápolis and Goiânia forming the Neoproterozoic Anápolis-Itauçu Complex. In the central part of the province the three large layered mafic-ultramafic complexes of Canabrava, Niquelândia and Barro Alto are described. Their probable age of formation is Paleoproterozoic, with deformation and metamorphism in the Neoproterozoic (Correia *et al.*, 1997). Within this region that can be considered as the internal zone of the Brasília Belt, there occur the metasediments of the Araxá Group, consisting mainly of micaschist and quartzite. They exhibit coherent petrological and structural features, essentially shallow angle foliation and thrust sheets verging toward the São Francisco Craton. They may correspond to an accretionary wedge thrust over the shelf sequence of its passive margin.

Figure 10 shows the Sm/Nd systematics for the Tocantins Province. It demonstrates clearly that a large number of granitoid rocks from this region (GO) correspond to juvenile material, differentiated from the mantle in Neoproterozoic times. Such granitoid rocks are considered to have formed in a series of magmatic arcs, indicating the existence of an important episode of crustal accretion, forming a large region immediately SE of the Transbrasiliano Lineament (Pimentel and Fuck, 1992). Some granitoid intrusions into the Araxá Group, in the southeastern part of the province show a similar mantle signature.

Rocks from the mantled gneiss domes occurring within the Araguaia Belt yielded different types of isotope signatures. Some of them (TO_1) indicated Paleoproterozoic ages, and derivation from crustal protoliths of probable late Archean age. Others (TO_2) resulted late Archean in age, but their isotope systematics also indicated reworking from still older crustal material.

Within the Archean cratonic fragment of Crixás, a few gneissic rocks (CX) yielded Sm/Nd T_{DM} model ages roughly concordant with the Rb/Sr available determinations indicating a juvenile mantle source for their protoliths. However, in other cases such as the Uvá (U_1 and U_2), as well as the Ribeirão (RI) granitoid gneiss, reworking from older crustal protoliths of Archean age is indicated. This evolutionary pattern can most probably be applied to many other crustal fragments of the tectonic province. For instance, in the large northeastern region of the Tocantins Province, including the localities of Natividade and Cavalcante (NC in Figure 10), the available Nd isotope determinations on Paleoproterozoic rocks indicate slightly negative initial ε_{Nd} values, suggesting a reworking from Late Archean crustal protoliths.

Mantiqueira Tectonic Province – northern sector

The northern sector of the Mantiqueira Tectonic Province includes a series of crustal fragments of different age and tectonic evolution, which took part of the Neoproterozoic agglutination of the Gondwana Supercontinent. The geodynamics involved the interaction of the continental masses of the São Francisco-Congo and Kalahari cratons, as well as the Rio de La Plata and Luiz Alves cratonic fragments, and the concomitant disappearance of the Adamastor Ocean. It constitutes a NE trending orogenic system, controlled by large strike-slip shear zones marking the borders of the different terranes that collided successively against the São Francisco Craton as well as against the Paraná cratonic area thought to exist below the sediments of the Paraná Basin. The complexity of the area is such that we will not attempt here to explain its diversity. The interested reader is referred to Brito Neves *et al.* (1999) or Campos Neto (this volume), where these authors put forward a model for the regional tectonic evolution, as an orogenic collage formed by several allochtonous terranes with different origins, including cratonic fragments and magmatic arcs. Figure 11, adapted from Campos Neto (this volume) gives an idea of the structural arrangement of the orogenic collage, and the tectonic nomenclature of the main terranes.

In the northern part of the sub-province the Araçuaí Belt is a passive continental margin at the eastern border of the São Francisco Craton. This Neoproterozoic marine environment opened toward the Rio Doce Gulf (Brito Neves *et al.*, 1999), an oceanic embayment formed slightly earlier, when the São Francisco and Congo cratons tried to separate from each other. Several basement terranes of Paleoproterozoic age, which are supposed to be witnesses of the aforementioned disruption episode, accreted back to their original position and closed the northern part of the Adamastor Ocean during the Brasiliano-Pan African Cycle. Remains of the Neoproterozoic oceanic lithosphere in that region were described by Pedrosa Soares *et al.* (1998).

Basement rocks of granitoid character are found within the Araçuai Belt, comprising the Guanhães and Mantiqueira terranes (Fig. 11). They are polycyclic medium to high-grade gneiss-migmatitic rocks, including some supracrustal relicts, displaying N-S structural trends but including an E-W mineral stretching lineation (Marshak and Alkmin, 1989). Their primary U/Pb zircon gave ages of about 3.0 Ga, but the rocks were affected by late Archean, Transamazonian, and finally Neoproterozoic tectonic and metamorphic events, the latter related to the last continental collision of the Araçuaí Belt (Alkmin and Marshak, 1998).

To the E, the granulite and kinzigite of the Juiz de Fora Terrane predominate, with Paleoproterozoic radiometric ages overprinted by the Brasiliano-Pan African episodes, but displaying Archean Sm/Nd model ages, and a Sr isotope systematics indicating their formation from crustal protoliths. Nevertheless, metaplutonic rocks of the Juiz de Fora Complex disclose contrasting Paleoproterozoic to Mesoproterozoic Sm/Nd T_{DM} model ages, suggesting possible juvenile additions to the older continental masses.

From this evidence, it seems that the northern segment of the Mantiqueira Province represents part of the reworked margin of a large continental mass that coalesced by the end of the Rio das Velhas Orogeny in the late Archean. It was later affected during the Transamazonian Orogeny by the action of the Mineiro Belt, where some juvenile material was accreted in the early Proterozoic, and still later it suffered the tectonic

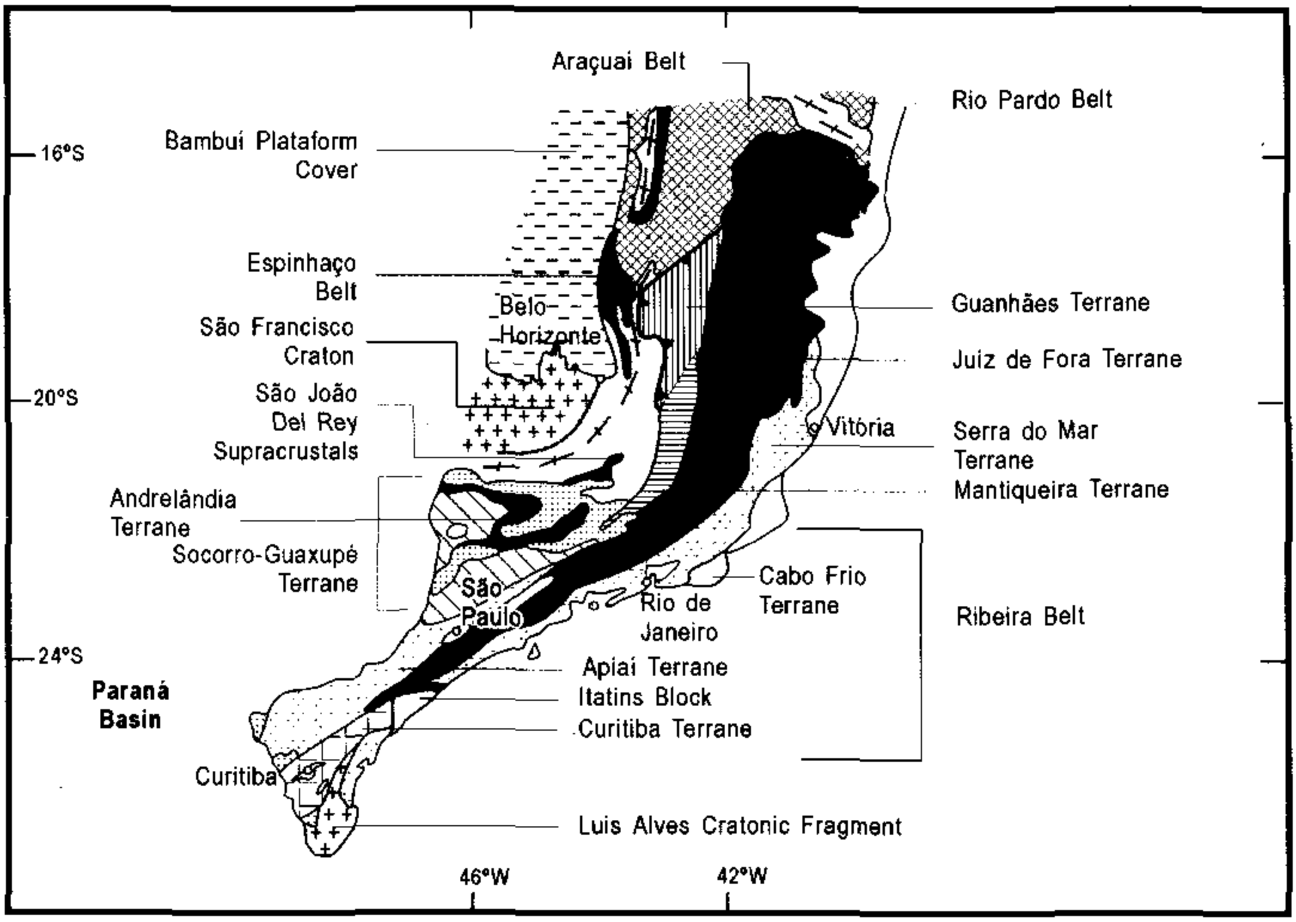
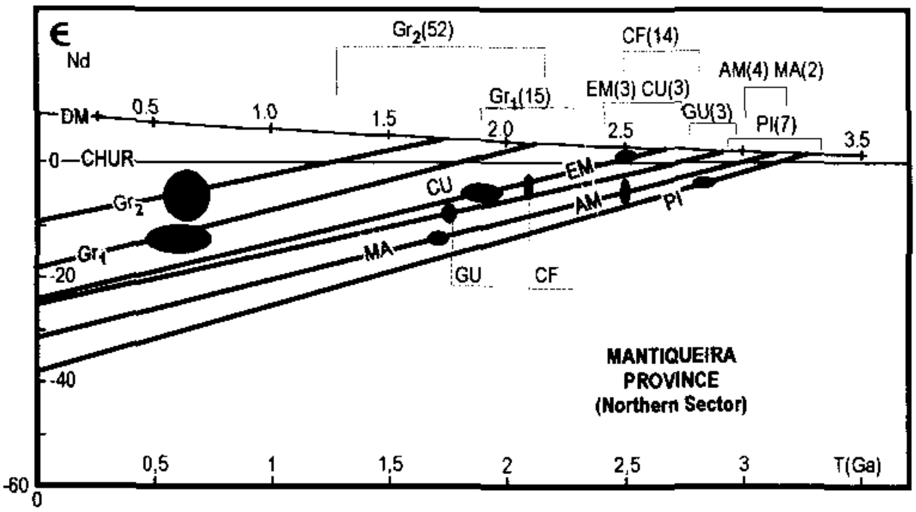


FIGURE 11 – Principal tectonic units of the northern part of the Mantiqueira Province.

FIGURE 12 – Diagram of Nd isotopic evolution for the northern part of the Mantiqueira Province.
EM = Embu Terrane; PI = Gneiss-migmatitic rocks near Piripá;
AM = Amparo grey gneiss;
CF = Granitoid rocks near Cabo Frio;
CU = Granitoid rocks of the Curitiba Massif; GU = Granitoid rocks of the Guanhães Terrane; MA = Granitoid rocks of the Mantiqueira Terrane;
Gr_1 and Gr_2 = Intrusive granitoid into Araçuai and Ribeira belts



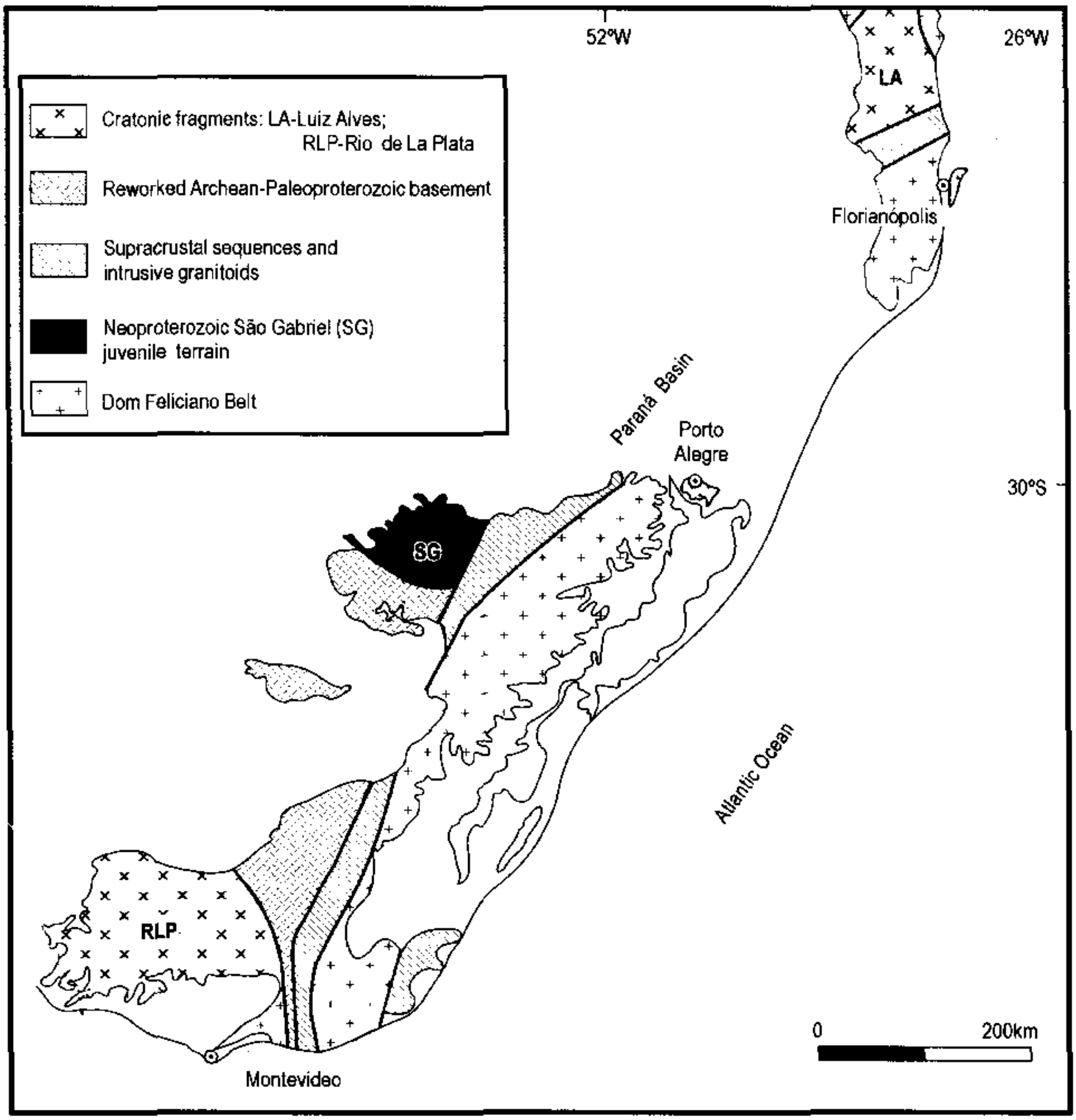


FIGURE 13 – Principal tectonic domains of southern Brazil and Uruguay.

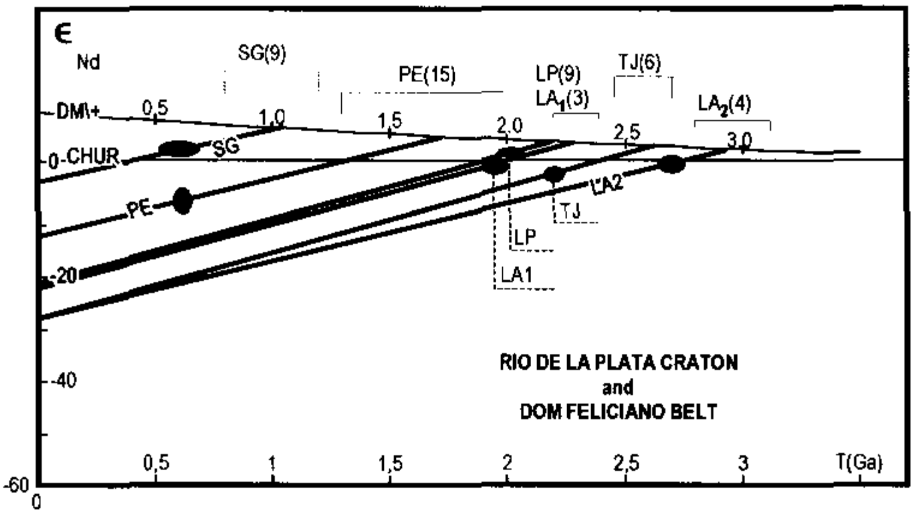


FIGURE 14 – Diagram of Nd isotope evolution for the Luís Alves and Rio de La Plata cratonic fragments and the Dom Feliciano Belt.
LA_1 and LA_2 = Luiz Alves high-grade gneiss; TJ = Basement inliers into the Dom Feliciano Belt; LP = Rio de La Plata granitoid rocks; PE = Pelotas Batolith; SG = São Gabriel Block

overprinting of the Neoproterozoic collision associated with the Araçuaí Belt. A similar polycyclic framework is also seen at the southern end of the São Francisco Craton, where granitoid basement rocks are found within the Mesoproterozoic to Neoproterozoic basins of São João del Rey, Carandaí and Andrelândia (Söllner and Trouw, 1997).

Reworked Archean basement rocks were also detected in some other localities, such as Piripá, Bahia, at the northernmost extremity of the Mantiqueira Province, and at its easternmost end in the Cabo Frio region.

The central segment of the northern sector of the Mantiqueira Province is a giant thick-skinned frontal nappe system, the Socorro-Guaxupé Terrane (Fig. 11), structurally related to the series of ENE trending thrusts of the Araxá Group (Campos Neto and Caby, 1999). This system includes basement rocks of different ages, including the Amparo high-grade gneiss of Archean age as well as magmatic arc assemblages and volcano-sedimentary units of Mesoproterozoic to Neoproterozoic ages. Deformation is mainly related to a large collisional event, when the São Francisco Craton collided against the Paraná cratonic area in one of the earlier episodes of the Brasiliano Orogeny in the context of agglutination of western Gondwana.

The southern segment of the region is dominated by the Ribeira Belt, and is cut by several very long NE trending fault zones, which separate many tectonic blocks. At least two main allochtonous units are defined: the Apiaí and Curitiba terranes (Fig. 11), both of them presenting gneiss-migmatite basement rocks, overlain by the volcano-sedimentary sequences of Mesoproterozoic to Neoproterozoic age. The source areas for these sequences could be continental relief adjacent to passive margins, or recently formed magmatic arc systems such as, for instance, the Cunhaporanga and Três Córregos complexes, which contain granitoid rocks of many different types in the 700 to 540 Ma age range, widespread in both Apiaí and Curitiba terranes.

Figure 12 shows the Nd evolution diagram for granitoid rocks from the northern part of the Mantiqueira Province. With the exception of the Embu Terrane (EM) within the Ribeira Belt, which yielded a juvenile late Archean Nd isotope signature, all the other "basement units" within both Araçuaí and Ribeira belts presented moderately to strongly negative initial ε_{Nd} values, indicating reworking from Archean crustal protoliths. Such units are identified as Piripá (PI), Amparo (AM), Cabo Frio (CF), Curitiba (CU), Guanhães (GU) and Mantiqueira (MA) in Figure 6. The oldest Sm/Nd T_{DM} model age of about 3.2 Ga was obtained for the late Archean Piripá basement terrain within the Araçuaí Belt, a value which compares with those available for the adjacent Gavião Block of the São Francisco Craton. The other basement units are younger, most of them indicating a Paleoproterozoic age and formation by reworking of older crustal material.

A large amount of granitoid rock belongs to many intrusive complexes, formed during the tectonomagmatic episodes of the Brasiliano Orogenic Cycle. Their Nd isotope signatures plotted in two clearly distinct groups in Figure 12, which differ in the initial ε_{Nd} values. Those with higher values (Gr_1) correspond to magmatic intrusions into basement rocks of the Ribeira Belt, formed probably by partial fusion of older crustal material, in which a Paleoproterozoic component is widespread. The other group (Gr_2) includes intrusive granitoid in both the Araçuaí and Ribeira belts, and especially those considered by Campos Neto and Figueiredo (1995) as belonging to the Serra do Mar Microplate, the youngest allochtonous terrane which joined the previous collage in the process of agglutination of western Gondwana. The latter granitoid rocks exhibited a large spread in Sm/Nd T_{DM} model ages, indicating that a probable Neoproterozoic juvenile component was present mixed with different amounts of older crustal material within their magmatic sources.

The Rio de La Plata and Luiz Alves cratonic fragments, and the Dom Feliciano Belt

The major geotectonic units of southern Brazil, Uruguay and the Province of Buenos Aires in Argentina are the Neoproterozoic Dom Feliciano Belt and its complex forelands, and the Rio de La Plata and Luiz Alves cratonic masses (Figs. 1 and 13). Two other tectonic units, of still imprecise tectonic significance are known in the region: the São Gabriel Block in the State of Rio Grande do Sul, Brazil; and the Punta del Este Terrane, in Uruguay. A more detailed description of these units and their tectonic evolution can be found in Basei *et al.* (this volume).

The Dom Feliciano Belt is a very coherent structural feature in its 1200 km long regional extension, with tectonic vergence toward the NW. Its internal organization shows three narrow and subparallel crustal segments: a granite belt at the southeastern end; a central schist belt, and a northwestern foreland belt with volcano-sedimentary sequences. The granite belt is formed by a series of essentially igneous complexes, of which the Pelotas Batholith in the State of Rio Grande do Sul is the most representative, formed in a magmatic arc tectonic environment. The granitoid rocks include a series of deformed calc-alkaline types (tonalite, granodiorite, etc.) and many weakly deformed to undeformed monzogranite and granodiorite bodies, together with isotopically very homogeneous leucogranite. The geochemical data for these granitoid rocks indicate that they were generated by magma in which crustal participation and reworking was very important (Babinski *et al.*, 1997). The schist belt includes supracrustal rocks in a few discontinuous polydeformed metamorphic complexes. They all possess crustal basements of Paleoproterozoic age, and were affected by strong metamorphism during the Brasiliano Orogeny. The foreland basins form a narrow belt over the cratonic borders, and their marine sedimentary features suggest that a sea opening to the SW may have interlinked them. They include fossils indicating a Vendian to Lower Cambrian depositional age, one of the youngest events related to the Brasiliano Cycle.

The crystalline terranes observed in the northern and central-western parts of the Dom Feliciano Belt that served as foreland for it are those of Luiz Alves and Rio de La Plata (LA and RLP in Figure 13). A major NE trending fault zone, the Pien Transcurrent Fault Zone, separates the Luiz Alves Craton from the Curitiba Microplate, here considered as a tectonic suture and representing the southern limit of the northern sector of the Mantiqueira Province.

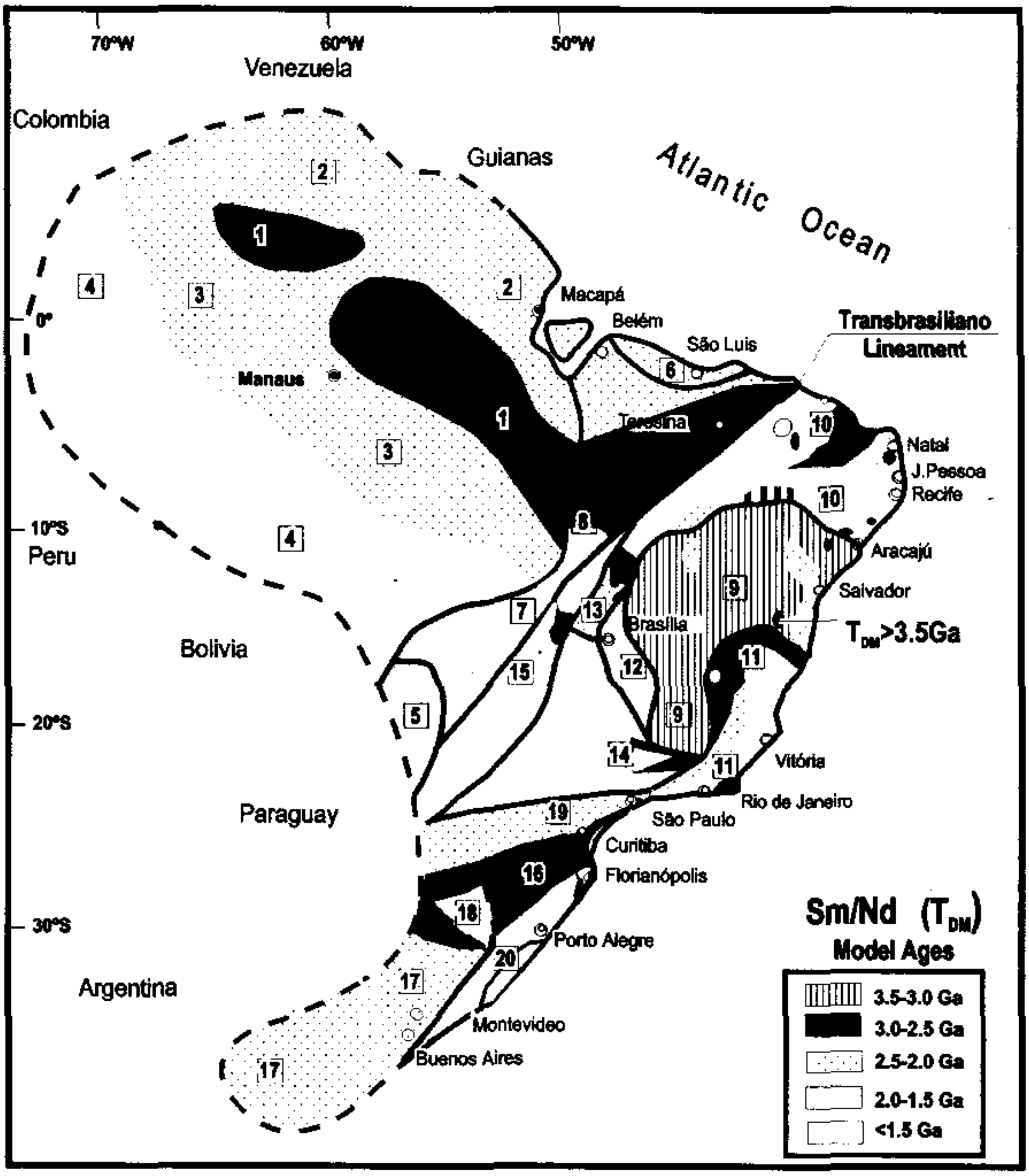


FIGURE 15 – Crustal domains of the South American Platform. 1 - Central Amazonian Province; 2 - Maroni-Itacaiúnas Province; 3 - Ventuari-Tapajós Province; 4 - Rio Negro-Juruena and Rondonian Provinces; 5 - Rio Apa cratonic fragment; 6 - São Luis cratonic fragment; 7 - Tectonic domain of the Paraguai-Araguaia Belt; 8 - Tectonic domain of the Tocantins Belt; 9 - São Francisco Craton; 10 - Tectonic domain of the Borborema Province; 11 - Tectonic domain of the Araçuaí Belt; 12 - Tectonic domain of the Brasília Belt; 13 - Tectonic domain of the Uruaçu Belt; 14 - Tectonic domain of the Araxá and Alto Rio Grande belts; - 15 Goiás Magmatic Arc; 16 - Luis Alves cratonic fragment; 17 - Rio de La Plata Cratonic fragment; 18 - São Gabriel tectonic block; 19 - Ribeira Belt; 20 - Dom Feliciano Belt

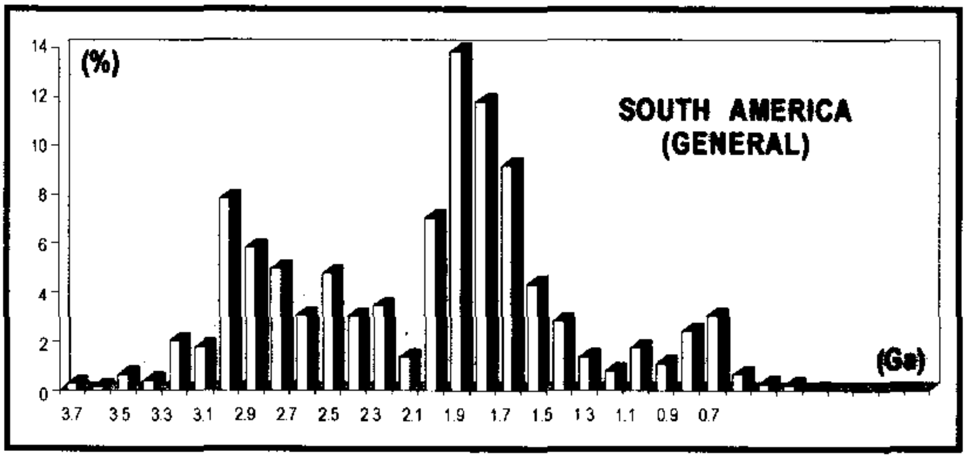


FIGURE 16 – Histogram of continental growth (or survival) for the South American Platform. Continental crust production based on Sm/Nd (T_{DM}) model ages.

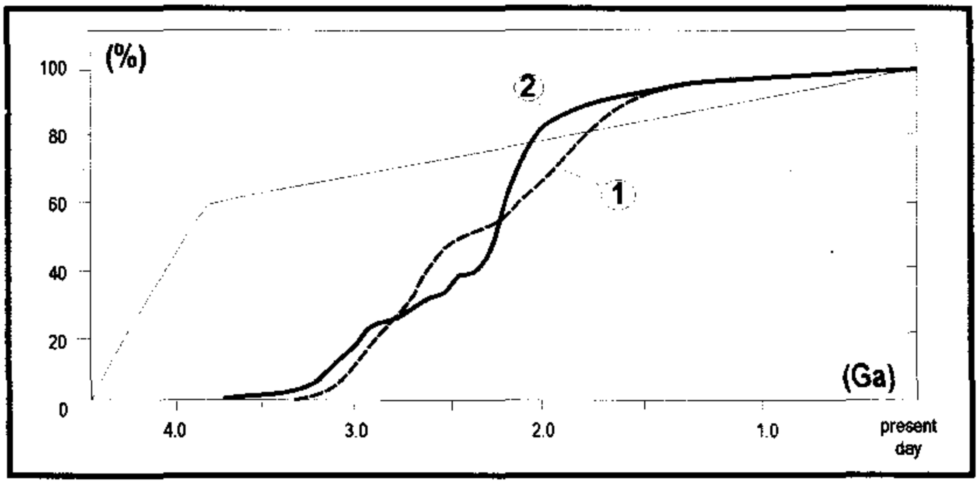


FIGURE 17 - Cumulative curves for continental growth (or survival) for the South American Platform.
(1) - From Cordani et al., *1988, based on Sr isotopes, and (2) - This work based on Nd isotopes.*

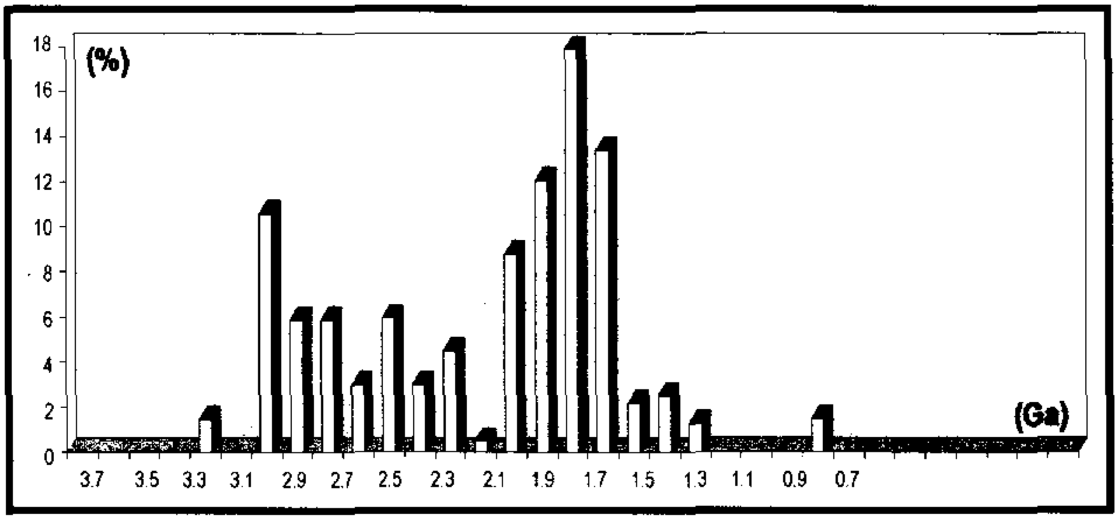


FIGURE 18 - Histogram of continental growth (or survival) for the northwestern part of the South American Platform.

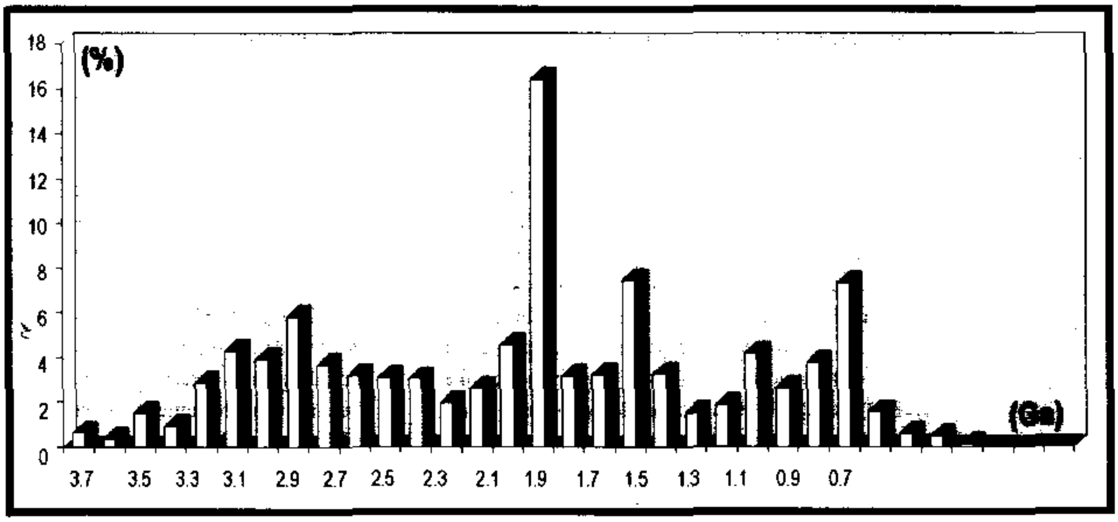


FIGURE 19 - Histogram of continental growth (or survival) for the southeastern part of the South American Platform.

The Luiz Alves Cratonic Fragment is a microplate in which orthoderived high grade metamorphic rocks predominate (Basei *et al.*, 1998). The main rock-types are charno-enderbitic granulite, chemically depleted in incompatible elements, and with radiometric ages in the 2.7 to 2.0 Ga interval. K/Ar ages of around 1.8 Ga indicate that the Neoproterozoic Brasiliano Orogeny did not thermally affect this unit. The Rio de La Plata Craton was defined by Almeida *et al.* (1973) and includes different terranes with ages of formation older than the Neoproterozoic, such as the Tacuarembó-Rivera (Brasil and Uruguay), Nico Perez and Piedras Altas (Uruguay) and Tandil (Argentina) terranes. The available radiometric ages indicate that the Brasiliano Orogeny thermally affected some of them, especially the Tacuarembó Terrane. In a general way, most age values are related to the Paleoproterozoic, considered to be the main crust-forming event for the geotectonic unit. The Piedras Altas Terrane occupies the southwestern end of the outcrop area of the Rio de La Plata Craton, and E-W trending orthoderived gneiss and migmatite, interbedded with some structurally concordant volcano-sedimentary belts make it up. This unit represents a juvenile crustal segment of Paleoproterozoic age (2.2 to 1.9 Ga), unaffected by later tectonothermal events, and cut by an important mafic dyke swarm of about 1.73 Ga in age (Bossi *et al*, 1993; Teixeira *et al.*, 1999). Radiometric ages of about 1.8 Ga are reported by Dalla Salda *et al.* (1992) in the Serra del Tandil (Argentina) where TTG granitoid plutons are considered to be tardi-tectonic in relation to the Transamazonian Orogeny.

The São Gabriel Bock, in the State of Rio Grande do Sul, Brazil (SG in Figure 13), is the only tectonic domain of the region under examination that was formed by Neoproterozoic juvenile material, with ages in between 750 and 600 Ma. It is situated to the W of the Dom Feliciano Belt, and consists of orthoderived gneiss with composition varying from diorite to tonalite and trondhjemite. Their mantle-derived isotope geochemistry demonstrates that they are very different in their petrogenesis from the Dom Feliciano granitoid plutons (Babinski *et al.*, 1995). Finally, the Punta del Este Terrane was defined by Preciozzi *et al.* (1999) as gneiss and migmatite formed in the 1.0 Ga - 900 Ma interval, but tectonically reworked during the Brasiliano Orogeny. The above-mentioned authors proposed that this terrane might represent a small piece of the Namaqua tectonic province, well represented in Southern Africa as one of the largest parts of the Kalahari Craton. It may have remained attached to the Dom Feliciano Belt after the break-up of Gondwana and the formation of the South Atlantic Ocean.

From the main area of the Rio de La Plata Craton, the Sm/Nd T_{DM} model ages (LP in Figure 14) indicated a juvenile derivation. In the same figure, the high-grade gneiss of the Luiz Alves Cratonic Fragment plotted in two groups (LA_1 and LA_2). Both groups plotted in the vicinity of the mantle evolution line, and correspond to two different episodes of mantle differentiation and continental crust formation; the older in the Late Archean, and the younger in the Paleoproterozoic.

Within the Dom Feliciano Belt, some basement gneiss (TJ) belonging to the Tijucas Fold Belt, with early Proterozoic radiometric ages, yielded Sm/Nd T_{DM} model ages of about 2.6 Ga and slightly negative initial ε_{Nd} values, indicating some reworking from older crustal materials.

In addition, a large number of intrusive granitoid bodies belonging to the Pelotas Batholith (PE), formed in Neoproterozoic times, yielded moderately negative initial ε_{Nd} values and Mid-Proterozoic Sm/Nd T_{DM} model ages. They exhibited a large spread in the individual values, indicating very probably their formation within a root of a magmatic arc where Neoproterozoic juvenile mantle-derived material was mixed with older continental crust.

A peculiar geotectonic feature of southern Brazil, the geotectonic significance of which is not yet completely understood, is the so-called São Gabriel Block. Here there occur deformed granitoid gneiss (SG), formed during the Brasiliano Orogeny, and yielded radiometric ages around 600 Ma, giving a mantle Nd isotope signature, with slightly positive initial ε_{Nd} values.

Concluding remarks

For the Archean rocks of the South American Platform, the isotope systematics indicates successive pulses of juvenile material, as in other parts of the world. The oldest rocks found so far, the aforementioned Sete Voltas and Boa Vista granitoid plutons from the Gavião Block within the São Francisco Craton, yielded Rb/Sr and U/Pb zircon SHRIMP ages of about 3.4 Ga., and Sm/Nd T_{DM} model ages up to 3.7 Ga.

For Paleoproterozoic and Mesoproterozoic rocks, both juvenile and reworked granitoid rocks are found. In the eastern part of the platform, corresponding to the São Francisco Craton and to the basement of the younger mobile belts, which surround it, reworked material predominates. However, for the western part, and especially within the Ventuari-Tapajós and the Rio Negro-Juruena tectonic provinces of the Amazon Craton, juvenile material is widespread. The geotectonic model for this region suggests the evolution of a very large oceanic lithosphere plate undergoing continuing subduction, with the formation of successive magmatic arcs, later accreted to the adjacent continental masses.

In Mesoproterozoic to Neoproterozoic times, granitoid rocks formed within the mobile belts of this age, or formed by intraplate processes over the cratonic areas, seem to have originated by reworking of older crustal material. Exceptions are to be made for the important Neoproterozoic magmatic arcs near the Transbrasiliano Lineament in central Brazil, and within the São Gabriel Block in southern Brazil.

Considering all the available Sr and Nd isotope information, the South American Platform was subsequently divided into crustal domains with internally coherent structural evolution and geochronological pattern. The resulting Figure 15 keeps the main boundaries of Figure 1, and the Amazonian Craton was further subdivided according to its internal tectonic provinces. Where the Precambrian basement is concealed beneath the sedimentary rocks of the Amazonas, Parnaíba and Paraná basins, the crustal domains were extrapolated with the help of some borehole information. The extension of the South American Platform beneath the Andean foredeep basins (Llanos, Beni, Chaco, and Pampas) was not considered, because of complete lack of information. The area shown in Figure 15 includes Brazil, French Guiana, Guyana, Surinam, Uruguay, and parts of Bolivia, Venezuela, Paraguay and Argentina, comprising 9.3 million km².

The Sm/Nd T_{DM} values from all the crustal provinces were employed in the construction of the histogram of Figure 16. From this figure, it is apparent that small amounts of continental crust older than 3.3 Ga survived within Archean fragments formed between 3.1 and 2.6 Ga. However, the main period of continental crust formation was between 2.2 and 2.0 Ga, corresponding to the Transamazonian Orogenic Cycle. The predominance of Paleoproterozoic juvenile material in the histogram of Figure 16 is due to the very large area within the Amazonian Craton consisting of juvenile magmatic arcs. Accretion of juvenile material continued until the Neoproterozoic, but at much slower rates. A small peak in the histogram corresponds to the Espinhaço/Rondonian-San Ignacio Cycle (1.4 - 1.3 Ga.) but the Brasiliano Orogenic Cycle is barely visible in the figure. The continental growth (or survival) curve derived from the Sm/Nd measurements (see the cumulative curve 2 in Figure 17) is not very much different from the one based in the Rb/Sr results. In Figure 17, it indicates that about 34% of the material encountered at present in the continental crust of the South American Platform was formed in the Archean. By the end of the Transamazonian Cycle in the Paleoproterozoic it was already 80%, and about 98% at the onset of the Brasiliano Cycle, in Neoproterozoic times.

The Transbrasiliano Lineament is a megafault zone and a probable megasuture active since the Neoproterozoic which separates a large northwestern continental mass, including the Amazonian Craton, from a southeastern continental mass, formed by a collage of cratonic fragments of different sizes of which the São Francisco Craton is the largest. When the crustal evolution of these two large continental masses is considered individually (see Figures 18 and 19), a few conclusions can be made:

1 - Old Archean rocks (older than 3.3 Ga) can be found at present only within the southeastern part of the South American Platform.

2 - On both continental masses, crustal evolution between 3.0 and 1.7 Ga is similar, suggesting that they were possibly contiguous within a Paleoproterozoic supercontinent.

3 - During Mesoproterozoic and Neoproterozoic times, the northwestern continental mass remained virtually unaffected by tectono-orogenic events, whereas the southeastern mass was broken up into smaller cratonic fragments, with the concomitant formation of regions of ocean floor. These smaller microcontinents and pieces of oceanic lithosphere took part in the agglutination and later fragmentation of at least two large supercontinents, Rodinia in the Mesoproterozoic, and Gondwana in the Neoproterozoic.

REFERENCES

Alkmin, F.F. and Marshak, S. (1998). Transamazonian orogeny in the Southern São Francisco Craton region, Minas Gerais, Brazil: evidence for Paleoproterozoic collision and collapse in the Quadrilátero Ferrífero. *Precambrian Research*, **90**, 29-58.

Almeida, F.F.M. de, Amaral, G., Cordani, U.G. and Kawashita, K. (1973). The precambrian evolution of the South American Cratonic Margin, South of the Amazon River. In: *The ocean basins and margins*, eds. Nairn, A.E.M. and Stehli, F.G., pp. 411-446. Chapter XI.

Almeida, F.F.M. (1977). O cráton do São Francisco. *Revista Brasileira de Geociências*, 7, 349-364.

Almeida, F.F.M. de, Hasui, Y., Brito Neves, B.B. and Fuck, R.A. (1981). Brazilian Structural Provinces: an Introduction. *Earth Sciences Review*, 17, 1-29.

Arndt, N.T. and Goldstein, S.L. (1987). Use and abuse of crust-formation ages. *Geology*, 15, 893-895.

Babinski, M., Chemale, F., Hartmann, L.A., Van Schmus, W.R. and Silva, L.C. (1995). *Acresção juvenil no bloco São Gabriel e* retrabalhamento crustal no cinturão Dom Feliciano durante o ciclo Brasiliano: evidências isotópicas. *VI Simpósio Sul Brasileiro de Geologia, Porto Alegre, RS, Brazil, Extended Abstracts*, 105-107.

Babinski, M., Chemale, F., Hartman, LA., Van Schmus, W.R. and Silva, L.C.-1997-U-Pb and Sm-Nd geochronology of the Neoproterozoic Granitic-Gneissic Dom Feliciano Belt, Southern Brazil. - J.S. Amer. Earth Sei., vol 10 (3-4): 263-274.

Barbosa, J.S.F. (1990). The granulites of the Jequié complex and Atlantic mobile belt, southern Bahia, Brazil. An expression of Archean – Proterozoic convergence. In: *Granulites and crustal evolution*, eds. Vielzeuf, D. and Vidal, pp. 195-221. Ph. Kluver Acad. Publishers.

Basei, M.A.S., McReath, I. and Siga Jr., O. (1998). The Santa Catarina granulite complex of southern Brazil: a review. *Gondwana Research*, 1, 383-392.

Bettencourt, J.S., Tosdal, R.M., Leite, J.R. and Payolla, B.L. (1999). Mesoproterozoic rapakivi granites of the Rondonia Tin Province, south-western border of the Amazonia craton, Brazil. I – Reconnaissance U-Pb geochronology and regional implications. *Precambrian Research*, **95**, 41-67.

Bossi, J., Campal, N., Civetta, L., Girardi, V.A.V., Mazzuchelli, M., Negrini, L., Rivalenti, G., Fragoso Cesar, A.R.S., Sinigoi, S., Teixeira, W. and Piccirillo, E. - 1993 - Early Proterozoic dike swarms from western Uruguai: *Geochemistry, Sm-Nd isotopes and petrogenesis.* - Chemical geology, vol 106: 263-277.

Brito Neves, B.B. and Cordani U.G. (1991). Tectonic evolution of South America during the Late Proterozoic. *Precambrian Research*, **53**, 23-40.

Brito Neves, B.B., Campos Neto, M.C. and Fuck, R. (1999). From Rodinia to Western Gondwana: an approach to the Brasiliano-Pan African Cycle and orogenic collage. *Episodes*, **22**(3), 155-166.

Campos Neto, M.C. and Caby, R. (1999a). Neoproteozoic high-pressure metamorphism and tectonic constraint from the nappe system south of the São Francisco Craton, southeast Brazil. *Precambrian Research*, **97**, 3-26.

Campos Neto, M.C. and Figueiredo, M.C.H. (1995). The Rio Doce Orogeny, Southeastern Brazil. *Journal of South American Earth Sciences*, **8**, 143-162

Cordani, U.G., Amaral, G. and Kawashita, K. (1973). The precambrian evolution of South America. *Geologische Rundschau*, **62**(2), 309-317.

Cordani, UG and Brito Neves, BB (1982). The geologic evolution of South America during the Archaean and Early Proterozoic. *Revista Brasileira de Geociências*, S. Paulo, **V12**(1-3), 78-88.

Cordani, U.G., Iyer, S.S., Taylor, P.N., Kawashita, K., Sato, K. and Mc Reath, I. (1992). Pb-Pb, Rb-Sr and K-Ar systematics of the Lagoa Real Uranium Province (South Central Bahia, Brazil) and the Espinhaço Cycle (ca. 1.5 – 1.0 Ga.). *Journal of South American Earth Sciences*, 5(1), 33-46.

Cordani, U.G., Sato, K. and Marinho, M.M. (1985). The geological evolution of the ancient granite-greenstone terrain of central-southern Bahia, Brazil. *Precambrian Research*, 27, 187-213.

Cordani, U.G., Teixeira, W., Tassinari, C.C.G., Kawashita K. and Sato, K. (1988). The growth of the Brazilian Shield. *Episodes*, 11(3), 163-167.

Cordani, U.G. and Sato, K. (1999). Crustal evolution of the South American Platform, based on Nd isotopic systematics on granitoid rocks. *Episodes*, 22(3), 167-173.

Cordani, U.G. and Torquato, J.R. (1981). Brazil-Africa geological links. *Earth Sciences Review*, 17, 155-176.

Correia C.T., Girardi V.A.V., Tassinari C.C.G. and Jost H. (1997)- Rb-Sr and Sm-Nd geochronology of the Cana Brava layered mafic-ultramafic intrusion, Brazil, and considerations regarding its tectonic evolution. *Revista Brasileira de Geociências*, 27: 163-168.

D'Agrella Filho, M.S., Pacca, I.G., Renne, P.R., Onstott, T.C. and Teixeira, W. - 1990 - Paleomagnetism of Proterozoic (1.01 to 1.08 Ga) mafic dikes in south-eastern Bahia State - São Francisco Craton, Brazil. - Earth Plan. Sci. Lett., vol 101: 332-348.

D'Agrella Filho, M.S. - 1999 - Contribuition of South American paleomagnetic date for the reconstruction of Proterozoic supercontinents: present states and future pespectives. - Aarhus Geoscience, vol 8: 27-35

Dalla Salda, L.H., Franzese, J.R. and Posadas, V.G. - 1992 - The 1800 Ma mylonite - anatectic granitoid association in Tandilia, Argentina. In: Basement Tectonics - R. Mason (ed.) ch. 7: 161-174.

Dalziel, W.D. (1997). Neoproterozoic-Paleozoic geography and tectonics: Review, hypothesis, environmental speculation. *Geological Society of America Bulletin*, 109(1), 16-42.

De Paolo, D.J. (1980). Crustal growth and mantle evolution: Inferences from models of element transport and Nd and Sr isotopes. *Geochimica et Cosmochima Acta*, 44, 1185-1196.

De Paolo, D.J. (1988). Age dependence of the composition of continental crust: evidence from Nd isotopic composition in granitic rocks. *Earth and Planetary Science Letters*, 90, 263-271.

Evans, D.A. (1998). True polar wander, a supercontinental legacy. *Earth and Planetary Science Letters*, 157, 1-8.

Hartnady, C., Joubert, P. and Stowe, C. (1985). Proterozoic Crustal Evolution in Southwestern Africa. *Episodes*, 8(4), 236-244.

Hoffman, P.F. (1991). Did the Breakout of Laurentia Turn Gondwanaland Inside-Out? *Science*, 252, 1409-1412.

Hurley, P.M., Almeida, F.F.M. de, Melcher, G.C., Cordani, U.G., Rand, J., Kawashita, K., Vandoros, P., Pinson, W.H. and Fairbairn, H.W. (1967). Test of continental drift by means of radiometric ages. *Science*, 144, 495-500.

Karlstrom, K.E., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W. and Åhäll, K. (1999). Refining Rodinia: Geologic Evidence for the Australia–Western U.S. connection in the Proterozoic. *GSA Today*, 9(10), 1-7.

Kotschoubey B., Hieronymus B., Rodrigues O.B. and Amaral R.T. 1996. Basaltos e serpentinitos da área da Serra do Tapa (PA). Prováveis testemunhos de um complexo ofiolítico pouco evoluído e desmembrado. Cong. Bras. Geol., 39, Salvador, SBG, Anais, vol.6, p.25-29.

Litherland, M., coord. (1986). The geology and mineral resources of the Bolivian Precambrian Shield. *Overseas Memoir - British Geological Survey*, 9, 153 p.

Marshak, S. and Alkmin, F.F. – 1989 – Proterozoic extension/ contraction tectonics of the Southern São Francisco craton and adjacent regions, Minas Gerais, Brazil: a kinematic model relating Quadrilétero Ferrifero, Sao Francisco basin and Cordilheira do Espinhaqo. – Tectonics, vol 8: 555-571.

Nutman, A.P., Cordani, U.G. and Sabaté, P. (1994). SHRIMP U-Pb ages of detrital zircons from the Early Proterozoic Contendas-Mirante supracrustal belt, São Francisco craton, Bahia, Brazil. *Journal of South American Earth Sciences*, 7, 109-114.

Pedrosa-Soares, A.C., Vidal, P., Leonardos, O.H. and Brito Neves, B.B. (1998). Late Proterozoic oceanic remnants in eastern Brazil: Further evidence and refutation of an exclusively ensialic evolution for the Araçuaí-West Congo orogen. *Geology*, 26, 519-522.

Pimentel, M.M. and Fuck, R.A. (1992). Neoproterozoic crustal accretion in central Brazil. *Geology*, 20, 375-379.

Pimentel, M.M., Whitehouse, M.J., Viana, M.G., Fuck, R.A. and Machado, N. (1997). The Mara Rosa arc in the Tocantins Province: further evidence for Neoproterozoic crustal accretion in Central Brazil. *Precambrian Research*, 81, 299-310.

Pinna, P., Jourde, G., Calvez, J.Y., Mroz, J.P. and Marques, J.M. – 1993 – The Mozambique Belt in northern Mozambique: Neoproterozoic (1100-850 Ma) crustal growth and tectogenesis, and superimposed Pan-African (850-550 Ma) tectonism – Prec. Research, vol 62: 1-59.

Preciozzi, F., Masquelin, H. and Basei, M.A.S. (1999a). The Namaqua / Grenville Terrane of Eastern Uruguay. *II South American Symposium on Isotope Geology, Actas, Cordoba, Argentine*, 338-340.

Sadowski, G.R. and Bettencourt, J.S. (1996). Mesoproterozoic tectonic correlations between eastern Laurentia and the western border of the Amazonian Craton. *Precambrian Research*, 76, 213-227.

Sato, K. (1998). *Evolução crustal da Plataforma Sul Americana, com base na geoquímica isotópica Sm-Nd.* Doctoral Thesis, Institute of Geosciences, University of São Paulo.

Schobbenhaus, C., Campos, D. de A., Derze, G.R. and Asmus, H.E. (1984). *Geologia do Brasil: Texto explicativo do Mapa Geológico do Brasil e Área Oceânica Adjacente.* D.N.P.M., Brasília.

Sollner, F. and Trouw, R.A.J. – 1997 – The Andrelandia depositional cycle (Minas Gerais/Brazil), a post-Transamazonian sequence south of the Sao Francisco Craton: evidence from U-Pb dating on zircons of a metasediment. – I. S. Amer. Karth Sci., vol 10 (1): 21-28.

Tassinari, C.C.G. and Macambira, M.J.B. (1999). Geochronological provinces of the Amazonian Craton. *Episodes*, 22(3), 174-182.

Tassinari, C.C.G., Cordani, U.G., Nutman, A.P., Van Schmus, W.R., Bettencourt, J.S. and Taylor, P.N. (1996). Geochronological systematics on basement rocks from the Rio Negro-Juruena Province (Amazonian Craton) and tectonic implications. *International Geology Review*, 38(2), 1161-1175.

Teixeira, W., Tassinari, C.C.G., Cordani, U.G., and Kawashita, K. (1989). A review of the geochronology of the Amazonian Craton: Tectonic implications. *Precambrian Research*, 42, 213-227.

Teixeira, W. and Figueiredo, M.C.H. (1991). An outline of Early Proterozoic crustal evolution in the São Francisco craton, Brazil: a review. *Precambrian Research*, 53(1/2), 1-22.

Teixeira, W., Carneiro, M.A., Noce, C.M., Machado, N. and Sato, K. (1996). Pb, Sr and Nd isotope constraints on the Archean evolution of gneissic-granitoid complexes in the southern São Francisco Craton, Brazil. *Precambrian Research*, 78(1/3), 151-164.

Teixeira, W., Renne, P.R., Bossi, J., Campal, N. and D'Agrella Filho, M.S. – 1999 – ^{40}Ar-^{39}Ar and Rb-Sr geochronology of the Uruguayan dike swarm, Rio de La Plata Craton, and implications for Proterozoic intraplate activity in western Gondwana. – Prec. Research, vol93: 153-180.

Thomaz Filho, A., Kawashita, K. and Cordani, U.G. (1998). A origem do Grupo Bambui no contexto da evolução geotectônica e de idades radiométricas. *Anais da Academia Brasileira de Ciências*, 70(3), 527-548.

Trompette, R. (1994). *Geology of Western Gondwanaland 2000 – 500Ma. The Pan-African/Brasiliano amalgamation of South America and adjacent Africa.* Balkema, Rotterdam, 350 p.

Unrug, R. (1996). The assembly of Gondwanaland. *Episodes*, 19, (1-2), 11-20.

Wegener, A. – 1912 – Die Entstehung der Kontinente. – Geol. Rundschau, vol 3 (4): 276-292.

Weil, A.B., Van der Voo, R., Mac Niocaill, C. and Meert, J.G. (1998). The Proterozoic supercontinent Rodinia: paleomagnetically derived reconstructions for 1100 to 800 Ma. *Earth and Planetary Science Letters*, 154, 13-24.

The Amazonian Craton

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The chapter was coordinated by Colombo C.G. Tassinari, and it was organized in two sections assembled by different authors. The text on the eastern part of the Amazonian Craton was prepared by Moacir J.B. Macambira, Jean M. Lafon and Colombo C.G. Tassinari. The western part was described by Jorge S. Bettencourt, Mauro C. Geraldes and Colombo C G. Tassinari. Jorge S. Bettencourt and Colombo C. G. Tassinari are responsible for the overall conclusions of this chapter.

The Amazonian Craton is one of the largest cratonic areas in the world. It underlies the northern part of South America, and covers an area of about 430 000 km^2. It is divided into two Precambrian shields: the Guaporé Shield and the Guiana Shield that are separated by the Paleozoic Solimões and Amazonas basins, (Fig. 1). The craton is surrounded by Neoproterozoic orogenic belts (Tucavaca in Bolivia, Araguaia-Cuiabá in central Brazil, and Tocantins in northern Brazil), and it has been relatively stable since 1.0 Ga ago. The cratonic area includes parts of Brazil, French Guiana, Guyana Suriname, Venezuela, Colombia and Bolivia.

The isotope studies and the definition of geochronological provinces provide a useful base from which to understand crustal evolutionary processes and their tectonic implications at continental scale. For this reason we have summarized the geochronology of the Amazonian Craton with the aim of describing its tectonic history during Precambrian times. We have commented on the isotope and geological data, emphasizing their geographical distribution in accordance with the geochronological provinces established in previous work (Cordani *et al.*, 1979; Teixeira *et al.*, 1989; Tassinari *et al.*, 1996; Tassinari and Macambira, 1999).

The geochronological provinces are defined, following the principles of Tassinari and Macambira (1999) as major zones within cratonic areas, where a characteristic geochronological pattern predominates, and the age determinations, obtained by different isotopic methods for different geological units, are very coherent. The divisions are made mainly on the basis of the age of the metamorphic basement and the geological characteristics. In general, a broad time-interval for the provinces was established because the geology of most of the Amazonian Craton is poorly known and the geological surveys have generally been on a regional scale, due to the dense vegetation. Therefore, the provinces mainly differ from each other in the age of their respective metamorphic terranes, lithological assemblages and their geological history.

Each geochronological province may contain anorogenic igneous rocks and sedimentary of widely different younger ages, in agreement with the orogenic history of the neighbouring areas. Furthermore, geochronological provinces may include some older preserved nuclei, when their tectonic evolution has an ensialic character or some younger metamorphic rocks produced by later reworking processes.

The geochronological provinces may include one or more orogenic events, within their respective period. The term orogeny is here considered as a period of metamorphic episodes accompanied by deformation, partial melting and syn-tectonic granitic intrusions, rather than in the wider usage of the term a complete orogenic cycle, involving subsidence, deposition of sediments, metamorphism, syn and post-tectonic magmatism and anorogenic episodes. In this way, within those areas better studied, such as the Serra dos Carajás, the southwestern regions of the Amazonian Craton, and part of French Guiana for which some detailed studies are available, it is possible to define several orogenies and distinct terranes within the same geochronological province, in like manner to the Grenville Province of North America.

The geographical boundaries between geochronological provinces in the Amazonian Craton have been reasonably well defined mainly by geochronological data with some geological and geophysical support, although some limits are still not well defined due to the overprint of age determinations and/or lack of reliable geological information. Therefore, some boundaries are still open to question, and detailed geological surveying and more precise geochronological data must be obtained to establish the precise location of the geochronological boundaries in the field.

There are several syntheses on the tectonic setting of orogenies affecting the Amazonian Craton, which can be divided along two different lines. The first line of reasoning follows authors such as Amaral (1974) and Almeida (1978), and proposes that Precambrian tectonics are characterized by platform reactivation and by ensialic orogenies, with sea-floor spreading and subduction being of lesser importance in the orogenesis. Hasui *et al.* (1984) and Costa and Hasui (1997) proposed a similar model for the evolution of the Amazonian Craton, mainly based on structural geology and geophysics. They considered the evolution of the Amazonian Craton as a whole by the diachronous formation of continental blocks or paleo-plates by collision during Archean and Paleoproterozoic times that resulted in the agglutination of a megacontinent. Crustal blocks include granite-greenstone terranes and medium-grade gneiss. The limits of the blocks are marked by shear belts including high-grade rocks from the tectonic extrusion of lower crust domains (granulite belts). The second line of reasoning was proposed by Cordani *et al.* (1979) and followed and modified by Tassinari (1981), Cordani and Brito Neves (1982), Teixeira *et al.* (1989), Tassinari *et al.* (1996) and Tassinari (1996). It is based on modern orogenic concepts that include continuous crustal accretion during Archean and Paleo-Mesoproterozoic times. This hypothesis, based on the predominance of calc-alkaline magmatism in the Proterozoic terranes, is strongly supported by isotope data. This chapter describes the geological history of the Amazonian Craton in line with the second of the two views.

The Amazonian Craton can be subdivided into six major geochronological provinces based on the age determinations, structural trends, relative proportions of rock-types, and some geophysical evidence (Tassinari and Macambira, 1999). The majority of the radiometric ages, which comprises about 3000 age determinations, was obtained by Rb/Sr, K/Ar, Sm/Nd and zircon U/Pb methods, although whole-rock and zircon Pb/Pb ages are also available. The recognized geochronological provinces of the craton (Fig. 1) comprise a stable Archean nucleus (Carajás-Iricoumé and Roraima blocks), which are included in the Central Amazonian Province (CAP), and Paleoproteorozic and Mesoproterozoic provinces such as Maroni-Itacaiúnas (2.25 - 1.95 Ga), Ventuari-Tapajós (2.0 - 1.8 Ga), Rio Negro-Juruena (1.8 - 1.55 Ga), Rondonian-San Ignacio (1.55 - 1.30 Ga) and Sunsas (1.30 - 1.0 Ga).

Sr, Pb and Nd isotope composition of igneous or orthogneissic rocks demonstrate that the Proterozoic crustal growth in the Amazonian Craton involved the addition of juvenile material as well as the reworking of the older continental crust. Part of the Maroni-Itacaiúnas and Rondonian-San Ignacio, and the whole of Ventuari-Tapajós and Rio Negro-Juruena provinces appear to have evolved through successive episodes of continental accretion with associated mantle-derived magmatism. By comparison, the Sunsas and part of the Rondonian-San Ignacio and Maroni-Itacaiúnas provinces may have been associated mainly with events involving continental collision. Sm/Nd model ages on granitoid samples from the Amazonian Craton indicate that about 30% of the continental crust was derived from the mantle during the Archean, and about 70% in the Proterozoic times. During the Proterozoic the main crustal *formation episodes took place around 2.0 Ga.*

Paleoproterozoic orogenies are typically developed in the border zone of the stable Central Amazonian Province, and are well represented in the eastern and northern part of this province by the Maroni-Itacaiúnas Province, and on the western side by the Ventuari-Tapajós Province. The paleoproteozoic belts included within the Maroni-Itacaiúnas Province weld three former microcontinents: the Archean Carajás-Iricoumé and Roraima blocks, and the Archean part of the West Congo Craton. The Ventuari-Tapajós Province seems to be slightly younger than Maroni-Itacaiúnas because its structural trends crosscut the structural pattern of the latter. The U/Pb, Rb/Sr, Sm/Nd and Pb/Pb age determinations have suggested that the Ventuari-Tapajós and Rio Negro-Juruena provinces evolved through successive magmatic arcs during the period from 1.95 to 1.55 Ga. The Rondonian-San Ignacio Province was developed through a magmatic arc phase between 1.55 and 1.4 Ga and thereafter by continental collision between 1.4 and 1.3 Ga. Finally, the Sunsas Province, composed mainly of metavolcano-sedimentary sequences and granitoid plutons, includes older terranes reworked between 1.3 and 1.0 Ga together with a small amount of juvenile material. Its evolution has been associated with the inversion of the marginal belt during continent-continent collision.

For purposes of discussion the crustal evolution of the Amazonian Craton will be divided into two parts based on geography: the eastern part, consisting of the Central Amazonian, Maroni-Itacaiúnas and Ventuari-Tapajós provinces, and the western part consisting of the Rio Negro-Juruena, Rondonian-San Ignacio and Sunsas provinces. This division is based on the fact that within the western part, the younger metamorphic and magmatic overprinting is more important than that which occurs on the eastern side.

Eastern Amazonian Craton

Central Amazonian Province (CAP)

The Central Amazonian Province is composed of the oldest continental crust of the Amazonian Craton that was not affected by the 2.2 - 1.9 Ga Transamazonian Orogeny. However, during the Paleoproterozoic it was the scene of expressive magmatic and sedimentary events. The basement of the CAP probably comprises a number of contrasting geological units in relation to their lithology, age and extent of geological knowledge. In the southeastern CAP, the Archean Carajás Metallogenetic Province represents the better-studied region of the Amazonian Craton. On the other hand, the western adjacent region and its continuity to the N of the Paleozoic Amazonas Synedise, is not well exposed; is poorly known and for which very little geochronological data are available. Taking into accounts these dissimilarities, and for the purpose of description and discussion of the geochronology, the CAP will be divided into two domains, separated by the Maroni-Itacaiúnas Province. The first domain is named Carajás-Iricoumé Block, which is subdivided into the Carajás and the Xingu-Iricoumé areas. The second domain consists of the Roraima Block. The geochronological pattern of the CAP *is summarized in Table 1.*

The Carajás-Iricoumé Block (Carajás Area)

The Carajás Area is the only well recognized and preserved Archean region of the Amazonian Craton. It represents the most important mineral province of Brazil, hosting deposits of iron, copper, gold, maganese, nickel, and others. The Maroni-Itacaiúnas Province to the N and Araguaia Belt to the E limit the area (Fig. 1). It is covered by the Phanerozoic sediments of the Parecis-Alto Xingu Basin to the S, and by the Paleoproterozoic volcanic rocks of the Uatumã Supergroup and sediments of the Gorotire Formation to the W. Of the regional geological surveys in the Carajás area, it is necessary to mention the synthesis made by the geologists of the Companhia Vale do Rio Doce-C.V.R.D. (Hirata *et al.*, 1982; DOCEGEO, 1988), besides those of Companhia de Pesquisas de Recursos Minerais-C.P.R.M. in the Grande Carajás Program (Araújo *et al.*, 1988; Araújo and Maia, 1991; Oliveira *et al.*, 1994; Macambira and Vale, 1997), summarized by Costa *et al.* (1995). Students and researchers from several Brazilian and foreign universities

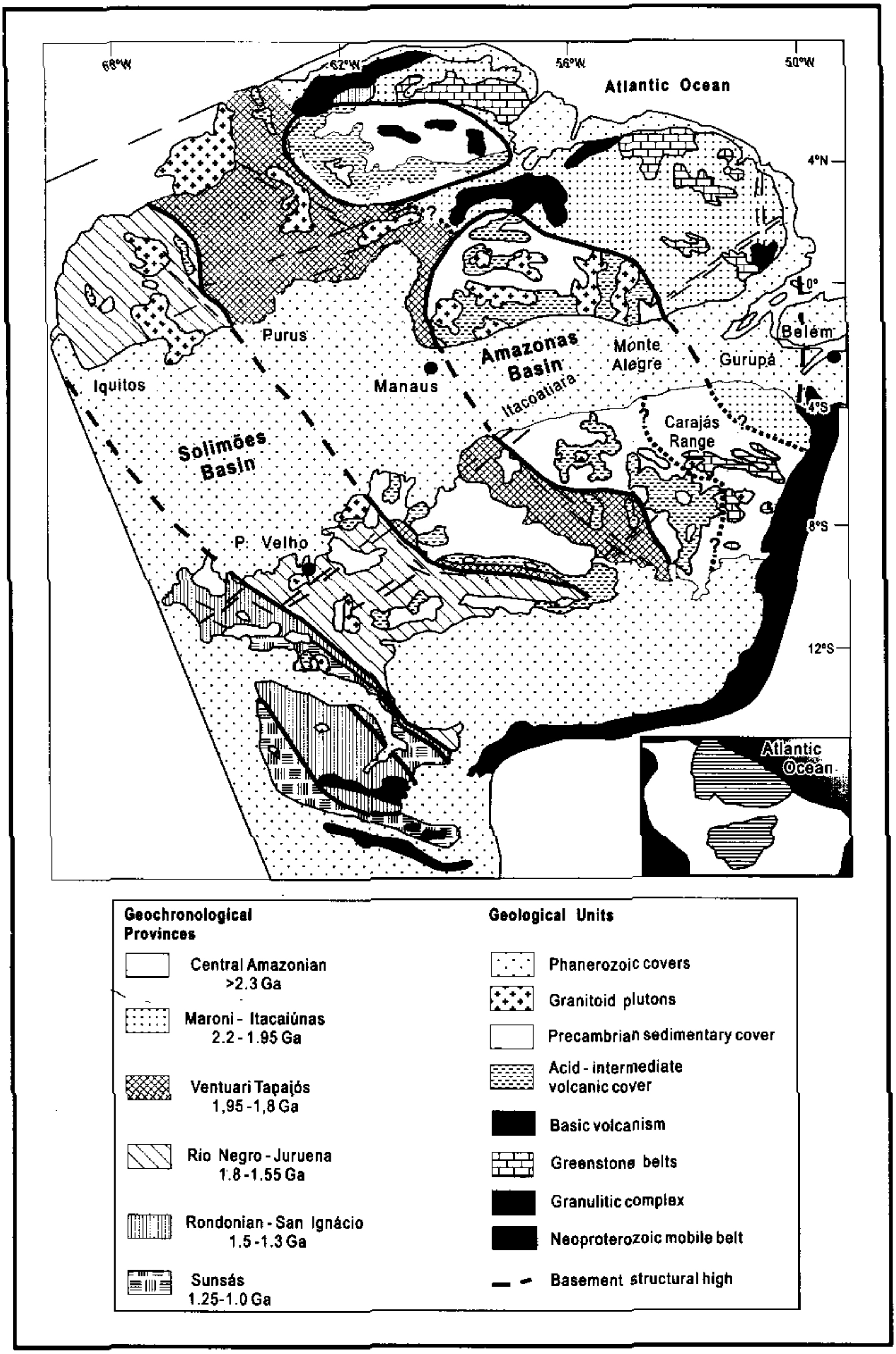


FIGURE 1 - Major geochronological provinces and main lithological associations of the Amazonian Craton (modified after Tassinari and Macambira, 1999).

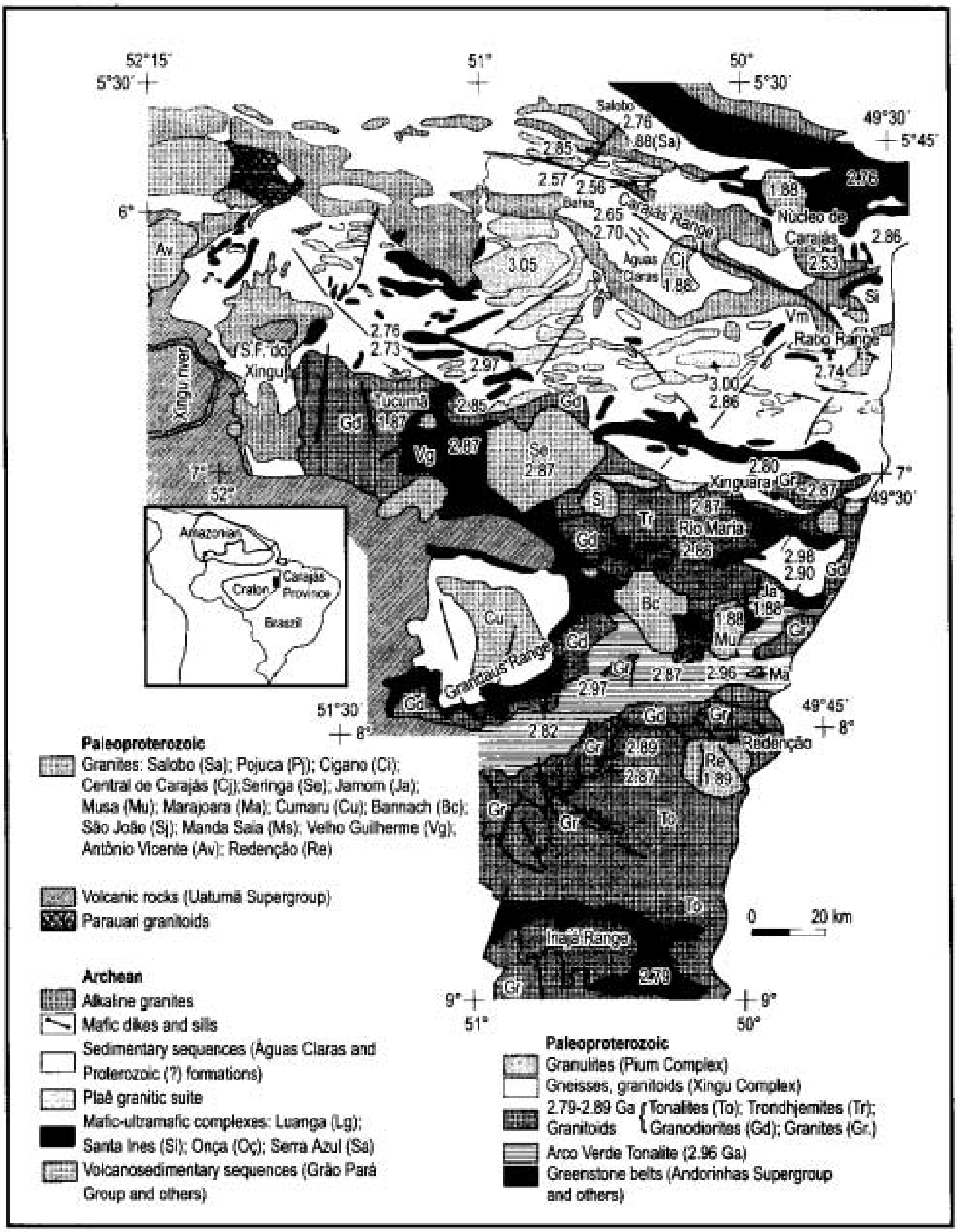


FIGURE 2 - Geology of the Carajás area, showing major unit locations and respective age determinations.

have also dedicated their studies to the CAP, including those attached to the of Universities of Pará, Brasília, São Paulo, Vale do Rio dos Sinos and Campinas.

Although a large amount of work has already been carried out in the Carajás area making a significant contribution to geological knowledge and leading to the development of geological models that are sometimes conflicting, important questions still remain open. The Carajás area was formed and tectonically stabilized in the Archean. Furthermore, it was only affected by a Paleoproterozoic extensive thermal event that was accompanied by the emplacement of granitic intrusions and felsic and mafic dykes. Phanerozoic mafic dykes have also been reported. The area was divided into three tectonic domains (DOCEGEO 1988; Costa *et al.*, 1995), named by the last-cited authors, as the Rio Maria granite-greenstone terranes, the Northern Itacaiúnas Shear Belt and the Southern Pau D´Arco Shear Belt (Fig.2). The three domains are roughly E-W structured, according to the regional foliation. The Rio Maria terranes are interpreted as a preserved nucleus whereas, at least part of both the semi-surrounding shear belts, are considered as the result of deformation and shearing of the units of the Rio Maria terranes. An important difference between the two shear belts is that the Itacaiúnas Belt also presents significant Neoarchean volcanism and plutonism whereas, in the Pau D´Arco Belt, the rock units appear to be lithologically and temporally similar to those of the Rio Maria terranes. The similarities between the Rio Maria terranes and the Pau D´Arco Belt led Althoff *et al.* (1991) and others to suggest that the southern belt could be considered as just an extension of the Rio Maria terranes.

The Itacaiúnas Belt is divided into two sub-domains: the E-W imbricated system, in the S, and the E-W, WNW-ESE and N-S transcurrent systems, in the N (Costa *et al.*, 1995). The Itacaiúnas Belt is mainly composed of high grade rocks (Pium Complex), gneiss (Xingu Complex), volcano-sedimentary sequences (*e.g.*, Grão Pará Group) and contemporaneous mafic-ultramafic complexes, and granitoid plutons (Plaque Suite, Estrela Granite and others). The oldest rocks of the Carajás area (Macambira and Lafon, 1995) and the Pau D´Arco Belt, consist of greenstone sequences, named the Andorinhas Supergroup (DOCEGEO, 1988), including TTG associations.

High grade terranes (Pium Complex)

According to Araújo and Maia (1991), the granulitic rocks of the Pium Complex occur as a number of elongated bodies (maximum length of 35 km), sub-parallel to the regional E-W foliation. They have so far been described only in the imbricated domain of the Itacaiúnas Belt, and were interpreted as fragments of lower crust emplaced along shear zones (Araújo *et al.*, 1988). The two main occurrences are those in the Pium River (in the central part of the Itacaiúnas Belt) and in the Catete River (in the western part of Itacaiúnas Belt), and they are mainly composed of mafic and felsic granulite, respectively. The mafic granulite seems to be older than the dominant felsic granulite, as xenoliths of the former have been found in the latter. The hypersthene calc-alkaline felsic granulite rocks are charnokite, enderbite and, subordinate charno-enderbite. The tholeiitic mafic granulite rocks are phaneritic, medium-grained, melanocratic, and isotropic hypersthene-plagioclase granulite (Araújo and Maia, 1991).

Rodrigues *et al.* (1992) analysed eleven samples of charnokite from the Catete River area and obtained an age of 3.05 ± 0.114 Ga (MSWD = 72) by the Pb/Pb on whole-rock method. The age was interpreted by the authors as the age of the crystallization of the protolith of the Pium Granulite. On the other hand, the analyses of the oscillatory zoned cores of the zircon of an enderbite from the Pium River area yielded, by the U/Pb (SHRIMP) method, an age of 3.002 ± 0.012 Ga (Pidgeon *et al.*, 1998). These authors share the same interpretation as the previous authors for the similar age, demonstrating that the two main occurrences of the Puim Granulite are coeval. Moreover, analysing the nebulously zoned rims of the zircon, Pidgeon *et al.* (1998) obtained, by the same method, an age of 2.861 ± 0.002 Ga, interpreted as dating the granulite facies metamorphism.

Granite-greenstone terranes

The granite-greenstone terranes of the Carajás area were first reported in the Rio Maria region (Cordeiro and Saueressig, 1980). A number of projects were carried out in these terranes that extended their occurrences over other areas (Hirata *et al.*, 1982; Macambira *et al.*, 1986; Medeiros *et al.*, 1987; DOCEGEO, 1988; Costa *et al.*, 1995).

The greenstone belt sequences

DOCEGEO (1988) proposed the term Andorinhas Supergroup to name all the greenstone sequences of the Carajás region. The unit is composed of mafic to ultramafic volcanic rocks (including komatiitic flows) interlayered with sediments (pellite, BIF, chert), at the base (Babaçu Group), grading from intermediate to felsic volcanic rocks associated with shale, greywacke and BIF, at the top (Lagoa Seca Group). The Babaçu Group is divided into the Igarapé Encantado and Mamão formations, whereas the Lagoa Seca Group is divided into the Fazenda do Quincas and Recanto Azul formations. Three geochemical series were defined in volcanic rocks: komatiite, low-K tholeiite and sodic calc-alkaline (Souza *et al.*, 1997). The E-W sequences, showing spinifex texture and pillow lava structure, are metamorphosed in the greenschist to amphibolite facies, and crosscut by shear zones, where the gold-mineralization associated with hydrothermal alteration are found (Huhn, 1991; Souza, 1994). The greenstone belts are covered by a pelitic sequence named by DOCEGEO (1988) as the Rio Fresco Group, and considered as Paleoproterozoic in age. However, Costa *et al.* (1995), taking into account the deformational similarities, considered the pelitic sequence as an upper part of the greenstone belt. Dating detrital zircon from this sequence, Macambira *et al.* (1998) reported ages of 3.2, 3.4 and 3.7 Ga, previously unrecorded in the rocks of this region.

Besides the Rio Maria greenstone belts, other undated units have been described in the Carajás area. In the southernmost part, these sequences are known as the Serra do Inajá Supergroup, which is divided into the Santa Lúcia and Rio Preto groups, and considered as chrono-correlated to the Andorinhas Supergroup (DOCEGEO, 1988). Costa *et* *al.* (1995) gave the name Tucumã Group to the greenstone sequences that occur in the homonymous region (central-western Carajás Area). These authors called Sapucaia Group the several elongated greenstone sequences of the imbricated domain of the Itacaiúnas Belt. They considered both Groups coeval with the Andorinhas Supergroup, but the Sapucaia Group was affected by a further deformational event. In the northeastern part of the Carajás Area, Hirata *et al.* (1982) reported the presence of a greenstone belt, named the Rio Novo Sequence. DOCEGEO (1988) proposed that this sequence had been an extension of the Andorinhas Supergroup, separated by further erosion, whereas Araújo and Maia (1991) believed that it was contemporaneous with the Grão Pará Group (2.76 Ga).

The zircon U/Pb results of the felsic volcanic rocks of the Lagoa Seca Group (Andorinhas Supergroup) of the Rio Maria region yielded ages of 2.904 + 0.029 / - 0.022 Ga (Macambira, 1992) and 2.979 ± 0.005 Ga (Pimentel and Machado, 1994). In the Tucumã region, a Pb/Pb evaporation age of 2.868 ± 0.008 Ga was obtained in zircon from a felsic volcanic rock of the Tucumã Greenstone Belt (Avelar *et al.*, in press). Minimum ages have been proposed for the greenstone belts through the dating of the zircon from the intrusive mafic-ultramafic layered complexes, such as that of Luanga (2.763 ± 0.006 Ga; Machado *et al.*, 1991) and that of Serra Azul (2.97 ± 0.007 Ga; Pimentel and Machado, 1994), in the Gradaus Ridge, to the W of the Rio Maria region. The neighbouring Guarapará Complex (olivine-gabbro, peridotite and dunite) seems to be contemporaneous with the Serra Azul Complex, as well as with the associated greenstone sequence.

The granitoid plutons

A voluminous set of Archean granitoid plutons and batholiths have been reported in the Rio Maria terranes. They have been named as the Arco Verde and Parazonia Tonalite, Rio Maria Granodiorite, Mogno Trondhjemite, and the Mata Surrão, Guarantã and Xinguara granites. Dall'Agnol *et al.* (1996) have classified these granitoid plutons according to geochemical criteria: the tonalite and trondhjemite are similar, but enriched in N_2O and depleted in K_2O in relation to the Archean trondhjemitic series. The granodiorite follows the trend of the Archean K_2O-moderate calc-alkaline series, whereas the granite bodies are highly fractionated calc-alkaline leucogranitoid similar to those associated with Archean late magmatic events. Costa *et al.* (1995) have applied the same terms used in the Rio Maria region to name similar bodies mentioned in the southern Redenção and western Tucumã regions. In the Rio Maria region, field relationships and geochronological data have shown that these granitoid plutons are younger than the greenstone belt sequences. The exception is the Arco Verde Tonalite, which is contemporaneous with the greenstone belts. The granitoid plutons, as well as the greenstone sequences, were deformed by a WNW-ESE to E-W ductile deformation that generated mylonitic zones (Dall'Agnol *et al.*, 1996). The younger granitoid bodies are considered to be syn-tectonic, showing evidence for contact metamorphism in the country rocks.

The Arco Verde Tonalite (Althoff *et al.*, 1991) presents an age of 2.957 + 0.025 / - 0.021 Ga. It is clearly intruded by the 2.87 Ga Mata Surrão Granite. Detrital zircon crystals from a greywacke of the adjacent Lagoa Seca Greenstone Belt, with a zircon U/Pb age of 2.971 ± 0.018 Ga (Macambira, 1992), were interpreted as having come from the Arco Verde Tonalite. The similarity of the morphology, internal structures and chemistry between both populations of zircon support this hypothesis. According to the last-named author, the Arco Verde Tonalite could represent part of a sialic margin of the greenstone basin. The greywacke resulting from the erosion of the Arco Verde Tonalite had already been deposited when the Lagoa Seca Group was being formed.

The available geochronological pattern constrains a short interval, around 2.87 Ga, for the Archean younger granitoid emplacement in the Rio Maria region (Macambira, 1992; Pimentel and Machado; 1994; Lafon *et al.*, 1994). Similar granitoid plutons of the central-southern region of Carajás seem to be coeval, taking into account the experimental errors and the constraints of the methods:

- Cumaru Granodiorite (Gradaus region): 2.817 ± 0.004 Ga (Pb/Pb on zircon; Lafon and Scheller, 1994);
- Rio Maria Granodiorite (Tucumã region): 2.852 ± 0.016 Ga (Pb/Pb on zircon; Avelar *et al.*, in press);
- Mata Surrão Monzogranite (Redenção region): 2.894 ± 0.019 Ga and 2.798 ± 0.028 Ga (Pb/Pb on whole-rock; Barbosa and Lafon, 1996);
- Arco Verde Orthogneiss (Redenção region): 2.872 ± 0.025 Ga (Pb/Pb on whole-rock; Barbosa and Lafon, 1996).

The evolution of the Carajás granite-greenstone terranes terminated with the intrusion of these granitoid plutons, which established a short episode (<150 Ma) of continental crust formation. On the other hand, Sm/Nd mantle-depleted model ages obtained in seven Archean granitoid samples from Rio Maria region (2.73, 2.86, and five samples ranging from 2.95 to 3.04 Ga), as well as their ε_{Sr} and ε_{Nd} values suggest that there was a short interval between the moment of mantle-extraction and the emplacement of these granitoid plutons (Sato and Tassinari, 1997; Dall'Agnol *et al.*, 1999a).

Regional Basement Rocks

The regional basement rocks were named by Silva *et al.* (1974) as the Xingu Complex, which are mainly composed of polymetamorphosed granodioritic rocks occurring over a large area of the southern part of the Amazonian Craton. With the advance of geological knowledge, new units were defined in the complex and its domain became reduced. A good example is the Rio Maria region, where the complex presently is no longer recognized as such. In the Itacaiúnas Shear Belt, the Xingu Complex is still an important stratigraphic unit, comprising gneiss, granitoid and amphibolite, and has been considered as the regional basement (Costa *et al.*, 1995).

The only U/Pb zircon ages available for the Xingu Complex are those obtained by Machado *et al.* (1991), in the northeastern Carajás Area. Analyzing four samples (amphibolite, gneiss and leucosome) of the same outcrop, they obtained similar ages (2.859 ± 0.002 Ga and 2.851 ± 0.004 Ga), interpreted as dating the last migmatization event affecting the region. Macambira and Lafon (1995) remarked that these ages are similar to those of the granitoid of the Rio Maria region. Costa *et al.* (1995) admitted that at least part of the Xingu Complex gneiss could be the result of the reworking of granitoid, similar to that preserved in the Rio

Maria region. Sm/Nd data support this hypothesis, since the model ages (T_{DM}), obtained from two samples (gneiss and tonalite) from the northeastern CAP, are 3.03 and 2.98 Ga, similar to those of the Rio Maria Granodiorite (Sato and Tassinari, 1997). Additionally, the authors observed that the Sr initial ratios and ε_{Nd} values suggest a short crustal residence for the material source of these granitoid plutons. On the other hand, the protolith (3.0 Ga) and the metamorphism (2.85 Ga) of the Pium Granulite are coeval with the mantle extraction and the emplacement of the Xingu Gneiss, suggesting that the same regional event generated both rocks. Another interesting piece of data was obtained by Machado *et al.* (1991), who used a zircon age of 2.851 ± 0.004 Ga for the Cascata Gneiss (base of the overlying Salobo Group) to propose that the group should be considered as part of the Xingu Complex. The data was further reconsidered by Lindenmayer *et al.* (1995), who interpreted the gneiss as a part of the complex. So, if this is confirmed, it represents another result indicating an age of 2.85 Ga for the Xingu Complex.

In addition, a Rb/Sr isochron age was obtained in gneiss considered as belonging to the Xingu Complex (Cunha *et al.*, 1981), from the Inajá Ridge region in the southernmost part of the Carajás area. The age of 2.696 ± 0.158 Ga (Sr initial ratio = 0.701), interpreted as that of the emplacement of the protolith of the gneiss, is not different, taking into account the experimental errors, from those of the other Archean granitoid plutons referred to so far.

Metavolcano-sedimentary sequences (Northern Carajás Area)

The Neoarchean metavolcano-sedimentary sequences of the northern Carajás area host the more important mineral deposits of the province, which have been described in the literature. The sequences form a WNW-ESE synclinal belt underlain by the Xingu Complex and cut by Paleoproterozoic granite intrusives and mafic dykes (Beisiegel *et al.*, 1973; DOCEGEO, 1988). The sequences are heterogeneously affected by several different types of overprinting (igneous crystallization, hydrothermal alteration, contact metamorphism, regional deformation, shearing recrystalization) and in general are poorly exposed. These aspects have made it difficult to reconstruct adequately the geological evolution of the sequences. In any event, they are lithologically and temporally different from those of the Andorinhas Supergroup.

Araújo and Maia (1991) redefined the Grão Pará Group as equivalent to the Itacaiúnas Supergroup (DOCEGEO, 1988), but including the sediments of the Rio Fresco Group. The group was thus divided into three formations (from base to top):

• Parauapebas Formation - mainly composed of basalt and dacite, with subordinated rhyolite, metamorphosed in the greenshist facies;
• Carajás Formation - essentially composed of BIF with subordinated jaspilite and very rare limestone units;
• Águas Claras Formation - composed of psammite and pelite with a subordinated chemical contribution, undeformed in the center of the belt. The sandstone beds generally are conglomeratic, subarkosic and locally brecciated. Nogueira *et al.* (1995) divided the Águas Claras Formation into the Lower Member (marine) and Upper Member (littoral and fluvial). Sequences similar to the Grão Pará Group have been reported to the N of the Carajás region including the Tapirapé, Misteriosa and Buritirama groups (Costa *et al.*, 1995).

Reliable geochronological data have been very important in the establishment of the contemporaneity of the volcano-sedimentary sequences of the northern Carajás area at *c.* 2.76 Ga (Wirth *et al.*, 1986; Machado *et al.*, 1991; Trendall *et al.*, 1998). From indirect dating, it was also demonstrated that the BIF are also coeval with the volcano-sedimentary sequences (Macambira, 1996; Trendall *et al.*, 1998). Dating of zircon from mafic dykes crosscutting the Lower Member of the Águas Claras Formation indicates that these sediments are older than 2.645 ± 0.012 Ga (Dias *et al.*, 1996) or 2.708 ± 0.037 Ga (Mougeot *et al.*, 1996). Zircon crystals from the Upper Member and aged 2.681 ± 0.005 Ga are interpreted as coming from a syn-depositional volcanism (Trendall *et al.*, 1998), indicating a minimum age for the sedimentation.

There is no consensus on the tectonic environment of the formation and evolution of the Carajás Basin. Some defend continental rifting (Gibbs *et al.*, 1986; DOCEGEO, 1988), whereas others propose an island arc model (Dardenne *et al.*, 1988; Teixeira and Eaggler, 1994). According to Araújo *et al.* (1988) and Araújo and Maia (1991), the structural evidence does not agree with the more recent hypotheses. For these authors, the Carajás Basin is the result of a transcurrent process and was filled by the volcano-sedimentary sequences (Grão Pará Group) in the distensive phase. In the inversion phase, the sequences were separated in lenses, which were imbricated. At that time, the southern Carajás area was tectonically stable.

Neoarchean intrusive bodies

Scattered mafic and felsic intrusive bodies occur in the Itacaiúnas Belt, crosscutting the Xingu Complex and the metavolcano-sedimentary sequences. They are generally elongated according to the E-W regional foliation, and interpreted as having a syn to tardi-tectonic emplacement. Some of them have been studied and dated. They will be described according to their similarities.

Plaquê Granitic Suite

Stratum-like granitic bodies, named the Plaquê Suite by Araújo *et. al.* (1988), have been reported in the Itacaiúnas Belt, especially in the imbricated domain. They are biotite and/or muscovite granite showing a varied degree of deformation, more intense in the border facies and considered as the result of the friction of a crustal collision during the development of the Itacaiúnas Belt (Araújo and Maia, 1991). Zircon crystals from a granitic body in the Tucumã region were dated at 2.736 ± 0.024 Ga (Avelar *et al.*, in press), and interpreted as the age of the Itacaiúnas Belt structuration.

A-type granite

Some foliated alkaline granitic bodies have been only identified in the transcurrent domain of the Itacaiúnas Belt crosscutting and metamorphosing the volcano-sedimentary sequences. The best studied granitic intrusive is the 2.53 Ga Estrela Granitic Complex (Rb/Sr isochron, Barros *et al.*, 1992) in the northeastern Carajás area. Similar bodies are the "Old Salobo" Granite (Lindenmayer *et al.*, 1995) dated at 2.573 ± 0.002 Ga (U/Pb on zircon, Machado *et al.*, 1991) and the deformed "Old Pojuca" Granite of the Itacaiúnas River dated at 2.525 ± 0.038 Ga (Pb/Pb on zircon, Souza *et al.*, 1997). Recently, Huhn *et al.* (1999) dated zircon from the alkaline Planalto Granite (Rabo Ridge) at 2.747 ± 0.002 Ga. Zircon from an associated dioritic intrusive body yielded an age of 2.738 ± 0.006 Ga, with an inherited zircon of 2.953 ± 0.002 Ga, enlarging the established interval for the occurrence of the Archean alkaline granite in the Carajás area. It is also interesting to note that the alkaline granite intrusives are contemporaneous with the tectono-metamorphic events (2.58 - 2.5 and 2.77 - 2.73 Ga) proposed by Machado *et al.* (1991) as having affected the northern part of the studied area.

Mafic-ultramafic layered complexes

Zircon crystals from the Luanga Complex, a sill-like differentiated intrusion composed of chromite bearing bronzitite, norite and leucogabbro (Medeiros Filho and Meirelles, 1985), were dated by Machado *et al.* (1991) at 2.763 ± 0.006 Ga. The age of the surrounding Rio Novo greenstone belt is unknown, but Oliveira *et al.* (1994) observed a great deformational contrast between both units considered as coeval. The complex is well preserved, showing only brittle deformation, whereas the greenstone is intensely deformed and metamorphosed in the greenschist facies. So, if the greenstone is 2.76 Ga old, the complex emplacement occurred in an immediate distensive event as was proposed for the Grão Pará Group by Araújo and Maia (1991).

In the Tucumã region, Macambira and Vale (1997) recognized a set of E-W elongated mafic to ultramafic bodies and proposed the term Cateté Intrusive Suite to name these complexes. The layered complex of Serra da Onça (serpentinite, piroxenite and gabbronorite) is the best known (Macambira, 1996). A Sm/Nd (whole-rock and mineral) age of 2.378 ± 0.055 Ga (MSWD = 3.9) was proposed for its emplacement, which could have occurred in a distensive system, taking into account that the body is undeformed and unmetamorphosed. The ε_{Nd} of the samples indicate a mantle source. A U/Pb (SHRIMP) age of 2.763 ± 0.006 Ga was obtained by Lafon and others (personal communication) on zircon from the Serra da Onça Complex. This age, which is the same as that of the Luanga Complex, could indicate that the complex, after its formation was stored in the low level of the continental crust, and during the 2.4 Ga distensional event ascended to the higher crustal level. This contemporaneity, as well as the lithological and structural similarities strongly suggests that the Serra da Onça and Luanga complexes are products of the same tectonic and petrogenetic process.

Among other mafic-ultramafic bodies of the imbricated domain of the Itacaiúnas Belt, Araújo and Maia (1991) referred to the Vermelho Complex, SW of the Rabo Ridge, hosting an important Ni ore deposit. It is composed of serpentinized dunite and peridotite metamorphosed at the greenschist facies. No geochronological data are avalaible for the complex, but the authors proposed a Paleo to Mesoproterozoic age. The Santa Ines Gabbro is another body showing the same NE-SW orientation as the Vermelho Complex, but these rocks have received little study.

Paleoproterozoic granite plutons and associated rocks

Paleoproterozoic granitic plutons and batholiths and associated felsic and mafic dykes are widespread in the Carajás area indistinctly crosscutting the Archean units. They are undeformed, high-level granite bodies emplaced *in a rigid crust and containing xenoliths from the country* rocks that are thermally metamorphosed (Dall'Agnol *et al.*, 1997). They are mainly composed of syenogranite and monzogranite with subordinated alkali-feldspar granite and granodiorite. Geochemically, they are alkaline, metaluminous and similar to the A-type and within-plate granite bodies (Dall'Agnol *et al.*, 1994). Dall'Agnol *et al.*, (1997) proposed a subdivision of the granite plutons into three groups, according to their magnetic susceptibility (MS), geochemistry and metallogenesis:

- The high MS, magnetite bearing Jamon and Musa granite *bodies, locally W-mineralized;*
- The moderate MS, Serra dos Carajás, Cigano and Pojuca granite bodies, locally with Cu and Mo mineralization;
- The low MS, Antonio Vicente, Velho Guilherme, Mocambo and Benedita granite plutons, frequently with Sn-mineralization.

The estimated time for the emplacement of the Paleoproterozoic granite of the Carajás area is well constrained at 1.88 Ga by U/Pb zircon dating (Wirth *et al.*, 1986; Machado *et al.*, 1991). These granite plutons are interpreted as the result of an extensive thermal event (Costa *et al., 1995*) probably responsible for the warming of the region that induced the *c.* 2.0 Ga Rb/Sr and K/Ar "Transamazonian" ages of some country rocks (Macambira and Lafon, 1995). The granite plutons show high Sr initial ratios (0.707 to 0.715, Macambira *et al.*, 1990) and, some from the Rio Maria region, show negative ε_{Nd} (- 9.3 to - 10.0, Dall'Agnol *et al.*, 1999a) suggesting an origin by the anatexis of crustal rocks, probably induced by underplating or intrusion of mantle-derived mafic magma (Dall'Agnol *et al.*, 1994). According to Nd isotope data and geochemical modeling, the 2.87 Ga quartz diorite of the Rio Maria region have the adequate composition to generate the Musa and Jamon granite plutons, as well as the associated dacite porphyry dykes (Dall'Agnol *et al.*, 1999a). However, Archean inherited zircon, as old as 3.2 Ga, found in these granite bodies (Machado *et al.*, 1991), indicate additional contribution or contamination of the granitic magma.

Rock units younger than 1.9 Ga

Due to the lack of clear field evidence and geochronological data, some geological units, such as certain sedimentary sequences and mafic dykes of the Carajás area, do not have a well established stratigraphic and temporal positions and could be younger than the 1.9 Ga granite. DOCEGEO (1988) considered all the sedimentary sequences of the area (Rio Fresco Group), covering the Archean units and locally cut by the Paleoproterozoic granite, of an age between 2.0 and 1.8

Ga. However, other authors (Figueiras and Villas, 1982; Ramos *et al.*, 1984) do not accept this regional correlation, since these sedimentary sequences show important differences according to the region of occurrence, stratigraphy, degree of deformation, and presence of coal. For Ramos *et al.* (1984), the coal layer of 3 m thick enclosed in a sedimentary sequence of the Fresco River Basin is impossible to find in a Precambrian sequence. In the Northern Carajás Area, the Aguas Claras Formation has been proved to be Archean but, other sequences, such as the Gorotire Formation, are believed to be younger (Costa *et al.*, 1995).

A number of mafic dykes indistinctly crosscutting the Archean units are widespread in the area. Some of them are Archean, probably associated with the volcanism of the Carajás Basin, whereas others are contemporaneous with the Paleoproterozoic granitic magmatism. However, some of these dykes have a Phanerozoic age (560, 500 and 225 Ma) as is the case of the dolerite dykes occurring in the Carajás Ridge according to the K/Ar dating (Gomes *et al.*, 1975; Cordani *et al.*, 1984; DOCEGEO, 1988).

Carajás - Iricoumé Block (Xingu – Iricoumé Area)

The Xingu-Iricoumé area is a NW-SE trending domain parallel to the younger Maroni-Itacaiúnas and Ventuari-Tapajós provinces. It is limited by the Carajás area to the SE; by the Central Guiana Belt to the NW; and divided by the Phanerozoic Solimões-Amazonas basins. The domain is poorly known, and the very few available geochronological results were mainly obtained by the Rb/Sr and K/Ar methods. The area is composed of Paleoproterozoic plutonic, volcanic and sedimentary rocks, which crosscut and rest unconformably on an undated basement complex. A pre-Transamazonian age, older than 2.3 Ga, is proposed for the basement, taking into account some Archean Nd model ages (T_{DM}) and the presence of granitoid plutons of *c.*1.96 Ga, intrusive in the northern Central Amazonian Province.

The oldest granitoid plutons of the Xingu-Iricoumé area are the calc-alkaline Água Branca Monzogranite (1.96 to 1.91 Ga; Santos and Reis Neto, 1982; João *et al.*, 1985; Almeida *et al.*, 1997), of the northernmost part of the domain. Additionally, there are some granitoid bodies included in the Parauari Granitic Suite (1.921 ± 0.069 Ga; Macambira, 1992), situated close to the Xingu River. These 1.96 Ga old granitoid plutons are interpreted as post-collisional in relation to the Transazonian Orogeny (João *et al.*, 1985), but those of the Xingu River are also considered as the first product of a distensive regional event (Costa *et al.*, 1995). Recently, Faria *et al.* (1999) reported that some granitoid plutons occurring in the southeastern part of the State of Roraima, believed to be the Água Branca Monzogranite, are peraluminous, similar to the S-type granite, and proposed the term Igarapé Azul Granite to name these rocks.

The Uatumã Supergroup (Santos, 1982) is composed of felsic to intermediate volcanic and plutonic rocks. In the northern area (Iricoumé), they are called Iricoumé and Mapuera groups, whereas those of the southern area (Xingu) are named the Iriri and Maloquinha/Rio Dourado groups, respectively. Zircon from the Iricoumé Rhyodacite collected from the tin Pitinga Mine yielded an age of 1.966 ± 0.009 Ga (Schobbenhaus *et al.*, 1994), similar to the age of the Água Branca Monzogranite. In the southern part of the area, a similar age (1.888 ± 0.002 Ga and 1.888 ± 0.007 Ga) was obtained on zircon from the Iriri Rhyolite of the Tapajós and Jamanxim rivers by Dall'Agnol *et al.* (1999b) and Moura *et al.* (1999), respectively. The former authors characterized a typical A-type signature for the studied fayalite-hedembergite rhyolite. Analyzing rhyolite and andesite from a region close to the São Felix do Xingu (Xingu River), Teixeira *et al.* (1998) obtained a similar age of 1.875 ± 0.079 Ga by the Pb/Pb on whole-rock method.

Sub-alkaline to alkaline granite showing similarities with the A-type and rapakivi granite, and interpreted as anorogenic (Dall'Agnol *et al.*, 1994), is widespread in the Xingu-Iricoumé area. In the northernmost domain, some of these bodies are highly mineralized with respect to Sn and intrusive in the Iricoumé Group. Examples include the Madeira (1.834 ± 0.006 Ga; Fuck *et al.*, 1993) and Água Boa granite bodies of the Pitinga tin mine, or the Água Branca Monzogranite and the Moderna Granite (1.814 ± 0.027 Ga; Santos *et al.*, 1997). In the southern area, very few geochronological data are available for these granite plutons. However, in the Tapajós Gold Province, enclosed in the adjacent Ventuari-Tapajós Province, recent geochronological results were obtained for the Uatumã Supergroup and Parauari Granite (Vasquez *et al.*, 1999; Lamarão *et al.*, 1999). These data indicate at least two Paleoproterozoic volcano-plutonic events: 1.89 - 1.88 Ga and 2.0 - 1.98 Ga. The older event is calc-alkaline, whereas the younger is sub-alkaline to alkaline, but a calc-alkaline plutonism is also reported, suggesting that a review of the terms Iriri Volcanism and Parauari Granite is needed. On the other hand, the Nd model ages (T_{DM}) indicate values between 2.6 and 2.5 Ga for the source of rhyodacite and intrusive granite from the Iriri-Xingu region (Sato and Tassinari, 1997). These Archean signatures contrast with the Paleopreoterozoic Nd model ages (T_{DM}) of the rocks of the neighbouring Tapajós Gold Province, which is included in the younger Ventuari - Tapajós Geochronological Province.

The Roraima Block

The Roraima Block occurs in the northern part of the AC and the 2.2 - 1.95 Ga Maroni-Itacaiúnas Province separates it from the Iricoumé-Carajás Block. It is completely covered by the 2.0 - 1.95 Ga Surumu acid and intermediate volcanic rocks (Schobbenhaus *et. al.*, 1994), and by the sedimentary sequences of the Roraima Group, which overlie the Surumu volcanic rocks.

The Roraima Group consists of a variety of sedimentary rock-types, which mainly include sandstone, feldsphatic sandstone, conglomerate and dark shale. The deposition of the Roraima Group was developed in the following principal environments: fluvial, deltaic, coastal lagoon, beach and shallow marine environments (Ghosh, 1981). The paleocurrent measurement on sedimentary sequences of the Roraima Group (Ghosh, 1981), suggest a northeasterly, easterly and southeasterly source of sediments, probably from areas occupied now by the Maroni-Itacaiúnas Belt.

Interbedded with the middle and upper Roraima sediments, sills of mafic rocks and zones of pyroclastic volcanic rocks dated at 1.65 Ga (Priem *et al.*, 1973) are observed. Furthermore the Roraima Group is intersected by several 1.88 - 1.6 Ga mafic sills and dykes (Snelling *et. al.*, 1969; Hebeda *et. al.*, 1973; Teixeira, 1978). Therefore the Roraima sediments may be at least as old as 1.88 Ga and their sedimentation occurred until at least 1.6 Ga.

An age older than 2.3 Ga for the Roraima Block metamorphic basement is here assumed based on the fact that the block is covered by ancient unmetamorphosed acid and intermediate volcanic rocks (1.95 Ga), which have the same age as the neighbouring high-grade metamorphic terranes of the Maroni-Itacaiúnas Province. Thus it is possible to interpret that the Roraima Block acted as a stable foreland to the marginal Paleoproterozoic Maroni-Itacaiúnas Belt.

Maroni - Itacaiúnas Province (MIP)

The Central Amazonian Province is surrounded to the N and NE by the 2.2 - 1.95 Ga MIP, a characteristic of which is a large exposure of metavolcanic and metasedimentary units, deformed and metamorphosed in the greenschist to amphibolite facies. In addition to which there occur granulite facies rocks and and gneiss-migmatite terranes. The southern boundary between the CAP and the MIP, in the northern part of the Serra dos Carajás area, is still uncertain, due to the lack of reliable geological and geochronological information (Table 2).

The Maroni Itacaiúnas Province extends over the eastern part of the Guiana Shield. In the aftermath of the original paper of Cordani *et al.* (1979), several synthesis have delineated the principal features of the MIP (Cordani and Brito Neves, 1982; Bosma *et al.*, 1983; Teixeira *et al.*, 1989; Gibbs and Barron, 1983, 1993). Most of the data on the MIP have been acquired before the 80s. Since that time, only a small number of works have been carried out; most of them in the eastern part of MIP (French Guiana and northern Brazil). New geochronological contributions including Sm/Nd, U/Pb and Pb/Pb results have recently improved the knowledge on the MIP (Gaudette *et al.*, 1996; Tassinari 1996; Sato and Tassinari, 1997; Fraga *et al.*, 1997; Lafon *et al.*, 1998; Vanderhaeghe *et al.*, 1998).

The geographical extension of the MIP includes the eastern part of Venezuela, the Guyana, Suriname, French Guiana and the easternmost part of northern Brazil (Amapá, northern Pará and northeastern Roraima). The MIP represents a widespread domain strongly marked by the Transamazonian Orogeny (2.2-1.95 Ga), which consists of large extensions of Paleoproterozoic crust and some remnants of Archean crust. The MIP covers most of the northeastern part of the Guiana Shield and also includes the northeasternmost part of the Guaporé Shield, S of the Solimões-Amazonas basins, where Paleoproterozoic Rb/Sr and K/Ar ages have been obtained on metamorphic sequences (Santos *et al.*, 1988). The limits of the MIP with adjacent provinces are not well constrained for geographical reasons (rainforest cover, and difficulties of field access) and to the lack of geochronological data. To the N, the Orinoco sedimentary basin limits the MIP. To the W, post-Transamazonian sedimentary cover and widespread Paleo and Mesoproterozoic igneous rocks (Roraima Group and Uatumã Group) overlie the MIP units. In the State of Roraima, the magmatic rocks of the Ventuari Tapajós Province limit the MIP. To the S, in the Guaporé Shield, the limit with the Central Amazonian Province has been estimated to lie to the N of the Carajás Archean Province (Cordani *et al.*, 1984).

The main tectonic features consist of roughly WNW-ESE structural trends from Venezuela to the State of Amapá in Brazil. NE to ENE trending structures are also present. In northeastern Venezuela, the NE-SW striking Guri Fault limits the Archean terranes of the Imataca Complex from the Paleoproterozoic granite-greenstone sequences and gneissic complexes. In Suriname, Guyana, and in northern Roraima, high-grade metamorphic belts are delineated by SW-NE oriented structures that crosscut the E-W trending structures of the granite-greenstone terranes (Bosma *et al.*, 1983).

The geochronological pattern of the Maroni-Itacaiúnas Province is very complete and concordant throughout the whole province, suggesting that its evolution took place during a major event in the Paleoproteorozic between 2.25 and 1.95 Ga (Table 2). In turn, remnants of some older Archean basement within the MIP have been identified. They generally consist of high-grade polymetamorphic rocks, such as the allocthonous (>3.0 - 2.0 Ga) Imataca Complex, in Venezuela (Montgomery and Hurley, 1978; Teixeira *et. al.*, 1999), and the exotic (2.9 - 2.6 Ga) Cupixi terranes in Amapá, Brazil (Lima *et al.*, 1986) that show a strong Paleoproteorozic metamorphic overprint. These ancient nuclei, together with 2.0 - 1.9 Ga Rb/Sr age determinations with high Sr initial ratios of 0.710 obtained on metamorphic rocks from the southern part of the MIP (Santos *et. al.*, 1988), suggest a partial ensialic character for the tectonic evolution of the Maroni-Itacaiúnas Province.

The isotope data for the high-grade metamorphic rocks, considered as the Central Guiana Granulitic Belt, including the Falawatra and Kanuku groups, indicate zircon U/Pb ages and Rb/Sr isochron ages between 2.1 to 1.9 Ga (Priem *et al.*, 1978), and Sm/Nd mantle-depleted model ages ranging from 2.2 to 2.0 Ga (Ben Othman *et al.*, 1984; Vignol, 1987). The geochronological results clearly indicate that the granulitic terranes were separated from the upper mantle during Paleoproteorozic times.

The Rb/Sr, Pb/Pb, Sm/Nd and U/Pb age determinations obtained for syntectonic granitoid plutons and gneissic-migmatitic terranes of "Série Ille de Cayenne", indicate ages in close agreement of 2.1 - 1.95 Ga, with coherent Sr initial ratios around 0.7018 - 0.7024, μ_1 value of 8.2 and positive ε_{Nd} values. These parameters suggest that the rocks were added to the crust during the Paleoproteorozic (Teixeira *et al.*, 1985; Milési *et al.*, 1995).

Detrital zircon U/Pb ages from metagreywacke and conglomerate yielded ages between 2.2 and 2.1 Ga (Gibbs, 1980; Milési *et al.*, 1995). The older ages were interpreted as the basement rock-forming ages. The associated bimodal volcanism gave a Sm/Nd whole-rock isochron of 2.1 Ga (Gruau *et al.*, 1985), which is in close agreement with the detrital zircon U/Pb ages around 2.1 Ga (Egal *et al.*, 1994).

The supracrustal sequences are intruded by different types of granite, with ages around 2.08 Ga (Milési *et al.*, 1995).

Based on the isotope data discussed above, Tassinari (1996) divided the MIP in two domains. The first domain consists of a "sialic" domain composed of reworked old Archean nuclei during the Transamazonian Orogeny as "inliers" within Paleoproterozoic crustal rocks and it is restricted to the southeastern part of the MIP (Pará and Amapá states in Brazil) and the northwestern part of the MIP (Imataca Complex in northeastern Venezuela). The second domain, which covers most of the MIP, corresponds to a "simatic or juvenile" domain developed by juvenile crustal accretion during the Paleoproterozoic era.

Most of the models have been the subject of controversy since the evolution of the MIP is based on geochronological and isotope results. For this reason it is important to review and discuss the results. Several syntheses of the radiometric results have been published (Gibbs and Barron, 1993; Teixeira *et al.*, 1989), but recent geochronological studies have furnished additional geochronological constrains on the evolution of the MIP (Sato and Tassinari, 1997; McReath and Faraco, 1997; Lafon *et al.*, 1998; Vanderhaeghe *et al.*, 1998). Even considering the new data that has appeared in the literature, it is necessary to emphasize the scarcity of reliable radiometric data in such a huge domain. Most of the data have been obtained by Rb/Sr and K/Ar methods, and with the exception of some restricted areas (Central Amapá region: Montalvão and Tassinari, 1984; Lima *et al.*, 1982; João and Marinho, 1982a; Bakhuis Mountains in Surinam: Gaudette *et al.*, 1976) all the Rb/Sr and K/Ar results gave a Paleoproterozoic age. This roughly constrains the Transamazonian evolution in the MIP between 2.1 - 1.8 Ga (Spooner *et al.*, 1971; Amaral, 1974; Lima *et al.*, 1974; Berrangé, 1977; Priem *et al.*, 1977, 1978, 1980; Gibbs and Olszewski, 1982; Bosma *et al.*, 1983; Montalvão and Tassinari, 1984; Teixeira *et al.*, 1984; Gaudette and Olszewski, 1985; Gruau *et al.*, 1985). These ages do not permit us to establish the geochronological succession of the different lithotectonic units in the MIP. Most of the available ages are widely dispersed within the 2.1 - 1.8 Ga range and they must been considered only as cooling ages (see for example the K/Ar results on minerals from the Vila Nova Group in Amapá; Hurley *et al.*, 1968). These dates mainly reflect the end of the Transamazonian Orogeny.

Recently, Sm/Nd (model and isochron ages), Pb/Pb and U/Pb zircon dating has permitted a solution to some problems that the Rb/Sr and K/Ar determinations have not solved and to better constrain the timing of the Transamazonian evolution of the MIP. Geochronological results showing Archean ages in the MIP are very scarce and mostly ambiguous. Only in the northeasternmost part of the MIP, U/Pb and Rb/Sr results on metamorphic and igneous rocks from the Imataca Complex are admitted as recording a complex Archean history (Montgomery, 1979; Montgomery and Hurley, 1978; Teixeira *et al.*, 1999). A Rb/Sr age of about 2.76 Ga on granulitic rocks from the Central Guiana Granulitic Belt in Suriname led Gaudette *et al.* (1976) to propose an Archean age for the protolith of the high-grade rocks but further Rb/Sr dating on whole- rock and U/Pb results on zircon (Priem *et al.*, 1978) do not indicate any involvement of an Archean crust in the metamorphic rocks even if this possibility has not to be definitively discarded (Priem *et al.*, 1978; Bosma *et al.*, 1983). In central Amapá (Cupixi - Tartarugal region), Rb/Sr ages of 2.45 and 2.9 Ga obtained on charno-enderbite and tonalitic orthogneiss, respectively, also suggest the existence of old Archean crustal domains in the MIP (João and Marinho, 1982a; Montalvão and Tassinari, 1984). Pb/Pb ages of about 2.55 - 2.49 Ga obtained on zircon from the Amapá Granulite (Lafon *et al.*, 1998) and Sm/Nd model ages of 3.1 - 2.94 Ga on the tonalite (Sato and Tassinari, 1997) strongly reinforce the involvement of an Archean crust in the southeastern part of the MIP. The existence of detrital zircon in quartzite associated with the Paramaca Formation in the S of French Guiana (Camopi region) suggests that the extension of the Archean crust reworked during the Transamazonian Orogeny in the MIP must be greater than previously believed. The range of ages on the detrital zircon (3.19 - 2.73 Ga) confirms that the segment of Archean crust is similar to that known in the Carajás Archean province (Lafon *et al.*, 1998) as previously suggested by Tassinari (1996).

The U/Pb and Pb/Pb on zircon ages and Sm/Nd, ranging between 2.26 and 2.11 Ga indicate an early Transamazonian age for the emplacement of the greenstone sequences (Gibbs and Olszewski, 1982; Gruau *et al.*, 1985; Gaudette *et al.*, 1996; McReath and Faraco, 1997; Norcross *et al.*, 1998; Vanderhaeghe *et al.*, 1998). The geochronological data are insufficient to establish if all the greenstone sequences are coeval at the scale of the MIP as well as the duration of the greenstone processes. In French Guiana, Pb/Pb and Sm/Nd ages; (Vanderhaeghe *et al.*, 1998) suggest a southward decreasing age for the greenstone sequences, but the large error (± 90 Ma) on the Sm/Nd age of the southern greenstone rocks (Gruau *et al.*, 1985) strongly limits such an interpretation. On a MIP scale, no evident relationships between geographical situation and age of the greenstone sequences may be pointed out. The greenstone belts occur in both juvenile and reworked domains of the MIP but the available Nd isotope data do not suggest any crustal contribution.

The orthogneiss and granitoid rocks widespread in the MIP have given ages Rb/Sr and K/Ar ages mostly between 2.1 and 1.9 Ga. In the juvenile domain, Pb and Sr isotopes (initial Sr: 0.702 - 0.704; $\mu 1 = 8.09$) show that this magmatism has a mantle-derived origin or comes from early Transamazonian crust involved in the orogeny (Teixeira *et al.*, 1985, 1989; Gruau *et al.*, 1985). This hypothesis is strongly reinforced by the existence of 2.22 Ga inherited lead in zircon from French Guiana granitoid plutons (Vanderhaeghe *et al.*, 1998). Granitoid rocks from the Archean reworked domain indicate high initial Sr values (0.712 < initial Sr < 0.760) related to the participation of an old crustal component in the generation of the granite (Santos *et al.*, 1988; Montalvão and Tassinari, 1984). Although high initial Sr has been obtained for high-K granite from central Amapá (Montalvão and Tassinari, 1984), the ε_{Nd} for the same rocks gave a positive value (+ 1.8), which exclude an origin through melting of old crustal material.

The most detailed chronology of the Transamazonian magmatic episodes has been obtained in the northern part of French Guiana (Vanderhaeghe *et al.*, 1998). These authors drawn attention to an episode of trondjhemitic magmatism at 2.17 Ga followed by the emplacement of calc-alkaline intrusions at 2.144 to 2.115 Ga and a late high-K magmatism at 2.09 to 2.08 Ga. The main units of the MIP can be divided into high-grade metamorphic complexes, including granulite and amphibolite gneiss, greenstone terranes and other supracrustal units and granitoid and related magmatic rocks (Teixeira *et al.*, 1989).

High Grade Metamorphic Terranes

Granulite

The main occurrences of granulite in the MIP are found in the Imataca Complex (Venezuela), Central Guyana Granulitic Belt (Suriname, Guyana and northern Roraima) and in the Tumucumaque Granulite Belt in Amapá, Brazil (Kroonenberg, 1976; Lima *et al.*, 1982; Gibbs and Barron, 1983).

The Imataca Terrane

This complex belongs to a 450 km long and 100 km wide belt of high-grade rocks in northernmost Venezuela, limited to the S by the Guri Fault. These rocks are mainly felsic granulite with some intermediate and mafic granulite. They are considered to be of igneous origin and derived from Archean protoliths (3.7 - 3.4 Ga; Montgomery and Hurley, 1978; Montgomery, 1979; and 3.23 - 2.93 Ga and 2.82 - 2.60 Ga; Teixeira *et al.*, 1999). Metasediments are also present. The age of granulitic metamorphism is related to the Transamazonian event between 2.2 - 2.0 Ga (Montgomery and Hurley, 1978), when the Imataca allocthonous block had been juxtaposed to the Maroni - Itacaiúnas Belt.

The Central Guyana Granulitic Belt

The Central Guyana Granulitic Belt is a WSW-ENE elongated belt stretching over at least 1000 km from Roraima (Apiú Complex) to southern Guyana (Kanuku Complex) and Suriname (Falawatra Complex in the Bakhuis Mountains). Most of the rocks are felsic granulite (enderbite dominant and charnockite), with minor amounts of basic granulite and sillimanite gneiss. Two phases of metamorphism have been identified. Whether these phases represent independent metamorphism events or a prograde and a retrograde stage of the same event is yet an outstanding problem.

All the attempts to date the belt gave Transamazonian ages, even if an Archean age has been suggested for the protoliths of the granulite (Gaudette *et al.*, 1976). For the Bakhuis Granulite, a U/Pb zircon age of 2.026 ± 0.02 Ga (Priem *et al.*, 1978) and Sm/Nd model age of 2.3 Ga (Ben Othman *et al.*, 1984) were obtained. The Kanuku Granulite in Guyana gave ages of 2.05 Ga (Rb/Sr, Spooner *et al.*, 1971) and 2.2 Ga (Sm/Nd model age, Ben Othman *et al.*, 1984). In the State of Roraima, zircon U/Pb, Rb/Sr and Sm/Nd ages resulting from analysis of the Kanuku Charnockite are scattered between 2.02 and 1.85 Ga (Lima *et al.*, 1986; Gaudette *et al.*, 1996). Nevertheless, the set of radiometric data, together with the Pb/Pb age of 1.966 ± 0.037 Ga on zircon from a charnockite intrusion (Fraga *et al.*, 1997), strongly suggest that the high-grade rocks are related to a late phase of the Transamazonian Orogeny (*i.e.*, 2.0 Ga or younger). Fraga *et al.* (1997) discussed the nature of this high-grade episode, and suggested that it may be related to charnockitic magmatism. These authors also demonstrated the existence of 1.56 Ga charnockite bodies in the same area, very close to the boundary with the Ventuari-Tapajós Province, associated with 1.54 Ga rapakivi-type granite (Gaudette *et al.*, 1996), reflecting an overprinting of a Mesoproterozoic thermal event related to younger orogenies.

The Tumucumaque Granulitic Belt

The Tumucumaque Granulitic Belt is found in the southeasternmost part of the MIP, mainly in Amapá, with an apparent NW-SE trend. The main occurrence is the Tartarugal Grande Metamorphic Suite in central Amapá. As in others areas, most of the rocks are felsic gneiss and granulite (charnockite, enderbite and hyperstene-free gneiss). Mafic rocks with minor occurrences of quartzite and kinzigite have also been described (João and Marinho, 1982a). These mafic-sedimentary rock associations have been considered as the high-grade equivalents of the greenstone terranes (Vila Nova Group) of Amapá. Rb/Sr dating of charnockite and enderbite indicate an age of 2.45 ± 0.074 Ga (João and Marinho, 1982a) or 2.674 Ga (Montalvão and Tassinari, 1984) depending on the samples included in the isochron calculation. Zircon crystals from a garnet granulite have furnished Pb/Pb ages in the range of 2.58 - 2.49 Ga (Lafon *et al.*, 1998). These ages clearly indicate the existence of Archean protoliths in the Tumucumaque Belt, but do not constrain the age of the granulitic episode.

Gneiss and Migmatite

Other high-grade rocks including amphibolite gneiss and migmatite have been described in the MIP. They have been generally associated with the granulitic rocks even if their relationships are not well defined. In Venezuela, granitic gneiss, amphibolite and migmatite occur in the Imataca Complex. Petrological studies indicate that they represent felsic rocks metamorphosed in the amphibolite facies. A large migmatitic body (La Ceiba Migmatite) has given a Rb/Sr age of about 2.7 Ga for the migmatization.

The Supamo Complex also includes quartz-feldspathic gneiss and migmatite. In the southeastern part of Venezuela, U/Pb and Rb/Sr dating of granitic and quartz-diorite augen gneiss furnished a range of 1.86 to 1.76 Ga for the amphibolite facies metamorphism (Gaudette and Olszewski, 1985). In Guyana, the eastern part of the Kanuku Complex consists mainly of rocks metamorphosed in the amphibolite facies, whereas granulite is dominant in the western part (Berrangé, 1977). The Bartica granitoid and gneiss, may be considered as the equivalent in Guyana of the Supamo Gneiss, and represent an early phase of the Transamazonian magmatism at about 2.25 Ga (Gibbs and Olszewski, 1982).

In southwestern Suriname, amphibolite and sillimanite gneiss are associated with granulite in the Coeroeni Group. These rocks are considered to be mainly of sedimentary origin and have a metamorphic history similar to the Falawatra Group, but were formed at lower pressures (Bosma *et al.*, 1983; De Vletter *et al.*, 1998). Quartz feldspathic gneiss is also described in northeastern Suriname (De Vletter *et al.*, 1998). Geochemical and petrological characteristics suggest that they could represent metamorphosed early Transamazonian granitoid plutons. Amphibolite grade gneiss, mostly of igneous origin, is also present in the northeastern part of the State of Roraima, in association with granulite (Kanuku Complex). The relationships with the granulite are not well defined. Gaudette *et al.* (1996) considered these as syn-tectonic intrusives in the granulite, whereas Fraga *et al.* (1997) suggested that granulite was emplaced within the surrounding orthogneiss. U/Pb ages between 1.94 and 1.88 Ga have been obtained for the protholith emplacement of the orthogneiss which led Gaudette *et al.* (1996) to exclude them from the MIP.

Orthogneiss and migmatite are widespread in French Guiana (Choubert, 1974; Gruau *et al.*, 1985; Marot, 1988, Vanderhaeghe *et al.*, 1998). Metamorphic conditions are generally in the amphibolite facies with local anatexis. In the northern part of French Guiana, the Ille de Cayenne and Central Guyana complexes are considered as high-grade equivalents of Paramaca Volcanics and metamorphic conditions are related to regional contact metamorphism with numerous calc-alkaline plutonic intrusions. (Vanderhaeghe *et al.*, 1998). In the State of Amapá, migmatite also occurs in the central part of the state and is related to high-K magmatism at about 2.06 Ga (Montalvão and Tassinari, 1984). In the Altamira region (Guaporé Shield) amphibolite facies metamorphism, migmatization, and anatexis also affects most of the basement rocks at about 1.99 to 1.93 Ga (Santos *et al.*, 1988).

Greenstone terranes

The Maroni-Itacaiúnas Province is mainly underlain by metavolcano-sedimentary sequences, metamorphosed in the greenschist to amphibolite facies, which are associated with Paleoproteorozic granite-greenstone terranes (Gibbs, 1980). Low-grade to medium-grade volcano-sedimentary rock associations are distributed throughout the MIP, with their major extension in the northwestern part of the Province (Venezuela/Guyana). They show similarities to Archean greenstones of which they are considered as the Proterozoic equivalent (Choudouri, 1980; Gibbs, 1980; Gibbs and Barron, 1993). In like manner to typical Archean greenstone belts, they consist of lower sequences of mainly metavolcanic rocks (ultramafic rocks and basalt with subordinate andesite) and upper sequences consisting mainly of felsic volcanics and metasedimentary rocks. The thickness of the whole sequence may attain 10 km. Gibbs and Barron (1993) have reviewed distribution, rock-types, stratigraphy and correlation between the different belts in each country in detail, but these authors do not speak about the associated granite intrusives, which characterise the granite greenstone associations. Bosma *et al.* (1983) in Suriname, and Gruau *et al.* (1985) and Vanderhaeghe *et al.* (1998) in French Guiana discussed the relationships between granitoid plutons and greenstones sequences. We will comment on this point in the chapter on the granitogenesis in the MIP.

Venezuelan greenstone belts are roughly oriented N-S. They have been included in the Pastora Group that consists of two principal belts: the Paragua-Caroni Belt (western belt) and the Guasipati Belt (central belt). No geochronological data are available for the Pastora Group, but a Paleoproterozoic age has been generally accepted.

In Guyana, three greenstone belts trending NW-SE have been described (Gibbs, 1980), and have been included in the Barama - Mazaruni Supergroup (from N to S: Barama Belt, Cuyuini Belt and Mazaruni Belt). In the Barama Belt some komatiitic rocks have been mentioned (Gibbs and Barron, 1993). An U/Pb age of 2.245 ± 0.086 Ga has been obtained on zircon from metagreywacke of the Cuyini and Mazaruni belts, which support a Paleoproterozoic age for the formation of the greenstones (Gibbs and Olszewski, 1982).

The greenstone belts in Suriname are grouped in the Marowijne Group, and occur in the northern and eastern part of the Suriname. Bosma *et al.* (1983) distinguished a lower Paramaka Formation (mostly mafic volcanics with subordinate intermediate volcanic rocks) from an upper Armina Formation consisting mainly of metagreywacke layers. Metamorphism of the Marowijne Group took place in the greenschist facies, but amphibolite facies rocks are locally described in the proximity of diapiric intrusions of tonalite and trondjhemite. Based on geochemical criteria, Bosma *et al.* (1983) suggested an island arc - back-arc marginal basin for the depositional environment of the Paramaka Volcanics. A Rb/Sr dating on metavolcanic rocks from the Paramaka Formation gave an age of 1.95 ± 0.15 Ga (Priem *et al.*, 1980)

In French Guiana, Choubert (1974), Gibbs and Barron (1983), Gruau *et al.* (1985) and Marot (1988) have described greenstone belts with some modifications in the nomenclature, but all of them are included in the Paramaca Group. They occur mostly in the central and northern parts of French Guiana. Recent work (Egal *et al.*, 1995; Vanderhaeghe *et al.*, 1998) has focused on the greenstones from the northern part of the country (Cayenne - Régina region). The greenstones appear as two belts, which underlie the eastern part of French Guiana and Suriname. The southern belt consists of metamorphosed ultramafic and mafic units (including komatiite) with intermediate to felsic volcanics, followed by the deposition of mainly sedimentary rocks (Choubert 1974; Gruau *et al.*, 1985).

In the northern part of French Guiana, Vanderhaeghe *et al.* (1998) described the northern greenstone belt as two main complexes separated by a sedimentary basin (Orapu Basin). These complexes (Isle de Cayenne and Central Guyana complexes) include metavolcanic and volcaniclastic rocks of the Paramaca Formation (mainly dacitic and andesitic rocks with subordinate basaltic and rhyolitic rocks) that have undergone metamorphism from the greenschist to amphibolite facies. These sequences together with the sediments of the Armina Formation constitute the Lower Volcanic and Sedimentary Unit. Gruau *et al.* (1985) provided a Sm/Nd age of 2.11 ± 0.09 Ga for the metavolcanics for the formation of the greenstones of the southern belt. In the northern greenstone belt, the age of formation of the greenstones is indirectly constrained by the 2.216 ± 0.004 Ga, 2.174 ± 0.007 Ga and 2.144 ± 0.006 Ga on associated trondjhemite and calc-alkaline plutons (Vanderhaeghe *et al.*, 1998).

In northern Brazil greenstone sequences occur as sub-parallel belts defining trends roughly oriented in a NW-SE in Amapá and northeastern Pará. They have been grouped in the Vila Nova Group (Lima *et al.*, 1974) and their rock-types mainly consist of mafic and ultramafic schist at the base to dominantly metasedimentary formations at the top (mostly quartzite, BIF and metaconglomerate). The high Mg character of the mafic-ultramafic rocks suggests a komatiitic affinity (João *et al.*, 1979). The stratigraphic position and nomenclature have been controversial (Lima *et al.*, 1982; João and Marinho 1982a). This applies especially to the central part of Amapá because of the uncertain relationships with neighbouring high-grade rocks and ambiguous geochronological results in the country rocks, which led some authors to consider them as Archean (João and Marinho, 1982a). Rb/Sr results on whole-rock and K/Ar on mica and amphibole from amphibolite of the Serra do Navio greenstone belt range between 2.09 and 1.76 Ga (Hurley *et al.*, 1968; Basei 1977; Montalvão and Tassinari, 1984). Recently, McReath and Faraco (1997) reported a Sm/Nd age of 2.264 ± 0.034 Ga for the Vila Nova Group in northeastern Pará (Ipitanga Greenstone Belt) in close agreement with the previous Sm/Nd and U/Pb ages available for the others greenstones in the MIP (Gibbs and Olszewski, 1982; Gruau *et al.*, 1985).

Other supracrustal formations

Other supracrustal formations may be distinguished from the greenstone belts mainly because of the lack of basic volcanic rocks and the predominance of metasedimentary sequences (Gibbs and Barron, 1993). Relationships with the greenstone belts are generally controversial. For example, in Roraima, the NW to WNW trending Parima - Cauarane belts have been considered as greenstone belt sequences and dated at 2.235 ± 0.019 Ga (Gaudette *et al.*, 1996), whereas Gibbs and Barron (1993) described them as separate units. In southwestern Suriname, quartz-feldspathic and pelitic gneiss with amphibolite, quartzite and calc-silicate rock mainly constitute the Coeroeni Group (Kroonenberg, 1976). Bosma *et al.* (1983) considered the Coeroeni Group as an intracratonic basin similar to the present-day Amazonas River Basin. The age of the Coeroeni Group is poorly constrained by a Rb/Sr age of 2.042 ± 0.097 Ga (Priem *et al.*, 1977). In French Guiana, Ledru *et al.* (1991, 1994) have discussed the relationships between the Paramaca Formation and Orapu and Bonidoro series and resolved discrepancies on correlation with the Suriname formations (Armina and Rosebel formations). They describe a Lower Volcanic Sedimentary Unit and an Upper Sedimentary Unit which have been formed, respectively, in oceanic (volcanic rocks, tholeiitic and with calc-alkaline affinities) and intra-continental (fluvio-deltaic deposits) environments. The former stage corresponds to a pre 2.1 Ga magmatic accretion episode, whereas the latter stage is related to a post 2.1 Ga crustal recycling episode (Vanderhaeghe *et al.*, 1998).

Granitoid plutons

The most representative rocks in the whole MIP are granitoid and orthogneiss. Excluding the orthogneissic rocks from the Imataca Complex, and tonalitic orthogneiss from the central-southern part of Amapá in Brazil (Cupixi region), all the granitoid plutons present paleoproterozoic ages, related to the Transamazonian Orogeny. Distinctions between different types of granitoid have been proposed according to geographical occurrence, petrological and structural features, relationships with supracrustal sequences and, more recently, on geochronological criteria. Even if some local peculiarities between different countries can be pointed out, two main phases of granitogenesis are normally accepted in the MIP (Teixeira *et al.*, 1989; Gibbs and Barron, 1993).

The first granitic episode consists of syn-tectonic intrusions and orthogneiss, mostly of tonalitic, trondjhemitic and granodioritic composition and with high Na content (Supamo Granitoid in Venezuela, Bartica Orthogneiss in Guyana, "Guianais" Granite in French Guiana). These granite bodies are coeval with the main deformational event and are associated with the greenstone belts. The second granitic phase is represented by late to post-tectonic granite intrusives with K or calc-alkaline affinities ("Caraïbes" Granite in French Guiana). In Suriname, Bosma *et al.* (1983) described the Transamazonian granite intrusives in the domain of the MIP. Granitoid plutons associated directly with the greenstone sequences consist of mainly tonalitic to trondjhemitic diapiric intrusions within the Paramaca Formation, whereas the two-mica leucogranite results from anatexis of crustal rocks within the Armina Formation. This distribution roughly corresponds to the different granite-types described in French Guiana. Most of the available ages on the Transamazonian granites have been obtained by Rb/Sr and K/Ar methods and gave a range of ages between 2.3 - 1.8 Ga (Teixeira *et al.*, 1989; Gibbs and Barron 1993; Tassinari, 1996). In the northern part of French Guiana, Vanderhaeghe *et al.* (1998) refer to the existence of granitoid of calc-alkaline affinity (tonalite and granodiorite) associated with continental crust accretion (2.144 - 2.115 Ga) and late crustal derived K-rich granite and peraluminuous granite related to crustal recycling (2.093 - 2.083 Ga). U/Pb ages in the same range have been obtained by Norcross *et al.* (1998) for a post-tectonic intrusion in northwestern French Guiana. In northern Brazil (Amapá), the distinction between different phases of granite intrusion are difficult. In the central part of Amapá, K-rich granite has been dated at 2.06 Ga (Montalvão and Tassinari, 1984). Early Na-rich and late K-rich granite bodies have also been described in central Amapá (João and Marinho, 1982b)

Post Transamazonian units

The thermal evolution of the Transamazonian Orogeny is shown by the Rb/Sr and K/Ar systems of minerals indicating that they decreased from 500 to 300°C between 2.08 and 1.76 Ga (Montalvão and Tassinari, 1984; Tassinari, 1996). After the end of the Transamazonian Orogeny, the Maroni-Itacaiúnas Province behaved as a cratonized area, and no widespread magmatic-metamorphic event has yet been described. Locally, some magmatic activity has been observed. In Amapá, some felsic intrusions (Falsino Suite) and alkaline intrusions (Mapari Suite) have been dated at about 1.76 Ga (Tassinari *et al.*, 1984) and 1.68 - 1.34 Ga (Montalvão and Tassinari, 1984), respectively. In central Suriname, pyroclastic volcanic rocks interbedded in weakly metamorphosed continental sediments showed a 1.66 Ga Rb/Sr age (Priem *et al.*, 1973). Dolerite dykes of the same age or slightly older are also found in Suriname (Hebeda *et al.*, 1973; Norcross *et al.*, 1998) and in Guyana (Snelling and McConnell, 1969). In western Suriname, a 1.2 ± 0.1 Ga widespread mylonitization event (Nickerie Metamorphic Episode) has been determined by Rb/Sr and K/Ar analyses on mica. Locally, mica from rocks of southern Guyana also shows the same event (Snelling and McConnell, 1969). Permo-Triassic dolerite dyke swarm emplacement related to the opening of the Atlantic Ocean is well constrained between 221 Ma and 195 Ma in French Guiana and Suriname (Priem *et al.*, 1968; Deckart *et al.*, 1997).

Ventuari - Tapajós Province (VTP)

The western part of Amazonian Craton was assembled through a succession of magmatic arcs from 2.0 to 1.40 Ga involving the Ventuari-Tapajós, Rio Negro-Juruena and part of Rondonian-San Ignacio provinces. In this way the VTP corresponds to the first magmatic arc, which was accreted to the protocraton consisting of the Central Amazonian and Maroni-Itacaiúnas provinces. The VTP occurs in a prominent NW-SE trend and can be followed from the Ventuari River in southern Venezuela to the Tapajós River in Brazil (Fig. 1). The VTP, as a whole, exhibits a geochronological pattern slightly younger than the Maroni Itacaiúnas Province, with ages ranging from 2.0 Ga to 1.8 Ga (Table 3). The rock-types of the VTP, although not thoroughly documented, differ substantially from those of the MIP. The VTP consists mainly of calc-alkaline granitoid, whereas the MIP includes a large amount of metavolcano-sedimentary sequences and granulitic rocks among others.

Metamorphic Basement

The northern part of VTP is underlain by granite-gneiss having a granodioritic to quartz-dioritic composition; gabbro and amphibolite, which occur within the Ventuari petrotectonic domain (Barrios, 1983). U/Pb zircon ages between 1.85 and 1.83 Ga are available for the Macabana and Minicia gneiss and the Atabapo quartz-diorite, which are in close agreement with Rb/Sr age determinations of 1.83 Ga and Sr initial ratios around 0.7027 (Gaudette and Olszewski, 1981; Tassinari *et al.*, 1996).

The basement of the southern part of VTP, included in the Cuiú-Cuiú Complex (Pessoa *et al.*, 1977) and within the slightly younger Parauari Suite, has almost the same lithological assemblage described above, predominantly with calc-alkaline granodioritic rocks and gneiss of tonalitic composition, metamorphosed in the amphibolite facies. The zircon U/Pb determinations and whole-rock Rb/Sr ages vary from 2.0 to 1.85 Ga (Vignol, 1987; Gaudette *et al.*, 1996; Tassinari, 1996; Iwanuch, 1999). In general, the southern part of the province is older than the northern region, which suggest magmatic arc evolution and the production of new continental crust within the VTP. This began in the S and became progressively younger toward the N. The Sm/Nd crust-formation ages of 2.1 to 2.0 Ga (Sato and Tassinari, 1997) are found throughout the province, indicating the main period of continental accretion in the VTP. $\varepsilon_{Nd(2.0\ Ga)}$ values range from + 2.1 to - 1.6 (Sato and Tassinari, 1997), which suggest a mixing of some subordinate crustal component with predominantly mantle-derived magma. This mixture is due to the fact that external zones of VTP were developed on or adjacent to the Archean Central Amazonian Province.

The southern part of the VTP differs from the northern in having scattered occurrences of greenschist facies metavolcano-sedimentary sequences, which are included in the supracrustal Jacareacanga Suite (Santos *et al.*, 1997). They consist of schist, phyllite, metachert, banded iron formation units, and talc schist. The unit shows a U/Pb detrital zircon age and U/Pb zircon age of 2.1 to 1.92 Ga for the Parauari intrusive granitoid (Santos *et al.*, 1997).

Sedimentary Platform Covers and Associated Magmatism

The southern part of the VTP contains the alkaline to calc-alkaline Iriri acid to intermediate volcanic rocks and associated Maloquinha Granitoid with ages between 1.89 and 1.84 Ga (Santos *et al.*, 1997; Vasquez *et al.*, 1999; Moura *et al.*, 1999), and the acid to intermediate Teles Pires volcano-plutonism dated at *c.* 1.7 - 1.6 Ga (Basei, 1977; Silva *et al.*, 1980; Tassinari, 1996). The age of the volcanism indicates the time that the basins began to subside. These unmetamorphosed volcano-plutonic rocks mainly include rhyolite, rhyodacite, tuff, volcanic conglomerate, quartz-porphyry and dacite. Anorogenic ring granitic bodies with subvolcanic characteristics occur associated with the acid volcanism. In general these rocks are adamelite, granite porphyry, granite, riebeckite-granite, and rapakivi granite, locally with cataclastic textures. Geochemical characteristics suggest an alkaline, calc-alkaline and peralkaline composition for the granitoid (Silva *et al.*, 1980).

Sedimentary platform cover of the Gorotire and Beneficente groups, probably associated with foreland basins and dated at 1.8 - 1.6 Ga and 1.7 - 1.4 Ga, respectively (Santos *et al.*, 1997; Tassinari *et al.*, 1978), overlies these undeformed volcanic rocks that are intruded by 1.6 to 1.3 Ga mafic dykes. These sedimentary sequences, mainly composed of quartz-arenite, argillite, arkose, siltite, chert and carbonate rock such as limestone and dolomite, are typical of shallow marine deposits. The Beneficente Group is tectonically limited by NE-SW trending faults, and it was deposited across the southwestern margin of the Ventuari-Tapajós Province. The sedimentary deposition is coeval with tectonic evolution of the Rio Negro-Juruena Orogeny, which took place along the western side of the VTP between 1.8 and 1.55 Ga.

Rapakivi Granite and Associated Magmatism

The 1.55 Ga El Parguaza and Surucucus anorogenic, intra-plate, rapakivi granite bodies (Tassinari, 1996; Santos *et al.*, 1999) occur in the northern part of the VTP. Likewise, within the transition zone between the VTP and MIP there occur the 1.55 Ga rapakivi granite of the Mucajaí Suite and the associated Repartimento Anorthosite and the Serra da Prata

Charnockite that form part of a Charnockitic-Anorthositic-Rapakivi association, related to the Parguaza event (Fraga and Reis, 1995; Fraga *et al.*, 1997; Santos *et al.*, 1999).

The important anorogenic rapakivi magmatism occurring to the N of the VTP and dated at around 1.55 Ga may be a product of fractionation of magma from the base of the crust, involving material produced by partial melting of upper mantle and the lower-middle part of the continental crust. Although this magmatism occurred within a tectonicaly stable area of the VTP, it is related to orogenic processes of the neighbouring Rio Negro-Juruena Province due to the fact that convergent processes produced chemical changes in the upper mantle that may have led to crustal melting and rapakivi magmatism (Haapala and Rämo, 1995). The large volumes of rapakivi magmatic rock in the area show the importance of the continental accretion in the northwestern part of the Amazonian Craton, at the beginning of the Mesoproterozoic time.

Western Amazonian Craton

The western part of the Amazonian Craton, in like manner to the Grenville Province of Canada, is a multi-orogen region formed between 1.8 and 1.0 Ga where successive magmatism, metamorphism and deformation occurred that regionally affected and reworked precursor provinces, producing new complexes, as well as new juvenile continental crust.

Three major geochronological and tectonic provinces (*sensu* Cordani *et al.*, 1979; Teixeira *et al.*, 1989; Tassinari and Macambira, 1999) occur in the western part of the craton. These are the Rio Negro-Juruena Province (1.8 - 1.55 Ga), the Rondonian/San Ignacio Province (1.55 - 1.3 Ga), and the Sunsas Province (1.3 - 1.0 Ga). Recent advances in the understanding of the evolution of these provinces, based on new geochronological and geological data (Tassinari *et al.*, 1996; Tassinari, 1996; Sato and Tassinari, 1997; Sato, 1998; Van Schmus *et al.*, 1998; Geraldes *et al.*, 1999; Geraldes, 2000; Rizzoto, 1999; Payolla *et al.*, 1998; Pinho *et al.*, 1997; Bettencourt *et al.*, 1997, 1999) amongst others, provide the basis upon which it is possible to subdivided the provinces by orogenies and terranes. The orogenies are defined in terms of their foreland thrust-fold belts, Andean-type magmatic arcs, basement reactivation, transcurrent shearing, thrusting and deformed foredeep sedimentary prisms.

There remain many unanswered questions about the crustal evolution of the southwestern cratonic area, and a major practical obstacle in deciphering the complex evolution is the lack of a comprehensive nomenclature and consensus for describing various geochronological tectonic provinces, tectonic zones, mobile belts, orogens, terranes, domains, blocks, inferred arcs, detailed geological mapping, and precise geochronological data. As a result, conflicting interpretations have been made concerning the boundaries, distribution, evolution, and the real meaning of these provinces. These facts make it critical to unify and separately analyze the different orogens, which might allow temporal correlation of units and provide better constraints for plate tectonic reconstruction. Available information indicates that a variety of different geological environments are represented within the western Amazonian Craton, and consequently some terranes and orogenies can be separated from whole geochronological provinces. With this in mind we now propose the following major preliminary components and their subdivisions, showed in Figure 3:

- Rio Negro - Juruena Province (1.8 - 1.5 Ga)
  - Alto Jauru Greenstone Belt (1.79 - 1.75 Ga)
  - Cachoeirinha Orogen (1.58 - 1.52 Ga)
- Rondonian - San Ignacio Province (1.5 - 1.29 Ga)
  - Rio Alegre Terrane (1.5 Ga)
  - Santa Helena Orogen (1.47 - 1.42 Ga)
  - Rondonian-San Ignacio Orogen (1.4 - 1.29 Ga)
- Sunsas Province (1.3 Ga - 980 Ma)
  - Sunsas Cycle, in Bolivia (1.25 Ga - 900 Ma)
  - Aguapeí Thrust Belt (1.0 Ga - 950 Ma)
  - Nova Brasilândia Terrane (1.11 - 1.0 Ga).
  - Sunsas Orogenesis in Central-Northern Rondônia and Adjacent Areas (1.08 Ga - 980 Ma)

The Proterozoic basement in the western part of the Amazonian Craton, and in particular in the southwestern part, consists of igneous and metamorphic associations. The Rio Negro-Juruena Province basement rocks include several domains of distinctly different rock-types, including several volcano-sedimentary belts, felsic plutonic-gneiss and intrusive granitoid. Two main accretionary events are described for the Rio Negro-Juruena Province. According to Tassinari *et al.* (1996); Van Schmus *et al.* (1999); Geraldes *et al.* (1999) the first comprises the Jamari and Roosevelt terranes of Scandolara *et al.* (1999c) in northern Rondônia; the Alto Jauru greenstone belt of Pinho *et al.* (1997) and Geraldes (2000) in Mato Grosso. These events were defined by U/Pb ages from 1.79 to 1.717 Ga for gneiss and volcanic rocks. The second event, named the Cachoeirinha Orogen (Geraldes *et al.*, 1999) is defined in the same geographical area as the first event and consists of tonalite, granodiorite and granite with ages of c. 1.57 - 1.52 Ga (Tassinari *et al.*, 1996; Geraldes, 2000).

Rondonian-San Ignacio events have traditionally been regarded as extending from 1.5 to 1.29 Ga, with a wide geochronological distribution (Fig. 3). New geochronological data are now demonstrating that orogenic events such as the proposed Rio Alegre Terrane (1.5 - 1.49 Ga), Santa Helena Orogeny (1.47 - 1.42 Ga) and Rondonian-San Ignacio Orogeny (1.4 - 1.29 Ga) may be included in the Rondonian-San Ignacio Province.

Moreover it is postulated that the events previously defined between 1.6 Ga - 970 Ma through Rb/Sr isochron ages, within the southwestern part of the craton, need redefinition since Bettencourt *et al.* (1999a) argued that U/Pb zircon ages for the rapakivi suites of the Rondônia area are consistently older than the Rb/Sr isochron ages for the same plutons. The differences being 200 and 150 Ma for plutons older than 1.5 Ga, 100 Ma for those between 1.41 and 1.3 Ga and 4 to 30 Ma for the younger granite with ages situated between 1.08 Ga and 970 Ma. Similar situations were partially recorded in the southwestern Mato Grosso by Van Schmus *et al.* (1998) and Geraldes (2000).

The Sunsas Province is formed by two terranes already described (Sunsas in Bolivia, and Aguapeí in Brazil), the recently-defined Nova Brasilândia Terrane (Rizzoto, 1999a,

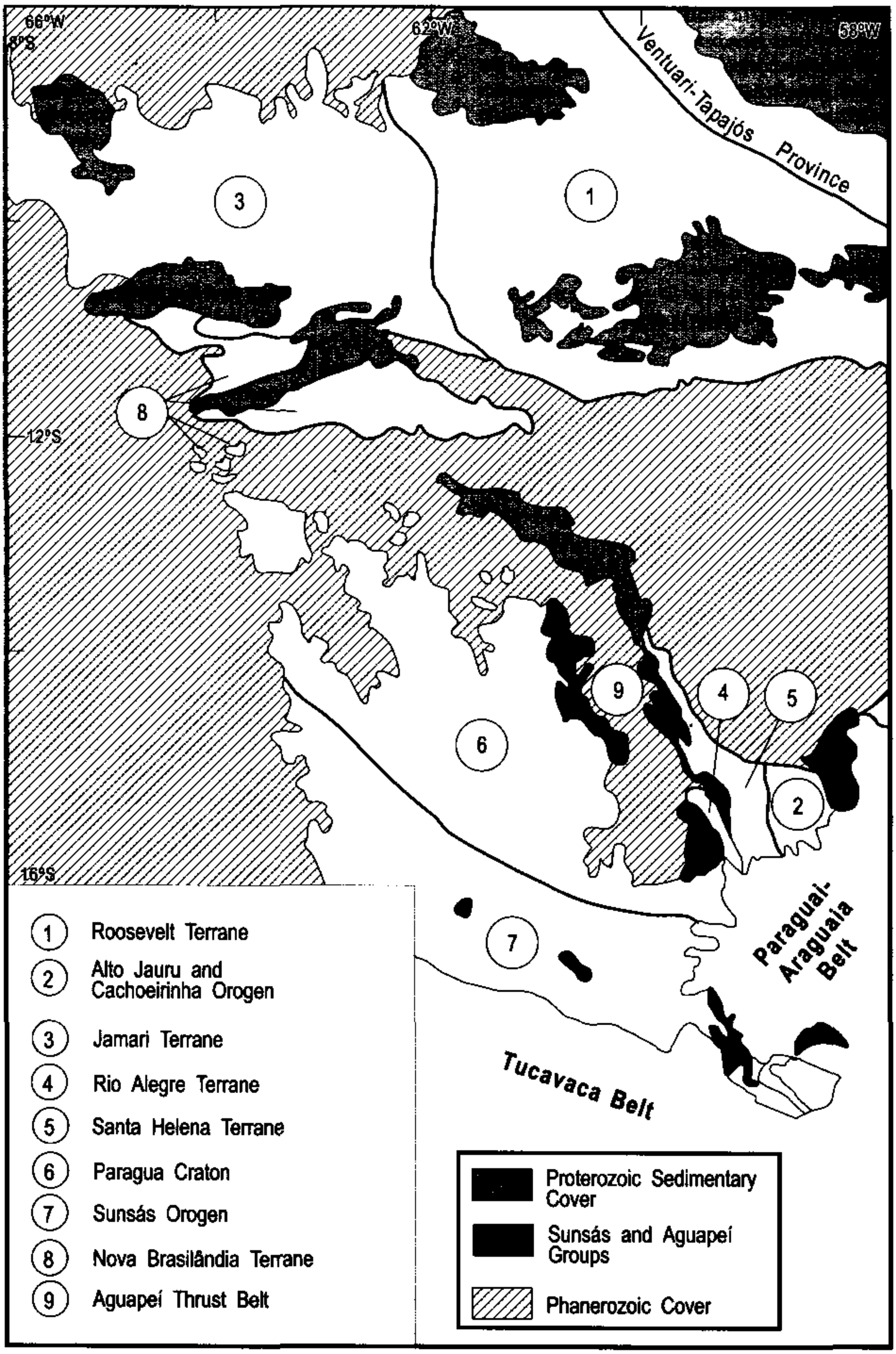


FIGURE 3 - Tectonic map of SW Amazonian Craton showing provinces and orogens.

b), and the redefined Sunsas event in central-northern Rondônia. This review, based mainly on the recent U/Pb, Sm/Nd geochronological data, chemical studies and less detailed geological surveys, is a preliminary attempt to model the Proterozoic evolution of this sector of the Amazonian Craton. It also aims to stimulate further research into the temporal correlation between the tectonic events in the southwestern part of the Amazonian Craton and those of the Laurentian and Baltic shields, thereby providing constraints for plate tectonic reconstruction.

Rio Negro - Juruena Province (RNJP)

The Rio Negro-Juruena Province was defined by Cordani *et al.* (1979) and Tassinari (1981), based on Rb/Sr and K/Ar ages between 1.75 and 1.5 Ga, as a geochronological province consisting of a juvenile mobile belt. Furthermore, Tassinari (1984), Tassinari *et al.* (1996) and Tassinari and Macambira (1999), based on other geochronological techniques, such as Pb/Pb, U/Pb and Sm/Nd, considered the Mesoproterozoic basement of the province as developed by successive magmatic arcs during 1.8 to 1.55 Ga. The basement includes several domains of different rock-types, including volcano-sedimentary belts, felsic plutonic-gneiss and granitoid intrusions. Extensive continental platform molasse or marine sedimentary rocks of Mesoproterozoic, Paleozoic and Mesozoic age fill rift basins within the province.

General Geological Aspects

The Rio Negro - Juruena Province is situated on the western side of the VTP, and its exposures straddle a NW-SE trend approximately 2000 km long and 600 km wide (Fig. 1) in the western part of the Amazonian Craton. The northern region (Uaupés) covers part of Brazil, Venezuela and Colombia and the southern region (only in Brazil) includes the states of Amazonianas, Rondônia and Mato Grosso. Juvenile Paleoproterozoic and Mesoproterozoic crust accreted to the western margin of the Ventuari-Tapajós Province during at least two orogenic events (1.8 - 1.7 Ga and 1.65 - 1.55 Ga) is exposed in the Rio Negro-Juruena Geochronological Province.

Metamorphic Terranes (Basement Rocks)

The basement rocks of the RNJP are composed almost entirely of 1.8 - 1.55 Ga granite-gneiss and granitoid of mainly granodioritic and tonalic composition. Generally the rocks are metamorphosed to the amphibolite facies, although some granulite facies rocks are also present. In the N there mainly occurs biotite-titanite monzogranite (Dall'Agnol and Macambira, 1992), and part of the basement complex extends into Colombia, where it is included in the Mitu Complex (Galvis *et al.*, 1979) and into Venezuela, where they form part of the Casiquiare Petrotectonic Association (Barrios, 1983). On the other hand, in the southern part, granite-migmatite terranes and gneiss of tonalitic composition compose the basement rocks. The general strike of the metamorphic rocks is NW-SE that is crosscut in some areas by NE-SW trending shear zones (Lima *et al.*, 1986).

The geochronological pattern available for the Rio Negro-Juruena Province basement (Table 4) is very complete and coherent, involving Rb/Sr, Pb/Pb and U/Pb (including SHRIMP) ages within 1.8 and 1.7 Ga and 1.65 - 1.5 Ga intervals. Sm/Nd mantle-depleted model ages ranging from 2.0 to 1.7 Ga, which suggest that the protholiths of the basement rocks are composed by mantle-derived magma and also by a mixing between juvenile material and magma produced by recycling of the earliest Ventuari-Tapajós crust. According to Tassinari *et al.* (1996) the K/Ar cooling ages on biotite from the granite-gneissic rocks yielded ages within 1.4 - 1.1 Ga interval. In this way, the older K/Ar ages are interpreted as referring to the uplift and tectonic stabilization of the area, and the younger K/Ar ages may be related to the reheating produced by anorogenic magmatism and a thermal-shearing Nickerie event of Priem *et al.* (1973).

The biotite-titanite monzogranite of the Rio Negro area (N of the RNJP) yielded whole-rock Rb/Sr and Pb/Pb isochron ages of 1.7 Ga and 1.63 Ga, respectively (Tassinari, 1984), and zircon U/Pb ages of 1.7 and 1.52 Ga (Tassinari *et al.*, 1996). Gneiss of tonalitic composition from the Casiquiare and Siapa domains in Venezuela gave a whole-rock Rb/Sr isochron age between 1.78 and 1.62 Ga (Barrios, 1983; Gaudette and Olzewski, 1981). In the southern part of the province, gneiss and granitoid rocks from the Porto Velho-Juruena area yielded SHRIMP U/Pb zircon ages of 1.75 and 1.57 Ga and whole-rock Rb/Sr and Pb/Pb isochron ages of 1.7 Ga (Tassinari *et al.*, 1996).

Younger metamorphic overprints in some regions of the RNJP occur in the southern part of the province as determined by zircon U/Pb ages of 1.42 Ga from gneissic rocks (Payolla *et al.*, 1998), and 1.34 Ga from granulitic rocks (Tassinari *et al.*, 1999). In the northern part of the province, superimposed geological events are shown by a Rb/Sr isochron age of 1.46 Ga obtained from the Uaupés Granitoid (Dall'Agnol and Macambira, 1992), and a Pb/Pb evaporation age of 1.52 Ga defined a two-mica granitoid from the Içana River area (Almeida *et al.*, 1997).

The virtual absence of older ages in the SHRIMP U/Pb zircon analysis; the low values of the initial $^{87}Sr/^{86}Sr$ ratios between 0.702 and 0.706; the μ_1 values around 8.1 and the positive and low negative $\varepsilon_{Nd(1.8Ga)}$ values relative to CHUR (+ 4 to - 2), suggest, for most of the Rio Negro-Juruena basement rocks, no contamination from Archean crust. However, these data suggest mixing between juvenile mantle-derived magma with some recycled subordinate older material.

Supracrustal sequences are scattered within the Rio Negro-Juruena Province. They are represented in the southern part of the province, by the Roosevelt metavolcano-sedimentary unit, which is composed by dacite, ryolite, andesite, tuff, volcanic breccia, claystone, sandstone, and banded iron formation units, metamorphosed in the greenschist facies. Zircon from dacite yielded U/Pb age of 1.74 Ga (Santos *et al.*, 1999), whereas a whole-rock Rb/Sr isochron for the volcanic rocks gave an age of 1.56 Ga (Tassinari, 1996). The older age is interpreted as the time of volcanism, and the younger age could be related to the metamorphism.

Small remnants of quartzite of Tunuí Group occur at scattered localities in the northern part of the province. The quarzite beds are crosscut by the 1.52 Ga Uaupés Granitoid (Almeida *et al.*, 1997), which allow us to suppose a time between 1.8 and 1. 52 Ga for the deposition and metamorphism of these sequences. The K/Ar cooling ages of 1.3 and 1.0 Ga obtained on muscovite from the metasediments (Pinheiro *et al.*, 1976) reflects the effect of the mylonitic Nickerie Episode with ages around 1.1 Ga (Priem *et al.*, 1973).

Sedimentary Platform Cover and Associated Magmatism

In the southern part of the Rio Negro-Juruena Province there are exposed undeformed volcano-sedimentary rocks of Caiabis Group, which can be divided into the Dardanelos and Arinos formations. Mainly volcaniclastic rocks, such as ignimbrite and tuff associated with greywacke, arkose and rhyolitic-rhyodacitic rocks compose the Dardanelos Formation. Samples from the acid volcanic rocks define a Rb/Sr reference isochron age of 1.44 Ga, which was interpreted as the age of the volcanism. Furthermore, the alkaline, tholeiitic and calc-alkaline basalt of the Arinos Formation, which is interlayered in the sedimentary rocks of the Dardanelos Formation, yielded whole-rock K/Ar ages of 1.4 and 1.2 Ga, respectively, for the lower and upper basaltic flows (Tassinari *et al.*, 1978). These geochronological results define the minimum depositional age of the sedimentary sequence. The alkaline basalt flows are associated with the 1.2 Ga Canamã alkali-syenite, which occurs in the northwestern part of the area underlain by the Caiabis Group.

The volcano-sedimentary rocks of the Caiabis Group are related to a rift system that has a NW-SE trend, produced by break-up of stable continental regions, as a reflex of orogenic activities in neighbouring areas, related to the evolution of the Rondonian-San Ignacio and Sunsas provinces. On the other hand, in the northern part of the Rio Negro-Juruena Province, where the volcano-sedimentary sequences are less commom, some volcanic rocks from the Traira River area, interlayered with some sandstone beds, yielded a Rb/Sr isochron age of 1.5 Ga (Pinheiro *et al.*, 1976).

Mafic magmatism associated with dykes swarm, which crosscut sedimentary sequences as well as basement rocks, occurs within three different periods; the older period between 1.4 and 1.35 Ga, and the two other periods ranging from 1.25 to 1.15 Ga, and between 1.0 and 0.95 Ga. These magmatic episodes are related to terminal igneous activity of the orogenies younger than 1.6 Ga. (Bettencourt *et al.*, 1995; Teixeira, 1978; Olszewski *et al.*, 1989; Tassinari, 1996).

Anorogenic Granitoid Magmatism

The anorogenic granitoid magmatism within the southern part of the Rio Negro - Juruena Province is represented by A-type granite and within-plate granite bodies, mainly of syenogranitic and monzogranitic composition, some of them with rapakivi textures, associated with gabbro, syenite, mangerite and charnockite. A rapakivi-granite suite of different age, petrology and geochemistry intruded the basement rocks of the Rio Negro-Juruena Province. These magmatic episodes were accompanied by a few bodies of anorthosite and mangerite, charnockite, mafic dykes and flows, as well as by an alkaline complex. These rocks are included in several granitic suites the ages of which are summarized in Tassinari *et al.* (1996) and Bettencourt *et al.* (1999a). These results showed that the typical anorogenic plutonism was emplaced periodically between 1.6 Ga and 970 Ma.

The rapakivi rocks in the states of Rondônia and Mato Grosso are represented by the Serra da Providência Intrusive Suite, which comprises the granite of Serra da Providência Batholith as well as satellite stocks as described by Leal *et al.* (1976, 1978) in the southeastern part of the Rondônia Tin Province. Gabbro, charnockite and mangerite were included in the suite by Rizzoto *et al.* (1996). This batholith comprises an elongated oval-shaped gabbro-charnockite-mangerite-granite *c.* 140 km long by 40 km wide (Bettencourt *et al.*, 1999a). The four main granitoid units recognized by Rizzoto *et al.* (1996) are porphyritic monzogranite (pyterlite) with subordinate wiborgite, porphyritic monzogranite, equigranular syenogranite and granite porphyre. An U/Pb zircon age reported by Bettencourt *et al.* (1999) yielded 1.606 ± 0.024 Ga for the oldest intrusion, whereas the porphyritic hornblende-biotite monzogranite has an interpreted age of 1.573 ± 0.015 Ga. Piterlite yielded an age of 1.566 ± 0.005 Ga, which is in close agreement with the zircon U/Pb SHRIMP age of 1.588 ± 0.016 Ga reported by Tassinari *et al.* (1996). The associated Ouro Preto Charnockite yielded a zircon U/Pb age of 1.532 ± 0.024 Ga (Van Schmus, personal communication).

The deformation of the Serra da Providência Intrusive Suite together with the surrounding older Rio Negro-Juruena basement probably evolved within a transpressional tectonic regime, which probably occurred during the Rondonian-San Ignacio Orogeny. This may have occurred before the emplacement of the undeformed 1.41 Ga Santo Antonio rapakivi granite (Bettencourt *et al.*, 1999a). Another possibility is that it might be chrono-correlated with the evolution of the Cachoerinha Magmatic Arc at *c.* 1.58 - 1.52 Ga as suggested by Geraldes *et al.* (1999) in the Jauru region in the southeastern part of the Rondônia Tin Province in the State of Mato Grosso.

Orogenies and Terranes in the Southern Rio Negro-Juruena Province

The Rio Negro-Juruena Province is one of the most poorly known provinces in the southwestern part of the Amazonian Craton, apart from the northern part of State of Rondônia and adjacent areas (Jamari and Roosevelt terranes) for which more precise geological and geochronological data are being obtained. This fact makes it difficult to distinguish the different orogens or terranes within this huge province, mainly in the northern part. However in the southern part of the province, in the southwestern region of the State of Mato Grosso, we can

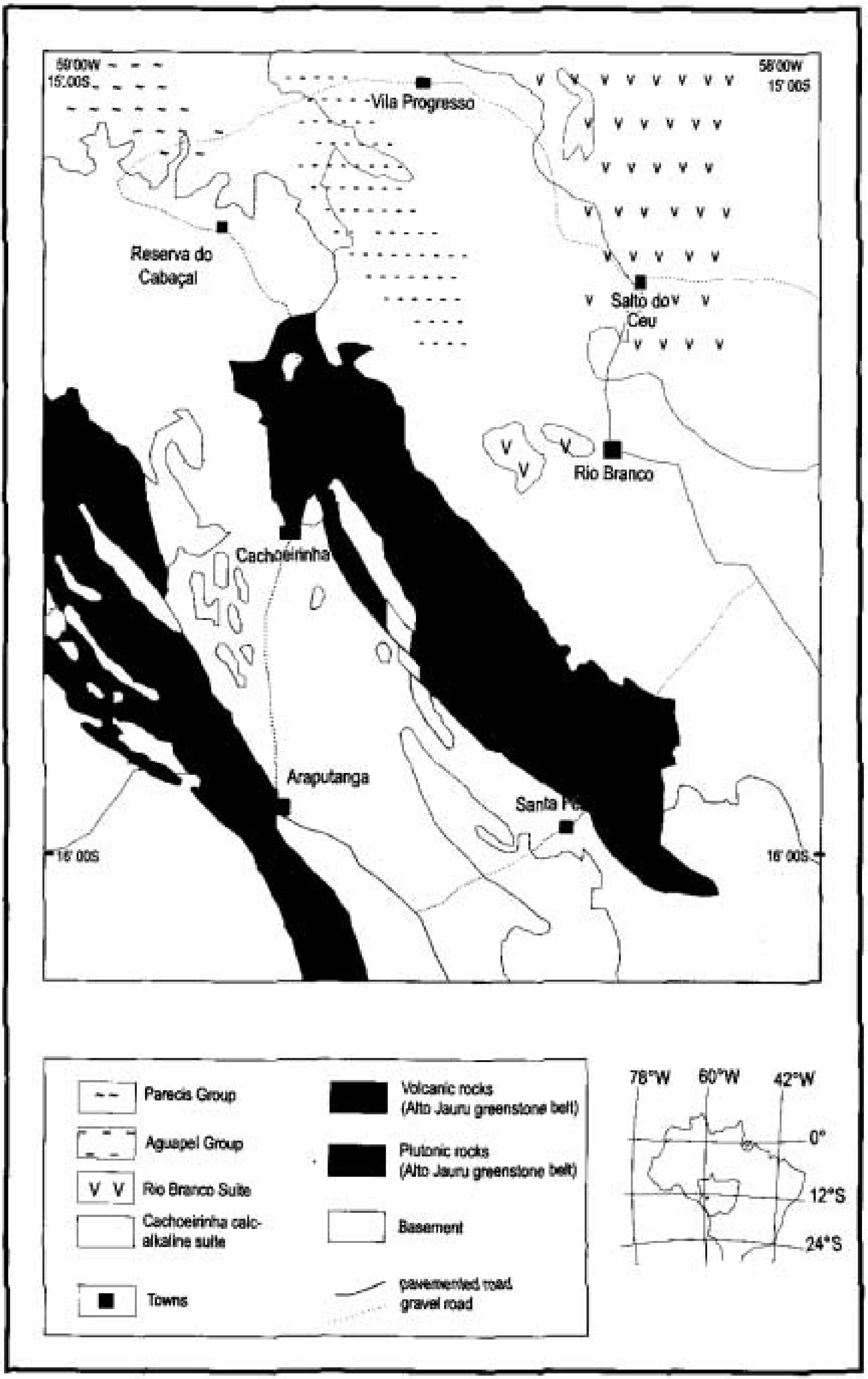


FIGURE 4 - Simplified geological map of the southern estern part of the State of Mato Grosso (SW Amazonian Craton), showing the Cachoeirinha Suite and Alto Jauru Greenstone Belt.

distinguish the Alto Jauru greenstone belt terrane and the Cachoeirinha Orogen which have been recently constrained by precise geochronological data (Geraldes *et al.*, 1999). The following discussion will highlight the main aspects of these two important units of the Rio Negro-Juruena Province.

Alto Jauru Greenstone Belt Terrane

The Jauru region, situated in the southwestern region of the RNJP in the State Mato Grosso, includes metavolcano-sedimentary sequences, interpreted by some authors as greenstone belts sequences. Teixeira *et al.* (1989) and Tassinari *et al.* (1996) considered this as an ancient nucleus preserved within the younger Rondonian-San Ignacio Province. The 1.8 - 1.55 Ga zircon U/Pb ages and 1.9 - 1.75 Ga Sm/Nd mantle depleted model ages obtained by Geraldes *et al.* (1999) on granite and orthogneiss from the basement rocks are in agreement with the RNJP geochronological pattern. The last authors considered the Jauru area as a possible extension of the RNJP.

Volcanic rocks of the Jauru region (Fig 4) were described and named initially by Saes *et al.* (1984) as the Quatro Meninas Volcanic Complex. Monteiro *et al.* (1986) named it the Alto Jauru Greenstone Belt. Three belts of metavolcanic and sedimentary rocks (from E to W named: Cabaçal, Araputanga and Jauru) are separated by granitic-gneiss terranes of tonalitic composition. The belts are intruded by Proterozoic dolerite and granitoid and are covered by Mesoproterozoic clastic sediments of the Aguapeí Group.

Monteiro *et al.* (1986) described the following sequences for the metavolcano-sedimentary unit: the lowest unit (basic volcanics) consists of massive amygdaloidal and variolitic lava and volcanic breccia, which contain lithic clasts of older rocks. Intermediate metavolcanic rocks (andesite lava and tuff, interdigited with felsic lava, tuff and metapelite) locally overlie these. The younger acid metatuff beds consist of dacite-rhyodacite lava, tuff and epiclastic rocks that host the main gold and silver-bearing ore body worked between 1984 and 1990. The acid metatuff beds are overlain by metasediments, which include local quartz-sericite-clorite schist. The interbedding of epiclastic debris is common, together with metachert and, locally, garnetiferous magnetite bands (BIFs). The aforementioned metachert and BIF units are interpreted as chemical metasediments by Pinho *et al.* (1997).

Geochronology and geochemistry of the Alto Jauru Greenstone Belt

Van Schmus *et al.* (1998) and Geraldes (2000) reported U/Pb zircon ages (Table 5) on volcaniclastic rocks (metatuff) yielding 1.767 ± 0.024 Ga, which was interpreted as the crystallization age. The Sm/Nd mantle-depleted model age of 1.87 Ga and the $\varepsilon_{Nd(T)}$ of + 2.4, indicate that the volcanism was derived from mantle-derived magma with little contamination from the supracrustal host rocks.

Plutonic rocks of tonalitic to granitic composition were described in the Jauru region as coeval with the volcanic rocks of the Alto Jauru Greenstone Belt. Ruiz (1992) mapped an area between the Cabaçal and Araputanga volcanic belts and described orthoderived gneiss (São Domingos Granite-gneiss) which is intruded by the Alvorada Granite. Carneiro *et al.* (1992) reported a Rb/Sr isochron age of 1.734 ± 0.226 Ga and initial Sr = 0.7019 for this gneiss, whereas Geraldes (2000) presented a U/Pb zircon age for the same rocks of 1.795 ± 0.01 Ga and a Sm/Nd mantle-depleted age of 1.93 Ga and ε_{Nd} = 2.16.

Ruiz (1992) defined the Aliança Gneiss, which yielded a U/Pb zircon age of 1.747 ± 0.013 Ga (Geraldes, 2000). In addition, Pinho (1996) reported a U/Pb SHRIMP result for volcanic rocks of Alto Jauru greenstone belt which yielded two age groups: the older group of 1.769 ± 0.029 Ga and the younger group of 1.724 ± 0.03 Ga, concordant with the U/Pb isotope dilution dating.

Chemical results for plutonic rocks (tonalite and granodiorite) from the Alto Jauru Greenstone Belt reported by Pinho (1990) indicate a TTG affinity, whereas the results reported by Geraldes (2000) indicated a calc-alkaline trend. In all cases, the authors interpreted these intrusive rocks as being generated in an arc-related environment. Pinho *et al.* (1997) reported chemical results obtained in volcanic rocks from Jauru Belt indicating an ocean floor origin. Different from the ultrabasic-basic rocks observed in the western part of the greenstone belt, the intermediate felsic unit has a predominantly calc-alkaline magmatic affinity. This led the authors to suggest that rocks formed at an ocean-ridge underlie the Jauru Belt (western part), and that the Cabaçal Belt was formed in an arc-related setting.

Metamorphism and Deformation

The metamorphism and deformation described in the rocks of the Alto Jauru Greenstone Belt are quite complex, and probably induced by younger superimposed events. After the crust-forming episode (1.79 - 1.74 Ga), this unit was intruded by a calc-alkaline suite dated at 1.57 - 1.52 Ga that resulted in migmatization and deformation of the rocks of the Alto Jauru Greenstone Belt. Subsequently, the development of the Santa Helena Arc (1.48 - 1.42 Ga) and the Sunsas/Aguapeí events (1.2 Ga - 950 Ma) undoubtedly overprinted the original structures as suggested by the Rb/Sr (1.7 - 1.4 Ga) and K/Ar (1.4 - 1.0 Ga) resetting ages.

Cachoeirinha Orogen

An important rock-forming event occurred between 1.57 - 1.53 Ga in the southwestern part of the border of the Amazonian Craton, and the orogenic products have been considered as part of the Rio Negro-Juruena Province. This event formed arc-related rocks chemically comparable with a calc-alkaline suite, varying from tonalite to granite, that intruded the older basement rocks, represented, locally, by the Alto Jauru Greenstone Belt (1.79 - 1.75 Ga), situated in the same geographic area (Fig. 2). The Cachoeirinha Orogen is bounded to the N by the Cretaceous Parecis sedimentary cover, to the S and E by the Neoproterozoic Paraguai and Araguaia-Tocantins belts and to the W by the Lucialva Lineament, separating the Cachoeirinha related-rocks from those of the younger Santa Helena Magmatic Arc.

The rocks that are now considered as belonging to the Cachoeirinha Orogen, were initially described by Figueiredo *et al.* (1974) and Barros *et al.* (1982) as belonging to the Xingu

Complex. Carneiro *et al.* (1992) suggested that the rocks in the São José do Quatro Marcos area underwent at least two rock-forming episodes. The first episode is represented by grey gneiss (Rb/Sr ages about 1.96 Ga), and the second episode is manifest by pink gneiss and granite with ages from 1.74 to 1.4 Ga (Rb/Sr ages). Granitoid plutons of different composition, with ages ranging from 1.7 to 1.4 Ga, have been separated from the basement (Saes *et al.*, 1984; Leite, 1989; Ruiz, 1992). Tassinari *et al.* (1996) reported U/Pb and Rb/Sr results constraining two major events for the Rio Negro-Juruena Province. They interpreted these as successive magmatic arcs, the older event from 1.8 to 1.7 Ga, and the younger event from 1.6 to 1.5 Ga. The latter event included some reworked material from the older magmatic arc. Recent U/Pb geochronology (Van Schmus *et al.*, 1998; Geraldes *et al.*, 1999; Geraldes, 2000) best constrain these two events in the so-called Jauru region, the first defined as the Alto Jauru Greenstone Belt (1.79 - 1.74 Ga), and the second as the Cachoeirinha Orogeny (1.57 - 1.53 Ga).

Tonalite, granodiorite and granite are the main rock-types related to the Cachoeirinha Orogen. The rocks are grey, equigranular, medium to coarse-grained and with mafic minerals that define the foliation. Plagioclase, quartz, amphibole and biotite comprise the main minerals. Zircon U/Pb ages from the tonalite yielded ages of 1.536 ± 0.011 and 1.549 ± 0.01 Ga, and the Sm/Nd mantle depleted ages are 1.77 Ga with $\varepsilon_{Nd} = +0.5$ and 1.83 Ga with $\varepsilon_{Nd(T)} = +1.0$, respectively (Table 6).

Granodiorite associated with the São Domingos Gneiss is observed in some parts of the Santa Cruz Batolith(Ruiz, 1992). They show lateral variations within the batholiths and display a strong foliation and banding. They are quartz-feldspatic in composition, and occur in felsic layers and in amphibole-enriched mafic layers. The Santa Cruz sample yielded an U/Pb zircon age of 1.587 ± 0.004 Ga ($T_{DM} = 2.05$ and $\varepsilon_{Nd(T)} = -0.8$), and the São Domingos Gneiss presented a zircon U/Pb age of 1.562 ± 0.036 Ga ($T_{DM} = 1.79$ and $\varepsilon_{Nd(T)} = +0.9$).

The gneiss-migmatite rocks associated with the Cachoeirinha Orogen present complex deformation (Ruiz, 1992), which produced irregular folds and discontinuous banding consisting of quartz feldspar-rich and biotite-rich layers. Recrystallized quartz grains with irregular contacts with feldspar and a mozaic fabric are interpreted (Ruiz, 1992) as the result of high-grade metamorphism, and the biotite orientation in bands is due to deformation.

Granitic rocks within the Cachoeirinha Orogen have a widespread geographical distribution over the Jauru region. The Cachoeirinha and Quatro Marcos granites yielded zircon U/Pb ages of 1.522 ± 0.011 Ga ($T_{DM} = 1.78$ and $\varepsilon_{Nd(t)} = +0.9$) and 1.537 ± 0.06 Ga ($T_{DM} = 1.75$ and $\varepsilon_{Nd(T)} = +0.5$), respectively.

Anorogenic plutons of granitic and granodioritic composition are described in the Jauru region. The Água Clara Granodiorite (Saes *et al.*, 1984; Monteiro *et al.*, 1986; Matos *et al.*, 1996) is the largest anorogenic batholith in the Jauru region (600 km^2) and consists mainly of granodiorite with subordinate tonalite and granite (Matos *et al.*, 1996). Geraldes (2000) reported a zircon U/Pb age of 1.485 ± 0.04 Ga, which was interpreted as the crystallization age, and a Sm/Nd mantle-depleted model age of 1.77 Ga and $\varepsilon_{Nd(T)} = 1.7$, suggesting a Rio Negro-Juruena crust as their magma source. The same interpretation may be considered for the Alvorada and Araputanga granite described by Monteiro *et al.* (1986), which show U/Pb ages of 1.389 ± 0.011 Ga ($T_{DM} = 1.77$ and $\varepsilon_{Nd(T)} = -1.3$) and 1.44 ± 0.06 Ga ($T_{DM} = 1.74$ and $\varepsilon_{Nd(T)} = -0.4$), respectively.

Rondonian - San Ignacio Province (RSIP)

The term Rondonian Province was introduced by Cordani *et al.* (1979), for an important tectono-magmatic event in the 1.45 - 1.25 Ga interval, which developed on the southwestern margin of the Rio Negro-Juruena Province (1.75 - 1.55 Ga). Subsequently, Teixeira and Tassinari (1984), based on more than 200 Rb/Sr and K/Ar determinations, defined the Rondonian Geochronological Province, which was represented by a mobile belt, aged 1.45 - 1.0 Ga. Rocks belonging to this orogen extended from the Ituxi-Abunã region (western part of the State of Rondônia) to the Jauru region (State of Mato Grosso) extending to the San Ignacio-Tunas region in Bolivia, thus including the rocks attributed to the San Ignacio Orogeny by Litherland and Bloomfield (1981). More recently, Tassinari *et al.* (1996) proposed the term Rondonian-San Ignacio Province (1.5 to 1.3 Ga), in order to include rocks belonging to the same event, both in Brazil and Bolivia.

General Geological Aspects

The Rondonian-San Ignacio Geochronological Province lies in the southwestern part of the Amazonian Craton, and it is bounded to the E by the Rio Negro-Juruena Province, and to the S and SW by the Sunsas Province. In several places the boundaries between these provinces are not very well defined, due to the fact that in most areas the effects of younger metamorphic overprinting are very strong.

The 1.55 to 1.3 Ga metamorphic basement, previously ascribed to the Xingu (Leal *et al.*, 1978) and Jamari complexes (Isotta *et al.*, 1978), is composed of granite-gneiss-migmatite terranes and granulitic rocks. In general, these rocks are metamorphosed in the amphibolite or granulite facies. Some Paleoproteorozic granulitic inliers, such as the Lomas Manéches Group in Bolívia (Litherland *et al.*, 1986) are scattered within the province. The Rondonian-San Ignacio Province (RSIP) shows strong NE-SW overprint structures, and is underlain by some regions with a clear ensialic character and others with juvenile magmatic characteristics.

The geochronological pattern for the RSIP basement rocks, include a group of Rb/Sr age determinations between 1.5 and 1.37 Ga, with $^{87}Sr/^{86}Sr$ initial ratios ranging from 0.703 to 0.710 (Amaral, 1974; Teixeira and Tassinari, 1984; Priem *et al.*, 1989; Tassinari, 1996) and zircon U/Pb ages varying from 1.49 to 1.33 Ga (Geraldes *et al.*, 1999). The granitoid plutons of the RSIP yielded two groups of Sm/Nd depleted-mantle model ages. The first and younger group obtained on samples from the Pontes de Lacerda region, around 1.5 Ga, which was considered by Geraldes *et al.* (1999) as being related to the Santa Helena Volcano-Plutonic

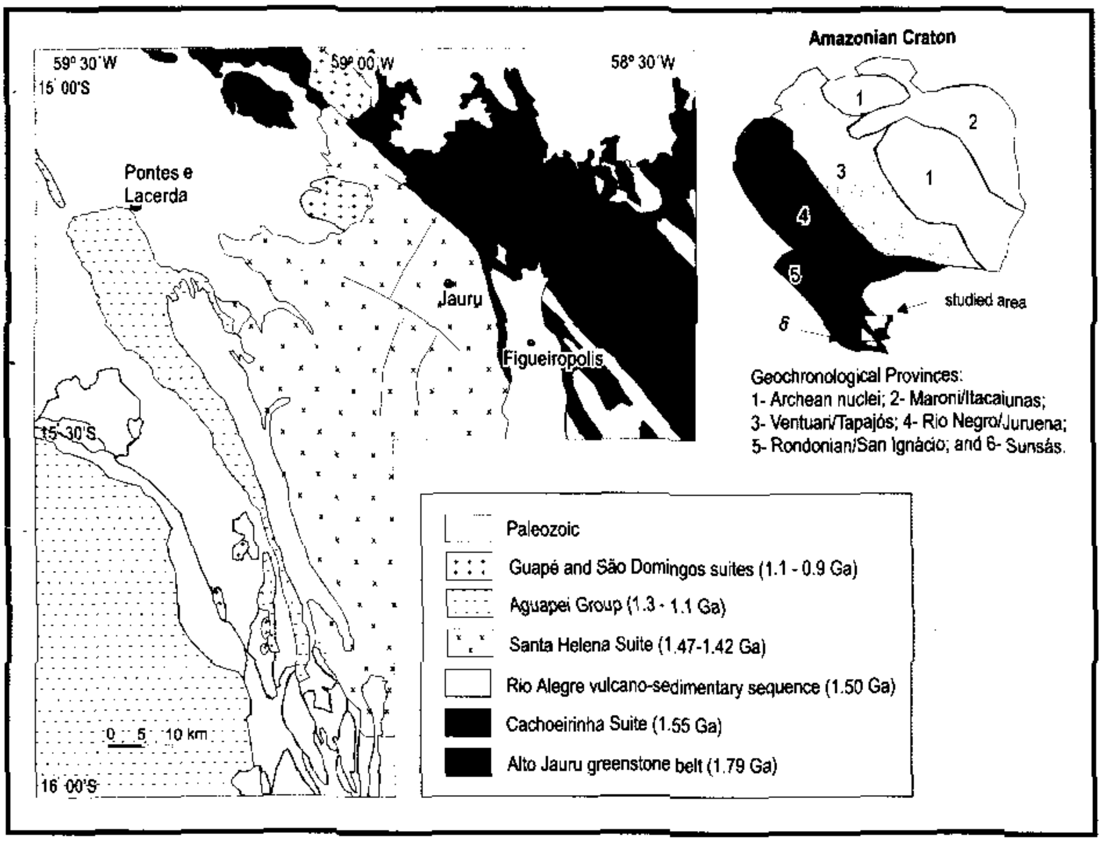


FIGURE 5 - Major geological units of Pontes e Lacerda region (Mato Grosso). The Aguapeí Thrust Belt marks the boundary between Rio Alegre Volcanosedimentary Sequence and the Santa Helena Suite. Santa Helena Suite and Cachoeirinha Suite are separated by the Taquaruçu-Lucialva Lineament.

Arc. The second and older group was composed of samples from the entire Rondonian-San Ignacio Province, with ages ranging from 2.0 to 1.7 Ga. This interval coincides with the estimated mantle-extraction period of the Rio Negro-Juruena Province (Geraldes *et al.*, 1999; Sato and Tassinari, 1997). The older ages indicate the development of crustal reworking within the province.

Metavolcano-sedimentary sequences, represented by the Ascencion Group, in Bolivia, indicate a Rb/Sr metamorphic age of 1.34 Ga (Litherland *et al.*, 1986). The K/Ar cooling ages between 1.35 and 1.3 Ga suggest the beginning of the tectonic quiescence period of the province. Post-orogenic and anorogenic magmatic activity is represented by the Santo Antonio Intrusive Suite (1.406 Ga), the Teotônio Intrusive Suite (1.387 Ga), the Alto Candeias Intrusive Suite (1.346 and 1.338 Ga), the São Lourenço-Caripunas Intrusive Suite (1.31 Ga), the Santa Clara Suite (1.08 Ga), the Costa Marques Group (1.02 Ga) and the Yonger Rondônia Granite (990 Ma), Bettencourt *et al.* (1999). The occurrence of the 1.1 Ga undeformed acid volcanism is very scattered within the province. Sedimentary rocks associated with alkaline basalt flows were deposited around 1.1 - 1.0 Ga in the NW-SE continental rift system that corresponds to the age of the basalt that intruded the lower sedimentary sequences of the Guajará Mirim Group.

In this review the term Rondonian-San Ignacio Geochronological Province is maintained, as defined by Tassinari and Macambira (1999), including one or more terranes and/or orogenies, where their respective geological evolution took place within 1.55 to 1.3 Ga period. In this way, recent geological and geochronological studies (Scandollara *et al.*, 1999b; Geraldes *et al.*, 1999) have led to the suggestion that the RSIP could include different terranes and orogenies, which are: (1) The Río Alegre Terrane; (2) The Santa Helena Orogen; and (3) The Rondonian-San Ignacio Orogen.

The Rio Alegre Terrane

The Rio Alegre Terrane comprises mafic-ultramafic plutonic and volcanic rocks associated with banded iron formation units and chert, which occur in the Rio Alegre Valley in the State of Mato Grosso (Fig. 5). They were initially described by Barros *et al.* (1982), and were correlated to the Rincon del Tigre Complex by Litherland *et al.* (1986). Menezes *et al.* (1993) included these rock-associations in the Pontes e Lacerda Volcano-sedimentary Sequence. Matos (1992) prepared a detailed geological map of this unit and renamed it as the Rio Alegre Volcano-sedimentary Sequence. Pinho (1990) correlated the mafic metavolcanic rocks of the Rio Alegre Valley with the Alto Jauru Greenstone Belt, but this correlation was not confirmed by new isotope data.

This unit occurs in a large area (50 x 200 km) bordered to the E by the Santa Helena Batholith and the Alto Jauru Greenstone Belt. The boundary to the S is covered by the sedimentary rocks of the Pantanal Formation (Quaternary) To the W it is overlain by flat-lying sediments of the Aguapeí Group; and to the N the boundary is not known.

Stratigraphic Units

The magmatic and volcanic metamorphosed rocks observed in the Rio Alegre Valley may be subdivided (Matos, 1992) as follow: (1) The Minouro Formation, consisting of metabasite, fine-grained with local porphyrytic (hornblende) textures, associated with fine-grained banded iron formation units (with magnetite-bearing layers), chemical sediments, chert, and clastic rocks; (2) The Santa Isabel Formation, which consists of metabasalt, metapyroclastic rocks and metariodacite presenting a high degree of oxidation and lateritization, and; (3) the São Fabiano Formation, which includes clastic and chemical metasediments (phylite, quartzite and carbonaceous layers), chert, and metavolcaniclastic rocks, including garnet-kyanite bearing muscovite-biotite schist. Amphibolite and meta-ultrabasic rocks, with nematoblastic textures, are subordinate units interlayered in mica-schist. The intrusive rocks are composed of differentiated complexes including gabbro and serpentinite with cumulate textures, metamorphosed in the greenschist facies.

Lithogeochemistry

Matos (1992) and Matos and Schorscher (1997a, b) based on geochemical studies on metavolcanic and meta-intrusive rocks from the Rio Alegre Terrane, suggested a sub-alkaline signature for these rocks, and a back-arc ocean-floor environment. Mineralogical alterations in these rocks are typical of ocean floor metassomatism including epidotization, carbonatization and sericitization. The geochemical data for the intrusive rocks led the authors to conclude that these resulted from the evolution and differentiation of tholeiitic magma.

Menezes *et al.* (1993) presented REE results for metabasalt from the Santa Isabel Formation showing MORB or immature island arc patterns. The authors suggest a magmatic origin from an enriched-mantle source or from continental margin collision. The existence of ocean-floor related rocks, metamorphosed in the greenschist facies, cut by pyroxenite and amphibolite may be interpreted as a collisional suture. In this way, future research on this unit might take into account the possibility that the rocks of the Rio Alegre Terrane might be an ophiolitic complex.

The major structural features characteristic of the Rio Alegre Terrane is a strong transposition of metasedimentary and metavolcanic rocks. Menezes *et al.* (1993) described this process as the result of an intense mylonitization. The main foliation strike is 30° - 50° NW, and the dip is 20° - 70° (Menezes *et al.*, 1993), indicating strain parallel to the border of the Amazonian Craton. The lineation variation (NW and SE) may have resulted from a progressive deformational event, where initial sub-horizontal foliation (thrusting) changed to a subhorizontal mylonitic foliation formed during strike-slip movement under ductile conditions.

Geochronological and Isotopes Constraints

Magmatic activity of the Rio Alegre Terrane occurred during a short period between 1.509 and 1.494 Ga (Table 7) (Geraldes, 2000). Sm/Nd results obtained on Rio Alegre rocks (T_{DM} from 1.67 to 1.54 Ga and $\varepsilon_{Nd(T)}$ values from + 2.5 to + 4.8, Table 7) suggest a juvenile origin for their magma sources. The Rio Alegre petrotectonic association represents an ocean floor complex associated with clastic and chemical sedimentary rocks, resembling an accretionary complex. This accretionary complex may have collided with the southwestern margin of the Amazonian Craton just after the evolution of the Santa Helena Arc (1.47 - 1.42 Ga) since the Santa Helena Batholith comprehends a region between the Rio Negro Juruena Province (E) and the Rio Alegre Accretionary Complex (W) (Geraldes *et al.*, 2000).

The Santa Helena Orogen

Proterozoic basement rocks, mainly of granitic composition, were initially recognized by Saes *et al.* (1984), who created the term Santa Helena Batholith to include the referred rocks. Subsequently, these rocks were studied by Menezes *et al.* (1993) and Lopes *et al.* (1992), who renamed the batholith as the Santa Helena Granite-Gneiss and, based on geochemical data, classified the associated rocks as intraplate A-type granite. Additional geological, geochemical and geochronological studies on this unit were carried out by Geraldes *et al.* (1997) and Van Schmus *et al.* (1998), which led the authors to propose the term Santa Helena Suite. The authors described igneous and meta-igneous rocks, represented by tonalite, orthogneiss and granite, as a calc-alkaline arc-related suite of 1.48 to 1.42 Ga. The extensive arc-related magmatism together with the new isotope and geochemical results provided a consistent record of this event, and a proposal is now made to elevate the Santa Helena Suite to orogen status.

The Santa Helena Orogen is bordered to the W by both the Rio Alegre volcano-sedimentary sequence and the Aguapeí Thrust Belt (Fig. 5). Limits to the E include several domains of different rock-types, including the Alto Jauru Greenstone Belt (1.79 - 1.75 Ga) and the Cachoeirinha Orogen (1.57 - 1.53 Ga). The northern and southern limits extend beneath the Paleozoic sedimentary cover.

Geochronological Results

The rocks of the Santa Helena Suite consist of granodiorite to granite and tonalite; the tonalite occurring mainly in the western part of the batholith (Lavrinha and Pau-a-Pique tonalite) (Geraldes, 2000). These rocks are leucocratic, grey to green in colour, medium to coarse-grained, and isotropic to slightly foliated. U/Pb zircon data from a hornblende-tonalite yielded an upper intercept (crystallization age) at 1.467 ± 0.025 Ga (Table 1), and the Pau-a-Pique body yielded U/Pb results with an upper intercept age at 1.481 ± 0.047 Ga. The Sm/Nd mantle-depleted model ages for these tonalite bodies are 1.53 Ga ($\varepsilon_{Nd(T)}$ = + 3.8) and 1.5 Ga ($\varepsilon_{Nd(T)}$ = + 4.1) respectively, indicating that in both cases the magma was derived from a source containing a very small, if any, contribution, of older continental crust (Table 8).

Granodiorite and some amphibolite are observed in the northern, central and western parts of the Santa Helena Batholith. The Alto Guaporé Gneiss, the Guaporé and the Triângulo granodiorite are grey in colour, and locally banded (granite-gneiss, biotite-gneiss and magnetite-gneiss). These rocks have zircon U/Pb ages of 1424±15 Ma and 1445±04 Ma, and Sm/Nd model ages (T_{DM}) from 1.56 to 1.49 Ga, with $\varepsilon_{Nd(T)}$ values varying from 2.9 to 4.0, indicating that the protolith of this gneiss contained a very small component of older crust.

The granitic rocks have a homogeneous distribution within the batholith. They show restricted compositional and textural variations and are mainly represented by grey to pink, usually equigranular, biotite granite. Previous Rb/Sr results for the Santa Helena Granite-Gneiss and Maraboa Pluton (Geraldes, 1996) yielded, respectively, 1.318 ± 0.024 Ga and 1.275 ± 0.125 Ga. Zircon fractions from Santa Helena Granite-Gneiss yielded an upper intercept (crystallization age) of 1.433 ± 0.06 Ga on the U/Pb concordia diagram. The U/Pb results for the Maraboa Granite, when plotted on a U/Pb concordia diagram yielded an upper intercept (crystallization age) at 1.449 ± 0.007 Ga. Sm/Nd model ages (T_{DM}) for these units are 1.62 Ga ($\varepsilon_{Nd(T)}$ = 3.1) and 1.7 Ga ($\varepsilon_{Nd(T)}$ = 2.5), respectively, indicating that the original granitic magma was derived from a source containing an older Rio Negro-Juruena continental crust component.

Other granitic rocks such as the Alto Guaporé Augen-gneiss, the Cardoso Magnetite-granite, the Santa Elina Granite and the Ellus Granite, were analyzed by the U/Pb technique, and when plotted on concordia diagram, yielded ages of 1.423 ± 15 Ga and 1.444 ± 0.021 Ga. The Sm/Nd model ages (T_{DM}) for these granites range from 1.57 to 1.51 Ga and $\varepsilon_{Nd(T)}$ values vary from + 2.7 to + 3.7, indicating that the parental magmas are composed mainly by mantle-derived material.

Deformation and Metamorphism

The structural analysis of the Santa Helena Batholith shows, according to Menezes *et al.* (1993), a penetrative foliation defined by mafic minerals or elongated k-feldspar and quartz. The regional strike, observed in augen-gneiss, is NNW, and the plunge direction varies from 30° - 60° NE, with the some oscillation to SE. The linear and planar structures of the Santa Helena Pluton are the result of a regional mylonitization, characterized by brecciation and quartz-enrichment at brittle conditions (Menezes *et al.*, 1993). U/Pb ages from zircon collected in centimetric veins sub-concordant with the augen-gneiss foliation yielded a *c.* 1.39 Ga (Geraldes, 2000), defining the deformation between 1.42 Ga (U/Pb age of vein host rock) and 1.39 Ga. This interval is interpreted as the minimum age for Santa Helena metamorphism.

Rock Geochemistry

A more recent geochemical study (Geraldes, 2000) indicates that the rocks vary from quartz monzogabbro and tonalite through granodiorite to granite *sensu latu*. The chemical results indicate a volcanic arc granite (VAG) affinity for the primitive and intermediate rocks, and the granite plots near the boundary of VAG and the within-plate granite (WPG), which is in agreement with the results of Menezes *et al.* (1993) that show that the granitic rocks of the Santa Helena Orogen may be classified as an A-type granitic suite.

Trace element modeling (Bell *et al.*, 1999) suggests that the Santa Helena rocks can be included within two groups: the first group with high Sr content, low Rb/Sr ratios, and relatively steep REE patterns (La/Ybn = 20) in which tonalite and granodiorite dominate. The second group has low Sr, high Rb/Sr ratios and relatively flat REE patterns (La/Ybn = 3.5) with large negative Eu anomalies in which granite dominates. Geraldes *et al.* (1999) reported REE patterns showing a higher fractionation between LREE and HREE in the primitive rocks than in the intermediate and fractionated ones. A positive Eu anomaly was found in the primitive rocks, a light negative Eu anomaly in the intermediate rocks, and a strong Eu negative anomaly in the granite, suggesting that the granitoid plutons were formed during a fractional crystallization.

The Anorogenic Magmatic Activities

The Rio Branco Suite characterizes the anorogenic magmatic activities related to development of the Santa Helena Orogeny. These were formerly described by Oliva (1979) as the Sierra Rio Branco Complex. Barros *et al.* (1982) carried out a detailed study, defining the Rio Branco Group. Subsequently, Leite *et al.* (1986) mapped the Rio Branco region and named these rocks as the Rio Branco Intrusive Suite, which was regarded as a differentiated layered complex. Geraldes *et al.* (1999) considered the anorogenic Rio Branco suite as part of a bimodal igneous suite, comprising coeval mafic and felsic rocks. The units are confined within the volcanic-plutonic rocks of the *c.* 1.77 - 1.52 Ga Rio Negro-Juruena Province (Bell *et al.*, 1999).

The felsic part of the Rio Branco Suite includes red to pink granite. The mafic members consist of gabbro, tholeiitic diabase dykes, and porphyritic basalt. Centimetric crystals of alkali-feldspar are bordered by plagioclase (rapakivi texture) in a plagioclase-bearing groundmass. Trace element content (Rb, Y and Nb) indicate a within-plate granite affinity, according to tectonic discrimination diagrams.

U/Pb isotope analyses of zircon from the granite yielded an upper intercept on concordia diagram of 1.423 ± 0.002 Ga, which we interpret as the crystallization age of the felsic rocks. U/Pb analyses of zircon from a gabbro yielded an age of 1.456 ± 0.024 Ga, in agreement with the felsic member ages. The $\varepsilon_{Nd(T)}$ values for mafic rocks range from + 1.24 to + 1.91, suggesting a mantle protholith. The $\varepsilon_{Nd(T)}$ values for felsic rocks are in the range from + 0.16 to - 0.96, which suggest that these rocks had a crustal component in their magma. Sm/Nd mantle-depleted model ages vary from 1.8 to 1.73 Ga for the mafic rocks and from 1.89 to 1.81 Ga for felsic rocks. Similar ages were found in the surrounding basement of the Alto Jauru Greenstone Belt.

The Rondonian-San Ignacio Orogeny

The Rondonian-San Ignacio Orogeny was first proposed by Litherland and Bloomfield (1981), and subsequently defined by Litherland *et al.* (1986, 1989) in Bolivia as the San Ignacio Orogeny. According to these authors, the orogeny is ascribed to the San Ignacio Schist Group (Rb/Sr metamorphic age around 1.344 ± 0.018 Ga). It was accompanied by a significant syn to post-tectonic granitoid magmatism, represented by the potassic calc-alkaline complex (Rb/Sr ages about 1.32 to 1.28 Ga) and by

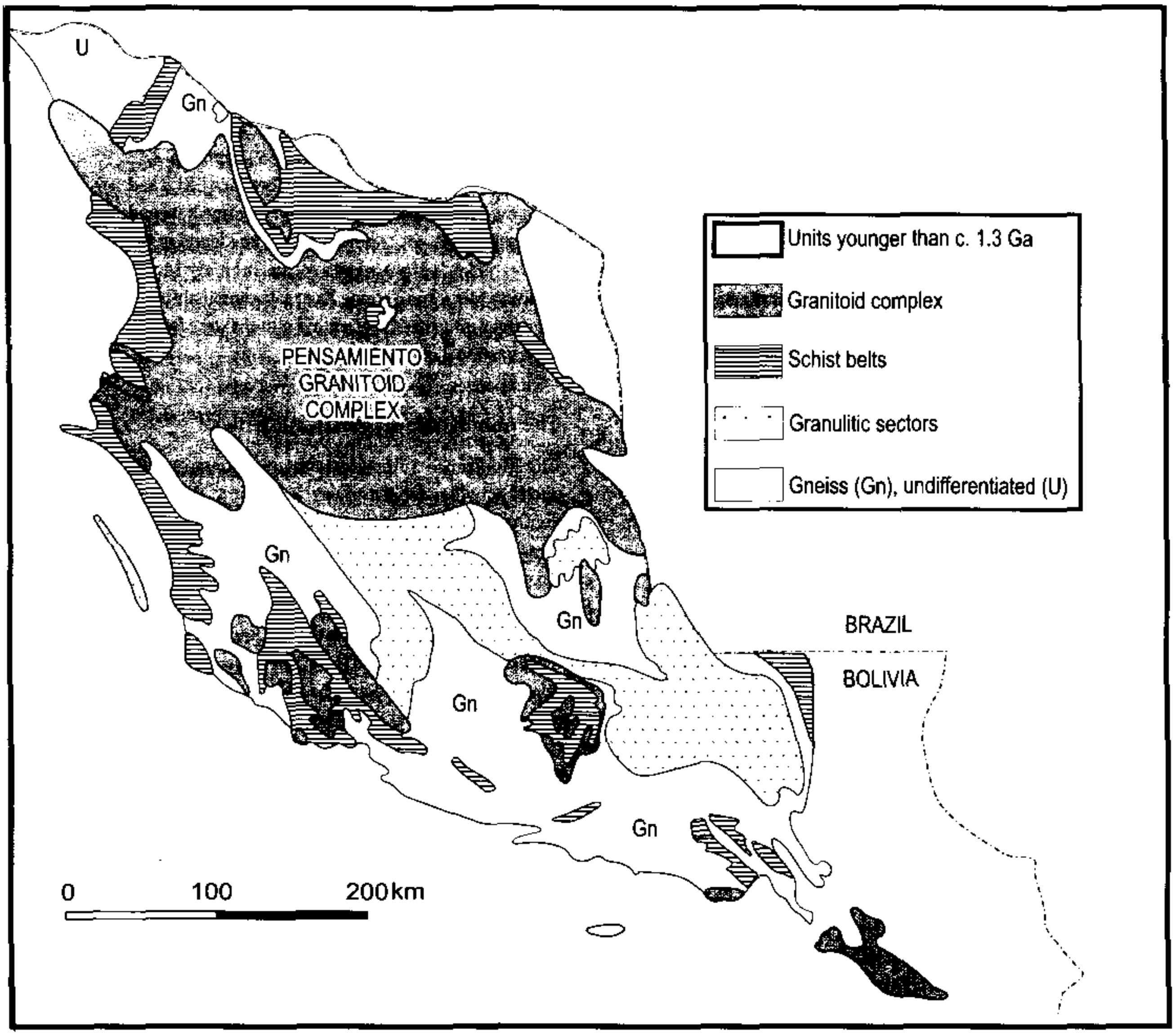


FIGURE 6 - Major geological units of c. 1.3 Ga or older in eastern Bolivia (modified after Litherland et al., *1986).*

the El Tigre Alkaline Complex (1.286 ± 0.046 Ga). In Brazil the correlative event is the Rondonian Episode (Rb/Sr ages from 1.5 to 1.3 Ga) of Teixeira and Tassinari (1984), which led them to propose the term Rondonian-San Ignacio Province which, at this time, unified both cycles. In the State of Mato Grosso, the Santa Helena Orogeny recently defined by Geraldes (2000), occurred within the same interval (1.48 - 1.42 Ga). The difference between both orogenies (Santa Helena and Rondonian-San Ignacio) is that the first has a magmatic arc evolution with a predominant accretion of juvenile material to the crust, whereas the Rondonian-San Ignacio Orogeny is characterized by an ensialic evolution with strong involvement of crustal reworking.

The San Ignacio event is manifest in widespread magmatism in eastern Bolivia, where the *c.* 1.3 Ga orogenic data are preserved in both the Rb/Sr and K/Ar systems. The northern border is not well defined due to the uncertainty of the extension of the Rio Alegre Terrane in the State of Rondônia, which is the eastern limit (Fig. 6). The cover of the Chaco Plain defines the western border. Farther to the S, within the basement rocks of the Sunsas Orogen, the K/Sr systematics are considerably disturbed, and yielded ages strongly reset by the 1.3 - 1.0 Ga Sunsas Orogeny.

The northern part of the State of Rondônia, and adjacent areas are here understood to be the segment of crust bounded to the S by the Nova Brasilândia Terrane (Sunsas Province) and to the N and E by Cenozoic and Mesoproterozoic sedimentary cover. This region has been subdivided by Scandollara *et al.* (1999b) into the Jamari, Roosevelt and Nova Brasilândia terranes, and the former subdivided in two domains: the Ariquemes/Porto Velho Domain and the Central Rondônia Domain. Up to now, the full extension of the Rondonian-San Ignacio event has not been reliably determined because the region was affected by three orogenic events: Rio Negro-Juruena, Rondonian-San Ignacio and Sunsas-Aguapeí. These overprint effects make it difficult to reconstruct the geological history of older events.

The following major lithological units represent the Rondonian-San Ignacio event (1.47 - 1.3 Ga) in Rondônia: minor remnants of felsic fine-grained syenogranite, gneiss, extensive bimodal rapakivi magmatism and AMCG magmatism in the São Lourenço, Caripunas and Lábrea regions. In addition, the Rondonian episode produced in that area widespread thermal effects and isotope resetting in the older rocks, high-grade metamorphism and sedimentary cover. Noteworthy, are the scarcity of supracrustal rocks and the small amounts of felsic/mafic gneiss suspected to be of igneous origin.

Pre-Rondonian-San Ignacio Basement Rocks

The metamorphic basement in Bolivia mainly comprises older rocks, in general evolved prior to the development of the *c.* 1.4 - 1.28 Ga San Ignacio Orogeny (Litherland *et al.*, 1986). The rocks consist essentially, of schist, gneiss and granulite. All dip steeply, and are of predominantly metasedimentary origin. These rocks are included within the Lomas Maneches Granulite Complex (hyperstene-bearing granulite, cordierite granulite) and Chiquitania Gneiss Complex, and both units have a complex geological history. The granulite and gneiss gave an interpreted Rb/Sr whole-rock regional reference isochron of *c.* 1.9 Ga (Litherland *et al.*, 1986) and Sm/Nd mantle-depleted model ages of *c.* 2.0 - 1.9 Ga (Darbyshire, personal communication.).

In Brazil, within the area affected by the Rondonian-San Ignacio Orogen, there occur some inliers composed of older rocks with ages between 1.77 and 1.47 Ga, which are mostly described in more detail by Payola *et al.* (1998), mainly in the Rondônia Tin Province. The pre-existing basement rocks are composed mostly of tonalitic gneiss and metasedimentary gneiss. These constitute part of remnant older crust referred to the Rio Negro-Juruena event. Granitoid and orthogneiss that represent three intraplate bimodal magmatic events developed over Rio Negro-Juruena crust at 1.57 - 1.56 Ga, 1.54 Ga, and at 1.53 Ga. These are related to the rapakivi Serra da Providência Intrusive Suite and associated rocks. They occurred at 1.42 Ga and are associated with the Rondonian-San Ignacio event. The Sm/Nd mantle-depleted model ages indicate that the parental magma of the granitoid plutons and orthogneiss were derived from a source containing a significant component of older crustal material (Geraldes *et al.*, 1999; Payolla *et al.*, 1998), and/or derived from a mixture of crustal and mantle sources (Payolla *et al.*, 1998).

Payola *et al.* (1998) based on U/Pb and Sm/Nd data defined three age-groups of the pre-existing Rondonian rocks: (1) greyish tonalitic gneiss (1.75 - 1.73 Ga; U/Pb ages and T_{DM} ages 2.2 - 2.06 Ga); (2) granitoid and ortogneiss (1.57 - 1.53 Ga, U/Pb ages with T_{DM} ages of 1.89 - 1.84 Ga) and; (3) fine-grained gneiss (1.42 Ga, U/Pb ages) and T_{DM} ages 1.75 Ga. These geochronological data clearly suggest the involvement of Paleoproterozoic material in the Rondonian-San Ignacio Orogen, showing their predominantly ensialic character.

The available K/Ar ages for Rondônia (Teixeira and Tassinari, 1984) reveal a complex thermal history for the area. K/Ar data on biotite and amphibole from basement rocks show three distinct cooling ages: 1.36 - 1.3 Ga, 1.26 - 1.24 Ga, and 1.1 Ga - 970 Ma. These were interpreted as reflecting the regional cooling related to three tectonic episodes. A zircon U/Pb SHRIMP age of *c.* 1.34 Ga obtained on high-grade granulitic rocks near Ariquemes (northern Rondônia) by Tassinari *et al.* (1999) has demonstrated the presence of this orogenic event in the area. This age is coeval with the emplacement of the Alto Candeias Batholith (1.34 - 1.33 Ga), which occurred in the same area. Despite the interpretation that considered the age of 1.34 Ga as the time of regional metamorphism assigned to the Rondonian-San Ignacio Orogeny, we also consider that the U/Pb age of 1.34 Ga could represent the Alto Candeias thermal event that resulted in the granulitization of the older Rio Negro-Juruena Gneiss. Recent U/Pb monazite age and Sm/Nd garnet-whole-rock isochron age of gneissic rocks of northern Rondônia suggest high-grade metamorphism at 1.33 - 1.31 Ga, which corroborates the SHRIMP results.

Metasedimentary units

The San Ignacio Supergroup crops out in the form of discrete belts throughout the Bolivian Precambrian area. This supergroup is mainly composed of quartzite, feldspatic metapsammitic and micaceous schist or phyllite with subordinate ferruginous, calc-silicate, metavolcanic and graphite-rich units (Litherland *et al.*, 1986). The basin was filled at some stage with arkose and lithic-feldspathic sandstone derived, presumably, from some nearby uplifted basement (Parágua Craton of Litherland *et al.*, 1986). The pelitic units mark a period of quieter deposition in a shallow-water, extending over the entire region. Ferruginous and carbonaceous shale, ironstone and volcanic rocks, which probably represent lower-grade relics of the depositional stage, should not be immediately interpreted as a magmatic arc (Litherland *et al.*, 1986), but could represent a continental margin prism related to the Rondonian-San Ignacio Orogen.

Metamorphism and Deformation

According to Litherland *et al.* (1986), two main phases of penetrative deformation accompanied by low to high-grade metamorphism have been related to the San Ignacio tectonic event. (1) The N to NE trending D_2 event involved upright folding with recumbence to the NE, which was accompanied by widespread migmatization. (2) The essentially NW rending, upright, D_3 folding that was accompanied by the generation of syn-kinematic K-feldspar-bearing granite dated at *c.* 1.35 Ga. Higher-level granophyre complexes and smaller mafic intrusions were also developed. D_3 was followed by phases of essentially non-penetrative deformation, which controlled the emplacement of some late to post-kinematic granitoid plutons and layered alkaline complexes.

Magmatic Units

A summary of the geology of the granitoid plutons of eastern Bolivia has been presented by Litherland *et al.* (1986, 1989). The San Ignacio event was accompanied by extensive syn to late-kinematic granite magmatism, which was designated by Litherland *et al.* (1986) as the Pensamiento Granite Complex. The age of the entire suite is contained within the interval of 1.4 to 1.25 Ga. The syn to late-tectonic magmatism is represented by the dominantly potassic calc-alkaline granite of the Pensamiento Complex which forms a large part of the Parágua Craton of Litherland *et al.* (1986). The terminal magmatic episode is represented by the layered cross-bedded nordmarkite of the El Tigre Alkaline Complex.

The granite series are dominantly potassic calc-alkaline granite, but some are of sub-alkaline composition. The dominant rock-types are monzo and syenogranite, gabbro and diorite. Sub-volcanic granophyre complexes and the layered cross-bedded nordmarkite of the El Tigre Alkaline Complex are observed in the southern zone of Bolivia. Locally, the post-tectonic granite exhibit rapakivi textures (*i.e.*, the Discordancia Granite). The Pensamiento Complex (Litherland *et al.*, 1989) consists of a number of dominantly syenogranite to monzogranite bodies of different ages. The few Rb/Sr isochron ages available for the granitoid plutons range between 1.391 and 1.291 Ga, and the initial Sr isotope composition, varies between 0.7003 and 0.7058 (Table 9). According to Litherland *et al.* (1989) the granitic bodies are syn to post-tectonic. The late to post-tectonic suites are deep level granitoid plutons derived from a mixing between lower crust and mantle sources with a prevailing mantle component. The emplacement took place during a period of stress relaxion, uplift erosion and decline in crustal temperature. The syn-kinematic granite bodies show a mixed crust-mantle isotope signature, which suggests generation mainly from short-lived sialic material.

The magmatic events ascribed to the Rondonian event in northern Rondônia, particularly in the Rondônia Tin Province, comprise intermittent distint bimodal intraplate rapakivi suites, which intruded the *c.* 1.75 - 1.53 Ga Rio Negro/Juruena crust. The episodes are defined by well-constrained U/Pb geochronology between 1.41 and 1.3 Ga (Table 10). This magmatism is represented by the suites as follows: the Santo Antonio-Teotônio (1.406 ± 32 Ga), the Alto Candeias (1.347 to 1.338 Ga) and the São Lourenço-Caripunas (1.314 to 1.309 Ga) rapakivi suites (Bettencourt *et al.*, 1999). The Igarapé Preto Intrusive Suite gave Rb/Sr age of 1.195 ± 0.05 Ga (Tassinari *et al.*, 1984), and could be included in the same event, considering that the U/Pb ages reported for similar rapakivi granite intrusives are consistently older than the Rb/Sr isochron ages of about 100 Ma (Bettencourt *et al.*, 1999).

Charnockite and mafic or ultramafic igneous complexes are associated with the Alto Candeias and São Lourenço-Caripunas suites. Contemporaneous diabase dykes are associated with the Santo Antonio rapakivi granite, whereas charnockitic and syenitic rocks of unknown ages are spatially related to the Alto Candeias rapakivi granites. Diabase, gabbro and anorthosite (Ciriquiqui basic and ultrabasic rocks) with conventional Rb/Sr age of *c.* 1.3 Ga crop out near the São Lourenço-Caripunas Batholith. The São Lourenço-Caripunas rapakivi granite and Ciriquiqui basic and ultrabasic rocks also constitute a large AMCG association. According to Bettencourt *et al.* (1999) there is the possibility that the suite, as defined herein, might consist of several sub-suites, each of which was emplaced over a short time.

Fazenda Reunidas Domain

In the western part of the Rio Alegre Terrane (along the Alegre River and the Aguapeí River), there occur granitoid plutons having a policyclic evolution ascribed to the Fazenda Reunidas Domain. The Aguapeí flat-lying sedimentary rocks bound these rocks to the W. Volcanic rocks of the Rio Alegre Volcano-sedimentary Sequence occur to the E and the Pantanal (Quaternary) sedimentary cover bound these rocks to the S.

In the southern part of Rio Alegre Valley the rocks comprise foliated grey granite denominated the Lajes Granite. Zircon U/Pb results (Table 11) (Geraldes, 2000) revealed a first age group around 1.606 Ga and a second at 1.31 Ga. The Sm/Nd mantle-depleted model age calculated for these rocks is 1.69 Ga, with $\varepsilon_{Nd(1.6\,Ga)} = +3.4$ and $\varepsilon_{Nd(1.3\,Ga)} = 0.0$. The older zircon U/Pb age was interpreted by Geraldes (2000) as the crystallization age, whereas the younger age was considered as isotope resetting in the U/Pb systems due to the superimposed magmatic or metamorphic event.

Tonalitic rocks were observed by Pinho (1990) along the Aguapeí River, near the Santa Barbara Ridge (Fazenda Reunidas Domain). They are associated with amphibolitic and intrusive monzosyenite, which has, in some places, a gradational contact with foliated tonalite, suggesting local melts. Geraldes (2000) reported concordant U/Pb ages for these tonalite bodies close to the contact with the monzosyenite, with values varying from 1.408 to 1.463 Ga. The display of these points along the concordia line suggests an isotopic resetting, probably due to the partial melt of the tonalite resulting in the monzosyenitic liquid with new U/Pb isotopic composition. The $\varepsilon_{Nd(1.4\,Ga)}$ value is + 3.6, which suggests a juvenile magma source for the tonalite. Other associated granitoid rocks presented similar geochronological results such as U/Pb ages of 1.412 ± 0.021 Ga and 1.4 ± 0.021 Ga with respectively (T_{DM} = 1.58 Ga and $\varepsilon_{Nd(T)}$ = 3.6) and (T_{DM} = 1.49 Ga and $\varepsilon_{Nd(T)}$ = 4.2).

The policyclic granitoid plutons from the Fazenda Reunidas Domain are younger than those of the Rio Alegre Terrane. However, the correlation between these associations is not well defined, and only further work will allow us to decipher the real tectonic meaning of these rocks.

Sunsas Province (SP)

The Sunsas Province is the youngest tectonic unit of the Amazonian Craton. It is best exposed in the southwestern part of the craton, where it consists of a zone with rocks generated by the erosion of older continental crust; the deposition and subsequent deformation and metamorphism of these sediments and older basement between 1.30 to 1.0 Ga (Litherland *et al.*, 1986). This metamorphic episode is associated with syn-tectonic magmatism. The 1.18 Ga Garzon Granulitic Belt, which occurs in Colombian Andes as a basement tectonic window, described by Kroonenberg *et al.* (1982) along with other Andean inliers (Santa Marta, Medellin, Arequipa/Antofala) (Tosdal, 1996), could be a possible extension of the Sunsas Belt towards the NW. The Tucavaca Belt (Lower Paleozoic) borders the Sunsas Province to the S, to the N by the older Rio Negro-Juruena Province, and to the W it extends under Phanerozoic sedimentary cover.

The start of the Sunsas tectonic evolution, which has been chrono-correlated with the Grenville Orogenic Cycle (1.3 to 1.0 Ga) in Laurentia and Baltica, was marked by an important phase of continental distension (rifting). Basaltic magmatism and the deposition in a continental margin environment of the sediments of Sunsas and Vibosi groups in Bolivia and the Aguapeí Group in Brazil represent this distensive phase. Furthermore, this basin was closed during the development of the Sunsas/Aguapeí Orogenic Belt. During the same orogenic episode there occurred the evolution of the Nova Brasilândia Volcano-plutonic Sedimentary Sequence and syn-tectonic magmatism. These episodes were followed by deformation and alkaline plutonism associated with the final stage of development of the Sunsas Orogen. Uplift and regional cooling occurred after *c.* 920 Ma, when cratonization was gradually achieved.

The Sunsas Geochronological Province may be subdivided into three main lithotectonic segments as follows: (i) the Sunsas Mobile Belt in Bolivia (Litherland and Bloomfield, 1981); (ii) The Aguapeí Thrust Belt in Brazil (Saes and Fragoso Cesar, 1996); and (iii) The Nova Brasilândia Metavolcano- sedimentary Sequence (Rizzoto, 1999).

The Sunsas Mobile Belt in Bolivia

The Sunsas Cycle was marked by an important event involving continental distension that was followed by alkaline plutonism and by the deposition of the Sunsas and Vibosi groups (constrained by Litherland *et al.*, 1986, at about 1.3 Ga - 950 Ma). This probably represents an extension of the same basin of the Aguapeí Group in Brazil (Saes, 1999).

The term Sunsas Orogeny was introduced by Litherland and Bloomfield (1981) to designate a period of sedimentation that included the erosion, deformation and metamorphism and reworking of the basement rocks and magmatism within the interval 1.3 Ga to 950 Ma. These geological events that took place in the southwestern part of the Amazonian Craton form the southern rim of the so-called Parágua Craton of Litherland *et al.* (1986). The deformation and metamorphism resulted in a WNW trending Sunsas Mobile Belt (Litherland *et al.*, 1989). The following summary is totally based on Litherland *et al.* (1986)

The Sunsas Mobile Belt (1.25 Ga - 900 Ma) in Bolivia is represented by reactivated basement, syn and post-tectonic granitoid, and sparse outcrops of metasedimentary rocks. The Sunsas deposits belong to two distinct environments: mobile belt and cratonic. Penetration-*deformed beds partially surround the so-called Parágua* Craton along its southern and eastern margins.

The geochronological database for this orogeny in Bolivia was summarized by Litherland *et al.* (1986). It is based mainly on Rb/Sr and K/Ar ages, and concludes that the metamorphism, deformation and plutonism of this belt extended from approximately 1.28 Ga to 950 Ma. The metamorphic basement and the San Ignacio Granitoid were reworked at this time as seen by the Sunsas tectono-metamorphic overprint, producing K/Ar age resetting.

The western and southern limits of the Sunsas Mobile *Belt are defined by the Rio Negro Front and the Santa* Catalina Straight Zone (Fig. 7). Both comprise a NW trending series of lineaments, which on the ground coincide with *en-echelon* mylonite belts. Z-shaped folds with limb sliding along short limbs producing minor dextral displacements associated with granulite downgrading. Whilst the Santa Catalina structure is clearly sinistral, the Rio Negro Front appears to show a regional dextral shift. The Granulite Complex to the S shows a sinistral swing.

The southeastern part of the Sunsas Belt can be divided into two tectonic segments. (1) N of Conception; where the San Ignacio structures are preserved to a considerable extent despite Sunsas reworking. (2) The San Diablo Front area, which is described in the southernmost part of the Sunsas Orogen and was interpreted as a suture zone between the Parágua Craton and the San Pablo Terrane (Saes and Fragoso Cesar, 1996; Saes, 1999). According to these authors, the Sunsas Orogen was the result of a collision of the Parágua Craton and the San Pablo Terrane, and the collage zone is represented by the San Pablo Front (Litherland *et al.*, 1986). This collage may be responsible for the metamorphism and magmatism during the Sunsas Cycle, and the granodiorite intrusives observed in San Pablo Terrane are the arc-related result over a SSW dipping subduction zone. There is no geochronological data for San Pablo Granitoid to confirm this hypothesis.

Sedimentation

The sedimentary record of this cycle is represented by the Sunsas and Vibosi groups, dated about 1.35 Ga, that overlie rocks of the San Ignacio Orogen with a marked uncorfomity (Literland *et al.*, 1986). The Sunsas Group is composed of quartzite, sandstone, shale and oligomitic quartz conglomerate, ranging from 1000 to 6000 m in thickness. The group includes three units, which are from the base to the top, the Arco Iris (or Guapama, or El Pucho), Cuatro Carpas (or El Elucion, or Lower Psammitic unit), and Bela Vista (or Cabecera, or Argillaceous unit).

The Arco-Iris unit (in Sierras Huanchaca-Ricardo Franco) comprises oligomitic conglomerate with quartz pebbles in a phyllitic matrix. Clast-supported conglomerate with a quartzitic matrix, poorly sorted siltstone wacke, conglomerate with a ferruginous (hematitic) matrix. Mica-schist, granitoid rocks, black quartz and volcanic rocks mainly form clasts. The Cuatro Carpas unit (in Serrania Huachaca) consists of a varied facies association with channel structures and siltstone and mudflake breccia. Cross-bedding and graded bedding may be result of sorting by current action rather than by turbidity flows. The Lower *Psammitic unit (in Sierra Santo Corazon) comprises typical* red sandstone, fine to coarse-grained psammitic types, of which the angularity increases with the amount of feldspar (microcline-dominated sub-arkose and arkose). The Argillaceous unit (Sierra Santo Corazon) consists of pelitic rock-types rarely found as massive units. They are normally intercalated with siltstone and psammite in semi-pelitic sequences, which are distinctly rhythmic in parts of Sierra Huanchaca.

Ripple marks, tabular-planar and festooned cross-bedding can be observed in the Arcos Iris, and Cuatro *Carpas units. Red mudstone beds, indicating oxidation and* dissecation (and hence deposition under subaerial conditions), are typical of shallow marine deposition. The Lower Psammitic unit consists of sandstone with tabular cross-bedding deposited in braided rivers and fine-grained sandstone deposited in eolian dunes. Lithology (mudstone)

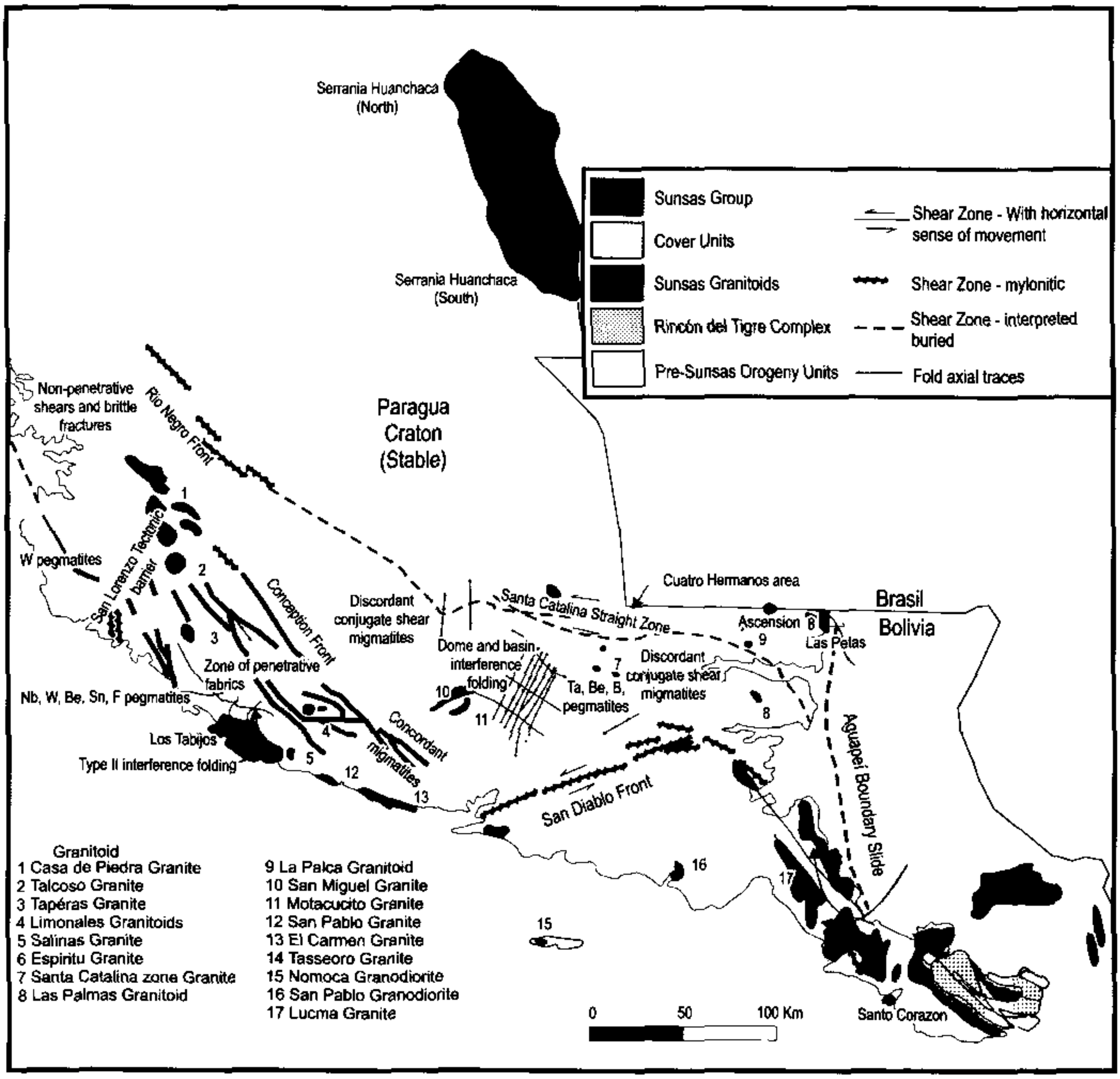


FIGURE 7 - Main units of the c. 1.0 Ga Sunsas Orogeny in eastern Bolivia (modified after Litherland et al., *1986).*

and sedimentary structures (rhythmic structures) in the Argillaceous unit are interpreted as result of deposition in deep marine water.

The Vibosi Group is a sequence about 2000 m thick of sandstone and arkose. Litherland *et al.* (1986) correlated the Vibosi Group with the Upper Psammitic unit and, consequently, included the Vibosi Group in the Sunsas Cycle. The basal Santa Isabel Formation (1100 m) is composed of medium-grained arkose. These beds are overlapped by the San Marcos Formation (750 m) with interbedded purple sub-arkose and fine-grained reddish-brown quartz sandstone. The Santo Columbo Formation (450 m), at the top of the sequence, comprises fine-grained, grey, and locally cross-bedded quartz sandstone.

The Magmatism

The Sunsas Cycle magmatic activity in Bolivia occurred between 1.005 Ga to 993 Ma (considering Rb/Sr isochron age) and is represented by the Rincon del Tigre Igneous Complex and several kinds of granitoid and migmatite. The Rincon del Tigre Igneous Complex occurs over an area of about 720 km², being a layered, thick, differentiated, mafic/ultramafic sill discovered by Hess (1960), which cuts sedimentary rocks of the Sunsas and Vibosi groups. The complex is subdivided in three units: lower ultramafic unit, middle mafic unit and upper felsic unit (granophyre). Darbyshire (1979) reported a Rb/Sr age of 992 ± 86 Ma for these rocks.

According to Litherland *et al.* (1986) the Sunsas Ganitoid includes both syn to late-kinematic bodies related to shear zones and late to post-kinematic types with little or no signs of deformation. The syn to late-kynematic types, may be foliated locally and consist mainly of medium to coarse-grained biotite or biotite muscovite syenogranite. The granitoid may show gradational contacts with nebulitic Sunsas gneiss-migmatite rocks. The main plutonic bodies of the Sunsas Province are: Espiritu Granite, Santa Catalina Zone Granite, Las Palmas Granitoid, La Palca Granitoid, San

Miguel Granite, Motacucito Granite, San Pablo Granite, El Carmen Granite, and Nomoca Granodiorite (Table 12).

The late to post-tectonic bodies are aligned within the Sunsas Orogen across the San Diablo Shear Zone and are represented by the Casa de Piedra, Talcoso, Tapeva, Salinas and Tasseoro granite bodies, and also by the San Pablo Granodiorite, and Luma Granite in the southeast fringe. They are circular or oval in plan view with intrusive contacts and little or no signs of internal foliation. They are dominantly medium to coarse-grained biotite monzogranite. Additionally, marginal swarms of crosscutting pegmatite and aplite veins are observed in the country rocks. The observed field features indicate that the granite emplacement occurred at a higher crustal level. The age determinations are still fragmentary, and only the Casa de Piedra Granite was dated by Rb/Sr at 1.005 ± 0.012 Ga. According to Teixeira and Tassinari (1984) this age is in good agreement with those already observed for the Young Rondônia Granite and Costa Marques Group in Brazil.

The Sunsas Migmatite occurs along sets of vertical conjugate shears. The migmatite is poorly developed in granoblastic granulite and on the Concepcion Front. These rocks are concordant with a new penetrative D_3 shear zone fabric. As a whole the paragneiss was migmatized in the upper amphibolite facies affecting the Rondonian-San Ignacio basement.

The Aguapeí Thrust Belt

The Aguapeí Thrust Belt occurs in the southwestern part of the Amazonian Craton in Brazil and comprises a sedimentary group deformed in a cratogenic environment. This unit partially covers the vulcano-plutonic basement ascribed to Rio Negro-Juruena and Rondonian-San Ignacio Province. The sedimentary sequence was described as Aguapeí unit (Fig. 8) by Figueiredo *et al.* (1974). Souza and Hildred (1980), Saes *et al.* (1992) and Menezes *et al.* (1993) who proposed the elevation of this unit to group status describing three formations: Fortuna, Vale da Promissão, and Morro Cristalina. Litherland *et al.* (1986, 1989) suggested the correlation between Aguapeí and Sunsas groups, indicating the link between Sunsas and Aguapeí basins and the contemporaneity of the respective metamorphic episodes.

The development of the Sunsas-Aguapeí sedimentary basin was controlled by the Jauru Terrane to the E in Brazil, and by the Parágua Craton of Litherland *et al.* (1986) to the W in Bolivia. The southern limit is unknown, and the basin probably extends northwards, encompassing the sedimentary rocks deposited in the Pacaás Novos, Uopiane, São Lourenço and Pimenta Bueno grabens. Deformational features were described by Saes *et al.* (1992); Menezes *et al.* (1993) and Geraldes (1996), suggesting the existence of a *c.* 950 Ma linear NNW mobile belt with different levels of deformation. Deformation included thrusting, folding and the development of shear zones correlated with the Grenville collage that resulted in the formation of the Rodinia supercontinent.

Sedimentary Units

The location of the Aguapeí Group in southwestern part of the Amazonian Craton is shown in Figure 8. In Bolivia the sedimentary rocks are observed at Serrania de Huanchaca and Santo Corazon (Litherland *et al.*, 1989), and in Brazil in the Ricardo Franco, São Vicente, Santa Barbara and Rio Branco ranges. According to Souza and Hildred (1980) and Saes *et al.* (1992), the Aguapeí Group includes three formations: a sandstone and conglomerate unit (Fortuna Formation); an intermediate pelitic unit (Vale da Promissão Formation) and an upper sandstone unit (Morro Cristalino Formation).

The sequence of the Aguapeí Group (in Brazil) and Sunsas Group (in Bolivia) record a complete cratonic oscillation, with a (1) transgressive phase with tide-dominated deposition of sandstone and conglomerate; (2) a marine progradation phase allowing the psammitic deposition in an oceanic, current-dominated environment; (3) an upper unit recording a marine regression with deposition of fluvial sandstone beds.

The above-described evolution allows the interpretation of the Sunsas-Aguapeí Basin as a depression originated by crustal extension within the Amazonian protocraton. Both the immature nature and anomalous thickness of the Santo Corazon area suggest the NNW trend of older basement structures which acted as paleosutures for rifting that originated the Sunsas-Aguapeí Basin. The Aguapeí Thrust Belt is characterized by a deformational event that thrust and folded the sedimentary rocks of the Aguapeí Group (Saes, 1999).

The K/Ar dating of hydrothermal seriates in veins in gold deposits associated with the Aguapeí sequence showed ages in the range 960 to 840 Ma, which may indicate the time of the hydrothermal sericitization. Pb/Pb dating in galena yielded model ages in the range 1.0 Ga to 800 Ma for the Onça Deposit, in agreement with K/Ar ages reported by Geraldes *et al.* (1997). These geochronological results suggest minimum ages for the Aguapeí Group.

Magmatism

The Guapé Intrusive Suite was first described by Barros *et al.* (1982), and also studied by Saes *et al.* (1984) and Menezes *et al.* (1993). The first-cited authors reported a Rb/Sr (whole-rock isochron) age of 950 ± 40 Ma. The last-cited authors carried out chemical studies and described bimodal anorogenic characteristics for this unit. Geraldes (2000) reported a T_{DM} age of 1.29 Ga and $\varepsilon_{Nd(950\,Ma)}$ value of + 1.27, suggesting a contribution of juvenile and crustal material for the magma genesis. The same author also reported two U/Pb zircon ages for the São Domingos Suite of 939 ± 19 Ma (T_{DM} = 2.21 and $\varepsilon_{Nd(T)}$ = - 7.1) and 914 ± 14 Ma (T_{DM} = 2.21 and $\varepsilon_{Nd(T)}$ = 7.6). This unit was interpreted as a shear-controlled S-type granitoid generated at the end of the Aguapeí thrust event. Sato and Tassinari (1997) found rocks with T_{DM} ages of *c.* 1.15 Ga in the area.

The Nova Brasilândia Terrane

The summary given below draws on published work and almost totally on Rizzoto (1999), Rizzoto *et al.* (1999a, b), to whom the reader is referred for detailed descriptions. The name Nova Brasilândia Metavolcano-sedimentary Sequence was introduced by Silva *et al.* (1992) for a group of supracrustal rocks. These rocks consist of amphibolite facies metamorphites as represented by mica-quartz schist, biotite paragneiss, calc-silicate rocks, and amphibolite that occur in the Nova Brasilândia d'Oeste and Alta Floresta d'Oeste regions (southeastern Rondônia), formerly recognized and named the Comemoração Epimetamorphites by Leal *et al.* (1978). More recently, and well documented by Rizzoto (1999) the entire sequence was redefined as a distinct tectonic unit the limits of which extend to northeastern and southeastern regions of the State of Rondônia.

Based on recent geological studies and U/Pb (SHRIMP) geochronological data, Rizzoto (1999) proposed the name Nova Brasilândia Group to embrace the main lithostratigraphical unit composed by dominantly mafic rocks (metagabbro, metadiabase and amphibolite) and by a metaplutonic-sedimentary sequence (biotite-feldspar quartz gneiss, mica schist and calc-silicate rock). These rocks are intruded by several high-level late-tectonic A-type granite plutons. However, the formal name Nova Brasilândia Terrane was proposed by Scandolara *et al.* (1999c) to replace the terms previously applied to these rocks.

Supracrustal rocks

In the type area the Nova Brasilândia Group was subdivided by Rizzoto (1999) into the Migrantinópolis Formation and Rio Branco Formation. The Migrantinópolis Formation consists of psammo-pellitic supracrustal rocks derived from deep-sea sediments, which consist of siliceous-clastic-carbonaceous turbiditic sediments that are intruded by subordinate sills, stocks and dykes of metagabbro-norite, amphibolite and metabasalt. The absolute age of supracrustal rocks remains unknown, but detrital zircon from a banded paragneiss indicate a U/Pb SHRIMP zircon age of 1.215 ± 0.02 Ga, which is interpreted as the maximum age of the deposition of the sediments. Inherited zircon having variable ages from 2.09 Ga to 1.417 Ga and Sm/Nd model ages for the banded paragneiss yielded values from 1.91 to 1.63 Ga, $\varepsilon_{Nd\,(T)}$ = - 3.8. These data support the assumption that a continental crust source was present on the northern flank of the Nova Brasilândia Group probably ascribed to granitoid plutons of the Serra da Providencia Intrusive Suite (1.57 to 1.53 Ga) and metadacite of the Roosevelt Sequence (1.75 Ga).

Sills and stocks of mafic meta-igneous rocks, metagabbro, metadiabase represent the Rio Branco Formation in addition to amphibolite interlayered with minor amounts of marl-turbidite (calc-silicate gneiss) and rare terrigenous silicic-clastic metaturbidite (quartz-feldspar gneiss and fine-grained mica-schist).

The 1.11 ± 0.01 Ga age for a metamorphosed (amphibolite facies) subophitic métagabbro is based on four zircon fractions, which was interpreted as the age of metamorphism that affected the sequence (Rizzoto, 1999). The metabasic rocks are characterized by $\varepsilon_{Nd(T)}$ varying from 3.1 to 5.0, which indicate a mantle source for these rocks, suggesting a juvenile-magma accretion to the continental crust at this time (Table 13). These data match the values and the interpretation previously advanced by Sato and Tassinari (1997). The geochemical and isotope data define the metabasic rocks as enriched mantle basalt of the tholeiitic series (P-MORB) and probably these sequences (turbidite and mafic magmatism) represent a passive rift environment.

Granitic Magmatism

The granitic intrusions in the Nova Brasilândia region took place in two distinct pulses at 1.098 ± 0.01 Ga and 995 ± 15 Ma, related to Rio Branco Granite and the Rio Pardo Granite Suite (Rizzoto *et al.*, 1999a, b). The Rio Branco Granite is mainly composed of a number of small syn-tectonic bodies interlayered with metabasite and calc-silicate gneiss. Monzogranite bodies have been strongly deformed and are controlled by the main transcurrent regional fault system. Samples of biotite monzonite gneiss have a U/Pb zircon crystallization age of 1.098 ± 0.01 Ga, which overlap with *peak-metamorphic age and deformation* of the Nova Brasilândia Group. The $\varepsilon_{Nd(T)}$ values of - 0.4 and the Sm/Nd model age (T_{DM}) of *c.* 1.63 Ga indicate that the original magma was derived from a source with an older crustal component and slight participation of mantle material (Table 13).

The Rio Pardo Granite Suite represents the younger tardi to post-tectonic granite pulse, which indicates the final pulse of the Sunsas Orogeny in the area. It is represented by oval shaped, epizonal plutons, locally grading into foliated charnockite. The granite intrusives are dominantly *hypersolvus syenogranite and subordinate subsolvus* monzonite showing rapakivi textures. A metamorphic, granoblastic texture is superimposed and the rock exhibits a protomylonitic texture. The granite intrusives are metaluminous, peraluminous, sub-alkaline A-types granite intrusives and have age counterparts in the Younger Rondônia Granite in northern Rondônia. Contemporaneous foliated coarse-grained grey to greenish charnockite bodies also occur.

Four zircon fractions of one porphyritic monzonite gave an upper intercept crystallization age of 995 ± 15 Ma. The *Sm/Nd model age around 1.5 Ga and* $\varepsilon_{Nd(T)}$ = + *0.5 suggest* that the granitoid plutons were derived from the recrystallization of an older continental crust with a short crustal life together with significant juvenile contribution. Five samples yielded a Rb/Sr isochron age of 1.003 ± 0.022 Ga and $^{87}Sr/^{86}Sr_{initial}$ = 0.7038 ± 0.0002 (MSDW = 0.3031), which is in close agreement with the new radiometric results.

Deformation and Metamorphism

It was postulated by Rizzoto *et al.* (1999a, b) that the Nova Brasilândia Group evolved during two periods of complex distension and convergent tectonic regimes. The model must consider two phases of deformation, as follows:

• An extensional tectonic setting: involving intracontinental rifting evolving to a progressive continental margin with the production of coeval 1.15 Ga mantle-derived tholeiitic melts (P-MORB type) represented by sills. The mafic rocks clearly precede the deformation and metamorphism that affected the foliated granitic rocks in the area.

• Later, during the compressional tectonic regime: the rocks of the Nova Brasilândia Group were affected by a severe transpressive regional metamorphic event in the upper amphibolite facies, under high-temperature and low pressure conditions, represented by the Rio Branco Transpressive Belt. A well-developed foliation and schistosity including widespread formation of mylonite accompanied this event by the development of transcurrent shear-zones.

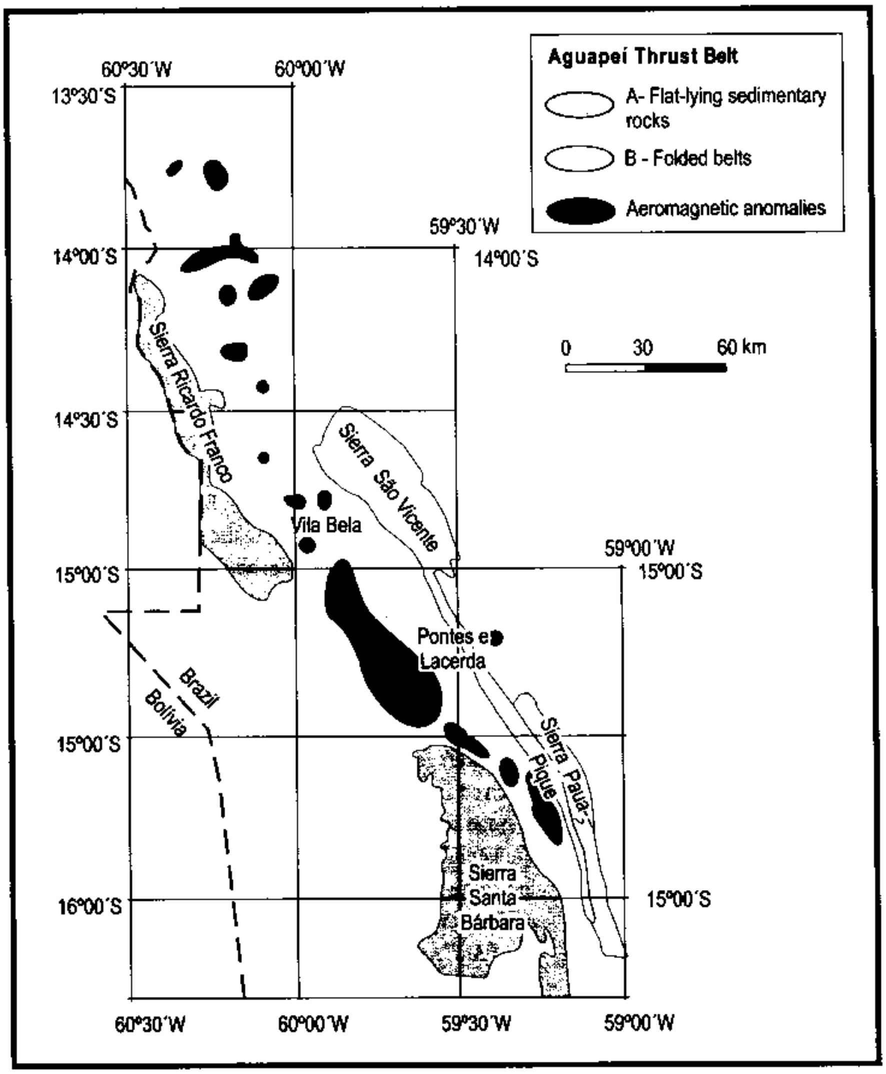
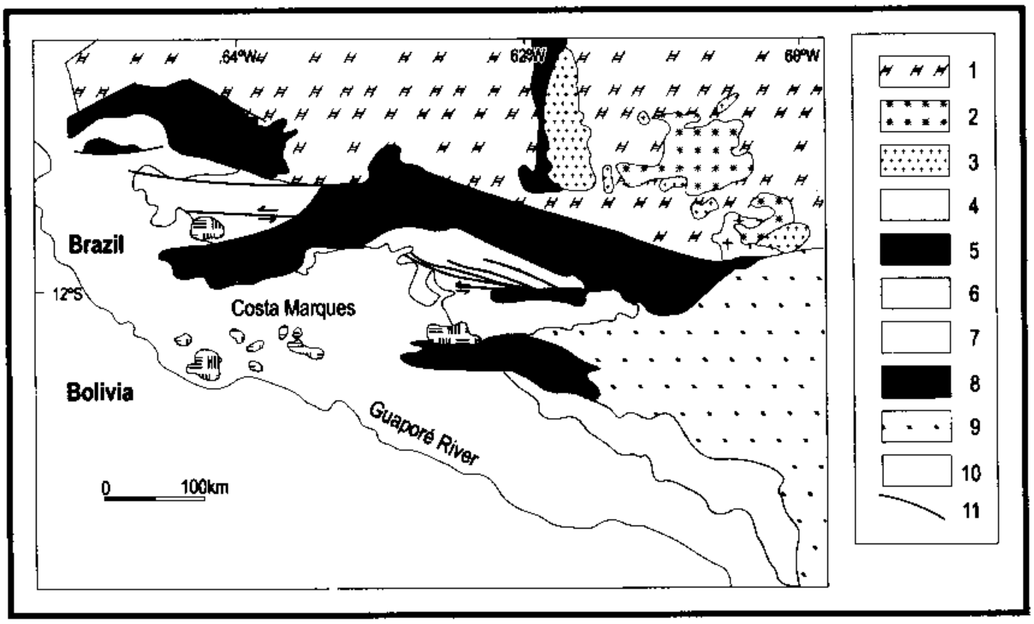


FIGURE 8 - Geographic distribuition of the Aguapeí Group morphostructures. The aeromagnetic anomalies delineate the cancealed Rio Alegre Terrane boundary (modified after Menezes et al.*, 1993 and Saes, 1999).*

FIGURE 9 - Major units of the Nova Brasilândia Group. 1 - Jamari Terrane; 2 - Roosevelt Terrane; 3 - Serra da Providência Intrusive Suite; 4 - Nova Brasilândia Volcano-sedimentary Sequence; 5 - Mafic rocks; 6 - Undeformed Sunsas Sedimentary Cover; (Tardi to post tectonic granitoids); 8 - Paleozoic Cover; 9 - Mesozoic Cover; 10 - Recent Cover; and 11 - Fault.



During this episode the mass transport was from SW toward NE. The anatectic syn-tectonic leucogranite resulting from partial melting of metaturbiditic sediments yielded a U/Pb age of 1.1 ± 0.008 Ga, which was interpreted by Rizzoto (1999) as dating the regional metamorphism in the area. This metamorphic episode also produced intense migmatization of the metaturbidite beds and recrystallization of the metabasic rocks.

Sunsas Effects in Northern Rondônia and Adjacent Areas

The effects of the Sunsas Orogeny in northern region of Rondônia (Fig. 10) and adjacent areas (states of Mato Grosso and Amazonianas, Brazil) occurred during the period between 1.15 Ga to 970 Ma. It includes a metamorphic overprint and deformation from 1.156 to 1.1 Ga and the emplacement of rapakivi granite intrusives, mafic dykes and granitic plutons between 1.08 Ga to 970 Ma in the older Rio Negro-Juruena and Rondonian-San Ignacio geochronological provinces. This superimposed metamorphic episode was previously recognized by Amaral (1974) as an autonomous reactivation on the ancient craton that he called the Rondoniense Reactivation (1.05 Ga - 900 Ma). These effects partially coincide in time with the Sunsas Orogenesis in Bolivia, the Aguapeí Thrust Belt and Nova Brasilândia Terrane, both in Brazil.

The Sunsas precursors in the region encompass the voluminous Rio Negro-Juruena (1.8 - 1.55 Ga) overprinted crust, mostly composed of paragneiss tonalitic gneiss, granitoid, migmatite and basic granulite, and rocks related to the Rondonian-San Ignacio event (1.41 - 1.39 Ga), already within the Rondonian-San Ignacio Province. These rocks show mainly bimodal mafic-felsic rapakivi magmatism with minor felsic gneiss. Sedimentary precursors are possibly represented by shallow marine sediments of the Beneficiente Group and by the Prosperança Formation.

The magmatism at 1.08 Ga - 997 Ma in northern Rondônia is composed of rapakivi granite and associated mafic rocks, including the Santa Clara Intrusive Suite (1.8 - 1.07 Ga) and Younger Rondônia Granite (1.0 Ga - 970 Ma). The granite bodies are mainly sub-alkaline, metaluminous to peraluminous, and show geochemical features of A-type within-plate granite (Leite Jr., 1996; Bettencourt *et al.*, 1997).

The Santa Clara Intrusive Suite includes the granite from the following massifs: Santa Clara, Oriente Velho, Oriente Novo, and Manteiga (Fig. 14). The older rock association is composed of porphyritic quartz-monzonite, monzonite and syenogranite with subordinated amounts of quartz-monzonite and minor pyterlite. Biotite and minor hornblende are the main mafic minerals and zircon, apatite, ilmenite, magnetite, allanite, fluorite and sphene are the essential minerals. A younger association includes syenite, trachyte and peraluminous and peralkaline granite.

The Younger Rondônia Granite intrusives were subdivided by Leite Jr. (1996) and Bettencourt *et al.* (1999) into two distinct suites. The first suite is composed of metaluminous to marginally peraluminous subsolvus and sub-alkaline types with minor associated quartz-syenite, quartz-monzonite, monzonite. The second suite, of restricted occurrence, shows a hypersolvus character as well as alkaline affinities. The field relationships suggest that the alkaline rocks are younger than the sub-alkaline types. The geographical distribution and the ages of the granitoid are shown in Figure 10 and Table 14, respectively.

The sub-alkaline suite consists of at least three distinct intrusive granite phases. Early intrusive bodies are composed mainly of coarse pyterlitic to porphyritic biotite syenogranite, late intrusive syenogranite and alkali-feldspar granite. The most recent intrusive rocks are rare, and are comprise mainly of (topaz), lithium-mica albite granite, and (topaz) quartz-feldspar porphyry. Primary tin and associated metal deposits are spatially related to granite of the two latter phases. The alkaline suite is composed of alkali-feldspar granite and peralkaline granite, alkali-feldspar syenite, trachyte and microsyenite and sub-alkaline quartz-feldspar porphyry and hybrid rocks (quartz microsyenite and quartz syenite). Biotite and more rarely sodic amphibole are the main mafic silicate minerals in the granite, but augite and/or hornblende are common in the syenite and microsyenite.

The Younger Rondônia Granite shows $\varepsilon_{Nd(T)}$ values of + 0.33 to - 3.25, T_{DM} between 1.66 to 1.73 Ga, initial Sr in the range of 0.707 to 0.709, $\delta^{18}O$= + 81 to 9.5 ‰ and has $^{206}Pb/^{204}Pb$ of 17.7 - 20.6 and $^{208}Pb/^{204}Pb$ of 37.3 - 43.2, indicating time-averaged Th/Pb > 4. Older crustal rocks are clearly involved in granite genesis. A source characterized by an average crustal to elevated Th/U ratio also contributed to the genesis of these granite intrusives. Oxygen isotopes indicate a calc-alkaline magma component or assimilation of high-level crustal material (Bettencourt *et al.*, 1999).

$^{40}Ar/^{39}Ar$ plateau ages on hornblende foliated granitic gneiss and augengneiss from the Ariquemes region (RNJP) provided by Bettencourt *et al.* (1996) defined ages of 1.156 ± 0.036 Ga and 1.149 ± 0.035 Ga, respectively. These data suggest a Sunsas metamorphic overprint. The progressively slightly younger dates obtained on the biotite, 1.001 ± 0.033 Ga and 912.8 ± 30.5 Ma, and more feldspar (antiphertite slow cooling rates during metamorphism and are consistent with K/Ar ages observed in the Younger Rondônia Granite (1.08 - 1.0 Ga).

On the whole, the $^{40}Ar/^{39}Ar$ dates of 1.15 Ga - 950 Ma define the approximate period of regional cooling of the Sunsas Orogeny in the area. This interval is also recorded in the adjacent Rondonian-San Ignacio Province, the Nova Brasilândia Terrane and in the Garzon-Santa Marta inliers in the Andean Belt in Colombia (Krooneberg, 1982). Regional metamorphic cooling at 1.1 Ga, or slightly after, is indicated over large areas of Grenville Province (Anderson *et al.*, 1999).

Tectonic setting

The now recognized 1.15 to 1.1 Ga evolutionary stage in the southwestern sector of the Amazonian Craton is coherent, and is demonstrated by proto-ocean mafic rocks and typical rift-related supracrustal rocks recognized in the Nova Brasilândia Terrane (Rizzoto *et al.*, 1999a). The 1.08 Ga - 970 Ma stage as represented by extensive rapakivi magmatism, mafic flows and plutonism, in like manner to the Grenville Province, is widely recognized and attributed to collisional orogenesis between Laurentia and Amazoniania. Sadowski and Bettencourt (1996) proposed that the spatial organization of Grenville-Sunsas age structure in the southwestern part of the Amazonian Craton is compatible with

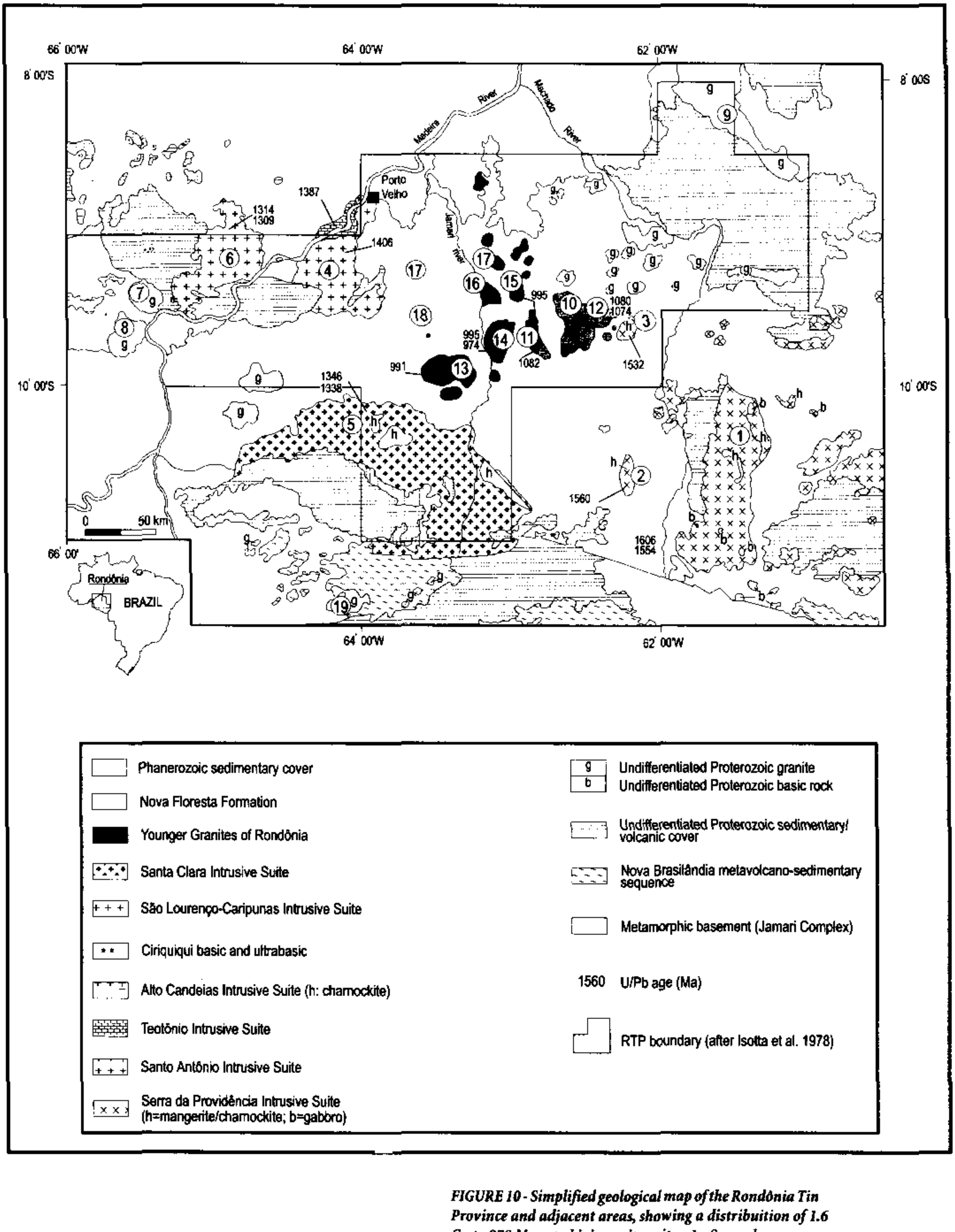


FIGURE 10 - Simplified geological map of the Rondônia Tin Province and adjacent areas, showing a distribuition of 1.6 Ga to 970 Ma rapakivi granite suites: 1 - Serra da Providência Batholith; 2 - Ouro Preto Charnockite; 3 - União Massif; 4 - Santo Antonio Batholith; 5 - Alto Candeias Batholith; 6 - São Lourenço-Caripunas Batholith; 7 - São Simão Massif; 8 - Abunã Massif; 9 - Igarapé Preto Massif; 10. Santa Clara Massif; 11. Manteiga Massif; 12. Oriente Novo Massif; 13 - Massangana Massif; 14 - São Carlos Massif; 15 - Pedra Branca Massif; 16 - Caritianas Massif; 17 - Santa Barbara Massif; 18 - Bom Futuro and Palanqueta hills; 19 - Costa Marques Group. Modified after Leal et al. *(1978), Schobbenhaus* et al. *(1981) and Bettencourt* et al. *(1999).*

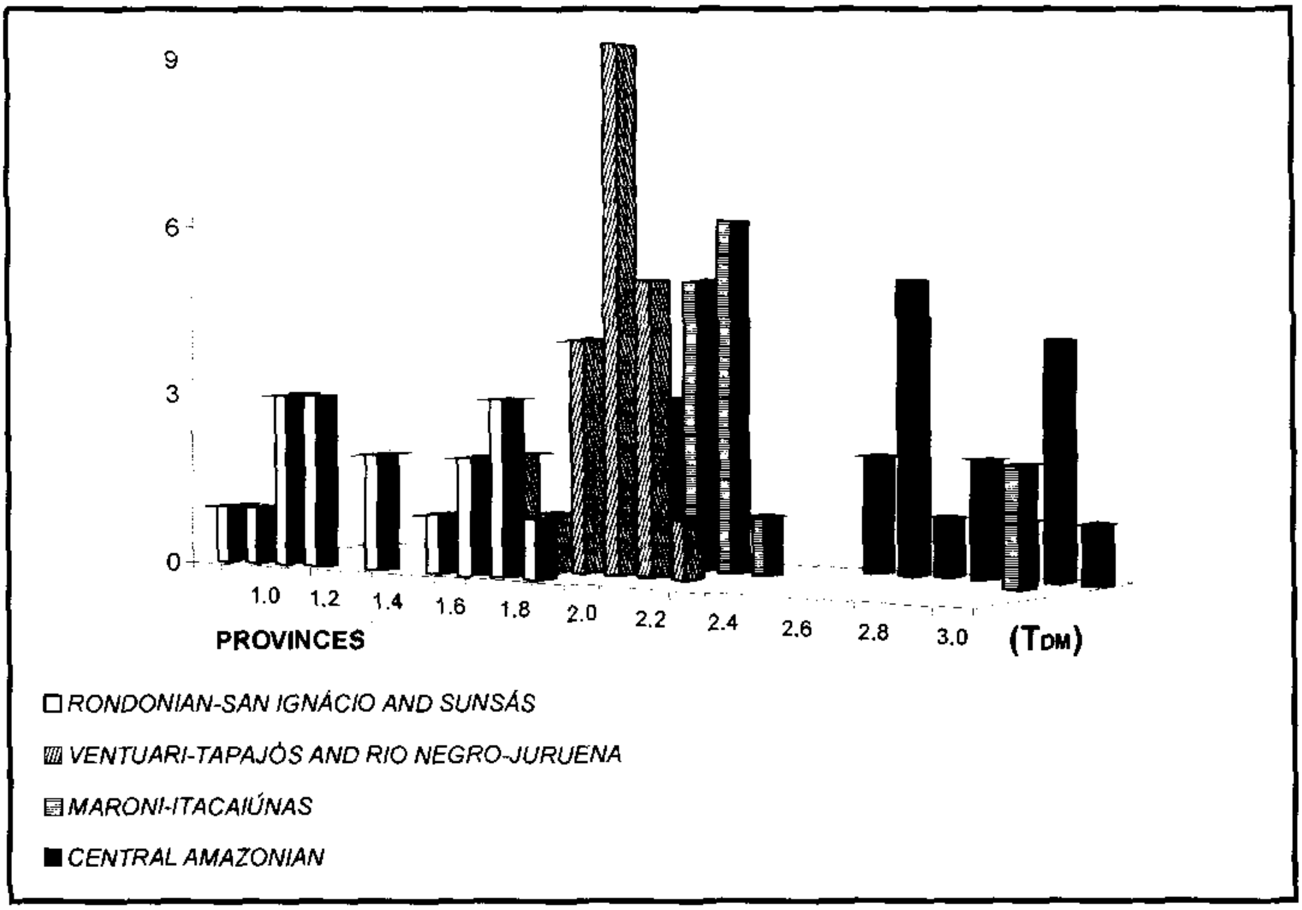


FIGURE 11 - Hystogram showing major accretionary events in the Amazonianian Craton according to depleted mantle Sm/Nd model ages.

a transpressional left-lateral component of collision, as mentioned by Hoffman (1992) and proposed locally by Gower (1996) in Canada. Following the collision, extensive strike-slip faults were accompanied by the intrusion of anorogenic granite and alkaline basalt that also resulted in the collapse of older cratonic sedimentary cover.

However a major extension-related underplating model as proposed by McLelland (1989) in terms of the thermal insulation below supercontinents, might also account from the proposed tectonic settings. It seems clear that a metamorphic event occurred in northern Rondônia between 1.156 Ga and 1.149 Ga, as indicated by the hornblende Ar/Ar data. This range of ages contrast with the time-span for the metamorphism of the Nova Brasilândia Terrane dated at 1.1 ± 0.008 Ga by Rizzoto (1999). As is the case in much of the Grenville Province these events, according to Gower (1996), show metamorphism between 1.01 Ga and 990 Ma. However, we predict that the northern Rondônia area could have been involved in a yet younger metamorphic event between 1.08 Ga to 970 Ma, but evidence for this remains to be confirmed.

Final considerations

Although the geochronological data for the Amazonian Craton is far from complete, the available geological and geochronological data show that the crustal evolution of the craton involved the significant addition of juvenile material during the Archean and Proterozoic, as well as the reworking of older continental crust. In this way we can estimate that about 30% of the continental crust of the entire cratonic area was derived from the mantle during the Archean and about 70% in the Proterozoic.

The distribution of the geochronological provinces of the Amazonian Craton suggests that the Archean protocraton was initially composed of independent microcontinents (Carajás-Iricoumé, Roraima, Imataca blocks, and the West Congo Craton, in Africa), which were amalgamated by Paleoproteorozic orogenic belts, between 2.2 and 1.95 Ga. The amalgamation of the Archean blocks that occurred at same time that juvenile Paleo-Mesoproterozoic continental crust began to be accreted to their western margins.

In this way, the Ventuari-Tapajós, Rio Negro-Juruena and part of the Rondonian-San Ignacio (Santa Helena Orogen) provinces represent a vast area of juvenile continental crust. These provinces were accreted to the protocraton between 1.95 and 1.45 Ga by successive magmatic arcs. It is likely that these arcs were the result of the subduction of oceanic lithosphere at the onset of the collision between a continental mass, composed at that time by the Central Amazonian and Maroni-Itacaiúnas provinces against another continental mass. This continental mass is now probably part of Rondonian-San Ignacio and Sunsas provinces and Laurentia-Baltica (Sadowski and Bettencourt, 1996; Tassinari *et al.*, 1996).

Between 1.4 and 1.0 Ga, as reported by Sadowski and Bettencourt (1996), the orogenic evolution of the Rondonian-San Ignacio and Sunsas Orogenies took place in ensialic environments, due to continental collision between the Amazonian Craton and Laurentia. Major mantle-crust differentiation in the Amazonian Craton took place at 3.1 - 2.8 Ga (CAP); 2.8 - 2.5 Ga (CAP); 2.2 - 2.0 Ga (MIP); 2.0 - 1.9 Ga (VTP and RNJP); 1.9 - 1.7 Ga, 1.6 - 1.5 Ga and 1.48 - 1.42 Ga (RSIP) and near 1.1 Ga (SP), as shown in Table 14. In Table 15 is shown the distribution of the Sm/Nd mantle-depleted model ages for the whole of the Amazonian Craton. These episodes, together with the 2.0 Ga to 900 Ma anorogenic magmatism related to rifting and continental breakup (basic-alkaline magmatism, A-type granitoid plutons, bimodal magmatism and the deposition of platform sediments, with decreasing ages from the NE towards the SW) are consistent with the hypothesis of lateral crustal growth during the Paleoproterozoic and Mesoproterozoic in the Amazonian Craton.

The Amazonian Craton in a Global Context

The 2.2 - 1.9 Ga geological evolution of the northeastern part of the Amazonian Craton is correlated with the West African Craton, whereas the southwestern part has a 1.75 Ga - 950 Ma tectonic history more compatible with Laurentia and Baltica. In this way, the main geological events that occurred in the Amazonian Craton can be chrono-correlated with cratonic areas.

The Northeastern Amazonian Craton

Various tectonic models have been proposed for the geodynamic evolution of the Maroni-Itacaiúnas Province and its relationship with the other domains of the Amazonian Craton. According to these models the Archean relics of the Imataca Complex are related to subduction with roughly N-S plate movements. Following minor Archean events, the major tectonothermal event occurred around 2.1 - 2.0 Ga (Transamazonian event) that is also related to subduction as manifest in the agglomeration of a succession of island arcs and back-arc basins thus explaining the production of juvenile material. These N-S oriented subduction processes led to formation of a mass of subducted sheets intruded by granitoid (granite greenstones terranes) responsible for the granulitic and amphibolitic metamorphism, as well as for the main orientation of the Central Guiana Granulitic Belt.

Ledru *et al.* (1994) and Vanderhaeghe *et al.* (1998) presented a model for the Transamazonian evolution of the Maroni-Itacaiúnas Province in northern French Guiana, where comparison with the crustal evolution of the West African Craton is emphasized. Two main stages of evolution have been defined. The first event corresponds to an early Transamazonian (*i.e.*, 2.2 - 2.14 Ga) major period of juvenile crust generation, characterized by oceanic crust subduction and continental accretion (formation of the granite-greenstones complexes).

The second stage is characterized by collision tectonics and crustal anatexis at about 2.09 - 2.08 Ga. This model is supported by detailed geochemistry, structural geology and geochronological studies, but the extension of this view to the whole MIP would be premature as it does not take into account the relationship with high-grade belts and reworked Archean nuclei. Delor *et al.* (1998) presented a model for the same region in which mantle plume processes are emphasized for the formation of juvenile crust, and the importance of modern collisional hypotheses for tectonic evolution are minimized.

For the entire Maroni-Itacaiúnas Province it is possible to define two main rock-forming events, the older event occurring between 2.26 and 2.11 Ga, and the younger event occurring between 2.0 and 1.86 Ga. The first period is characterized by the development of the granite-greenstone terranes, whereas the interval between 2.0 and 1.86 Ga is related to the formation of the Central Guyana Granulitic Belt and the gneiss-migmatite terranes. Within the period between 2.26 and 1.9 Ga there predominated continental accretion by mantle-derived material. During the period from 1.9 to1.86 Ga the crustal reworking processes appear to have been the most important of geological events.

In the West African Craton the Transamazonian Orogenic Cycle, also named the Ebournean Orogeny, comprises large granite-gneiss terranes and 2.25 - 1.95 Ga volcanic and sedimentary sequences. Between 2.1 and 2.0 Ga these sequences were affected by the Ebournean Orogeny. The emplacement of the granitoid plutons took place during three different periods: 2.18 - 2.17 Ga; 2.12 - 2.08 Ga and 2.0 - 1.95 Ga (Oberthur *et al.*, 1994), which were chrono-correlated with the granitoid plutons in French Guiana (Milesi *et al.*, 1995). In the West African Craton there are two volcanic episodes, the older episode dated at 2.19 - 2.17 Ga is characterized by tholeiitic magma. The younger episode, characterized by volcanism, occurred at 2.08 Ga and is composed of bimodal volcanic rocks (Delor *et al.*, 1992; Oberthur *et al.*, 1994; Milesi *et al.*, 1995). The age of the volcanism within the metavolcano-sedimentary sequences of the Maroni-Itacaiúnas Province is poorly constrained, but the available data, especially from French Guiana, suggest an age around 1.1 Ga (Gruau *et al.*, 1985).

The southwestern Amazonian Craton

A tentative time chart of events based on the new U/Pb and smaller amounts of Rb/Sr ages is shown in Figure 12. It also shows the relationships between sedimentation, calc-alkaline and rapakivi magmatism, mafic intrusions, thermal-metamorphic imprints and deformation among the orogens and terranes of the southwestern sector of the Amazonian Craton. Since Baltica, Laurentia and Amazoniania seem to have belonged to the same margin of a super-continent, in our discussion we will compare and contrast some of the evolutionary aspects of the orogens.

Events between 1.79 - 1.53 Ga

Little is known about the environment that might define the transition from Ventuari-Tapajós and Rio Negro-Juruena orogenic events. Likewise, the evidence that can be obtained for the precursors and detrital materials, limited to recent detrital U/Pb zircon ages at 1.9 - 1.7 Ga for metasedimentary gneiss from the Rondônia Tin Province provided by Payolla *et al.* (1998).

Data show that at least two main orogenic events occurred between 1.80 and 1.70 Ga and between 1.57 and 1.53 Ga within the Rio Negro-Juruena geochronological province. Calc-alkaline arc-related plutonism and volcanism point to two groups of magmatic rocks providing strong evidence for subduction related juvenile crust growth along more than 2000 km flanking the closest western margin of the Ventuari-Tapajós Province.

The first calc-alkaline magmatism is interpreted as recording the time of accretion of the island arc rocks to the proto-Amazonian Craton over a subduction zone dipping NW. The rocks were derived from mantle protholith magma with ages ranging from 2.0 to 1.7 Ga, probably by the recycling of the earliest Ventuari/Tapajós crust (2.0 - 1.8 Ga).

Gower (1996) provides a summary of correlatable events in Laurentia where crust of similar age, restricted to the pre-Makkovikian-pre-Labradorian (1.79 - 1.71 Ga) rocks is found in the Mealy Montains Terrane, Goose Bay area. In the Makkovik Province and in the Ketilidian Mobile Belt of southern Greenland, rapakivi granite and calc-alkaline crust have been dated at 1.74 Ga and 1.75 Ga, respectively.

The orogenetic events extending from 1.79 to 1.75 Ga in the southwestern part of the Amazonian Craton partially overlap in time with events in the Baltic Shield. These events include the Gothian-Kongsberian orogenesis (1.75 - 1.55 Ga) and the intermittent 1.85 - 1.65 Ga Trans-Scadinavian Igneous Belt (TIB-1: *c.* 1.81 - 1.77 Ga and TIB-2: *c.* 1.72 - 1.69 Ga) magmatic episodes as well as with the early rapakivi magmatism in the western part of the shield (Finland and Russia) (Larsen and Berglund, 1992; Wikström, 1996, *apud* Åhäll and Gower, 1997 and Sudblad and Ahl, 1997).

Events between 1.7 - 1.6 Ga

To date there is no unequivocal evidence as to whether subduction was continuous from 1.79 Ga. However, tectono-magmatic events within the interval 1.7 - 1.6 Ga are not yet recorded or reliably demonstrated in the southwestern part of the Amazonian Craton. Moreover, after the accretion of the 1.79 - 1.7 Ga calc-alkaline plutonic and volcanic rocks underlying the Alto Jauru Greenstone Belt (southwestern Mato Grosso), the Juruena region, and the Jamari and Roosevelt terranes, there seems to have occurred a long period of stabilization between 1.7 and 1.6 Ga.

According to Åhäll *et al.* (1996) and Åhäll and Gower (1997) magmatic activity in Baltica between 1.69 and 1.58 Ga produced calc-alkaline juvenile crust (1.69 - 1.65 Ga), alkaline-calcic late TIB granitoid (TIB-3: 1.68 - 1.65 Ga), and coeval calc-alkaline rocks. Farther to the E, in Finland, Estonia and Latvia, rapakivi magmatism produced the Wiborg-Estonia Group (1.65 - 1.62 Ga) as well as massifs temporally linked with calc-alkaline and alkaline-calcic Gothian magmatism (Rämo and Haapala, 1995; Laitakari *et al.*, 1996; Rämo *et al.*, 1996).

In Laurentia calc-alkaline magmatism extended from 1.68 to 1.65 Ga which led Gower (1996) to suggest southward subduction related crustal growth under the early Labradorian Arc. In Baltica the progressive eastward shift of magma composition from calc-alkaline to alkaline-calcic and finally through rapakivi granites strongly support eastward subduction (Åhäll and Gower, 1997). Gothian subduction continued in Baltica producing continental margin calc-alkaline magmatism and arc-accretion between 1.62 and 1.59 Ga, whereas no such activity has been recognized in eastern Laurentia (Gower, 1996).

Events between 1.58 - 1.52 Ga

Calc-alkaline juvenile crust from the Cachoeirinha arc-system (1.57 - 1.53 Ga) seems to be a complex crustal collage and may represent the root of a juvenile magmatic arc. The onset of the associated Cachoeirinha Granite (1.522 ± 0.011 Ga), the Quatro Marcos Granite (1.533 ± 0.006 Ga) and probably the Serra da Providência Intrusive Suite (1.575 - 1.52 Ga) farther N, temporally linked to the calc-alkaline and alkaline-calcic magmatism, might be expressions of the eastern Cachoeirinha subduction.

The Serra da Providência Intrusive Suite started to emplace outboard, farther N, more likely in an anorogenic setting, into increasingly more stable continental crust of the Rio Negro/Juruena juvenile arc system (1.75 - 1.72 Ga). The speculation that was offered by Bettencourt *et al.* (1999) was that the Serra da Providência Intrusive Suite could not be clearly linked to either extension at the end of the Rio Negro-Juruena Orogeny or to the development of a neighbouring orogeny between 1.6 and 1.5 Ga is now believed to be more likely related to the Cachoeirinha Orogeny.

In Baltica, continent-continent collision occurred at *c.* 1.58 Ga, and eastward subduction was renewed (Åhäll and Gower, 1997). Evidence is given by calc-alkaline *c.* 1.55 Ga mafic-ultramafic tonalitic intrusions (northern Telemark) and *c.* 1.53 - 1.5 Ga rapakivi magmatism in central Sweden. These events are correlatable with the onset of the Cachoeirinha Orogen in southwestern Mato Grosso, and in Finland, by the Aland Riga Group (1.58 - 1.54 Ga) and the *c.* 1.56 - 1.54 Ga Salmi Group rapakivi plutons (Laitakari *et al.*, 1996; Rämö and Haapala, 1995). There is no indication of metamorphic or magmatic activity during the period 1.6 - 1.55 Ga in Laurentia and, for this reason, there might have occurred the development of a passive continental margin basin (Gower, 1996; Åhäll and Gower, 1997).

Events at c. 1.5 Ga

The Rio Alegre rocks may be interpreted as having originated at a mid-ocean ridge at *c.* 1.5 Ga (U/Pb zircon ages) that underwent metasomatism under seawater (cloritization and epidotization). Metamorphism under greenschist to lower amphibolite facies (biotite zone to garnet-kyanite zone) and transposition until mylonitization (NW foliation) were associated with the accretionary process of the proto-Amazonian Craton during Mesoproterozoic times. The geochemical data on the intrusive rocks suggest an evolution as a result of differentiation of tholeiitic magma. Sm/Nd mantle-depleted model ages on volcanic and intrusive rocks vary from 1.87 to 1.77 Ga and $\varepsilon_{Nd(T)}$ values from + 2.0 to + 2.8, suggesting a mantle-derived magma (Geraldes *et al.*, 1999).

Geological relationships allow us to suggest that the tectono-metamorphic evolution of the Rio Alegre Terrane might represent the suture zone recording the end of the ocean plate subduction which gave origin to the Santa Helena Magmatic Arc. According to this hypothesis the Santa

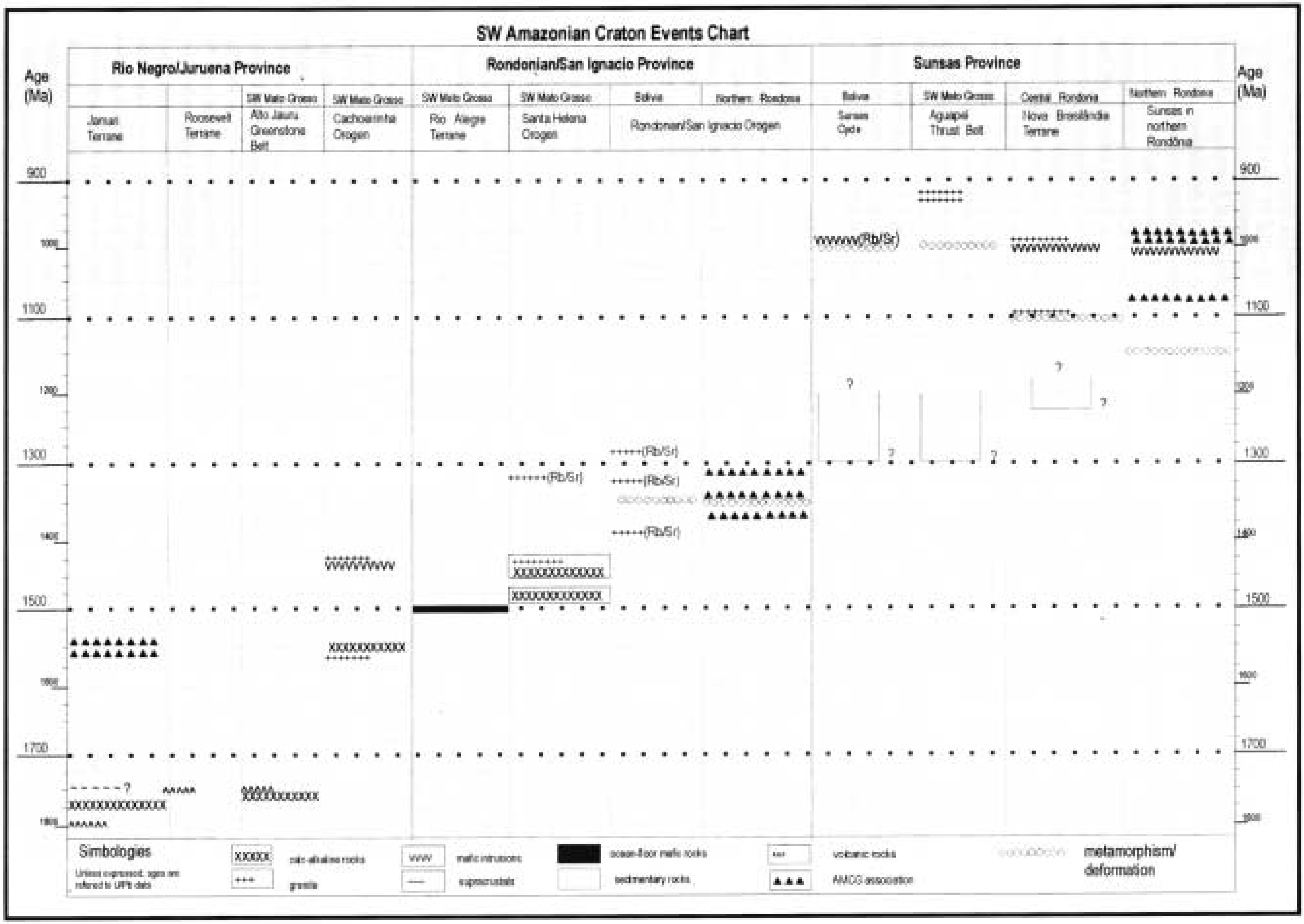


FIGURE 12 - A tentative time-correlation chart showing the orogenic events of the SW Amazonian Craton.

Helena Suite (U/Pb ages of 1.47 - 1.42 Ga and T_{DM} between 1.7 and 1.5 Ga) would have formed as a result of the ocean crust subduction, represented partially by the Rio Alegre Terrane. Detailed research on the Rio Alegre rocks should consider the possibility of an ophiolitic character for its genesis.

Sadowski and Bettencourt (1996) suggested that Amazoniania was also joined to Laurentia-Baltica at 1.6 Ga, and separated from it during Mesoproterozoic rifting known in Grenville Province as the Pinwarian Orogeny (1.51 - 1.45 Ga) that began a Wilson Cycle which probably ended with the formation of Rodinia (Gower and Tucker, 1994). This extensional event may be represented in the Amazonian Craton by the Rio Alegre Terrane, which leads one to suppose a possible link between Laurentia and Amazoniania pre-1.4 Ga.

Events between 1.45 - 1.3 Ga

A Mesoproterozoic history within this period is now classified into two stages; Santa Helena Orogen (1.48 - 1.42 Ga) and the Rondonian-San Ignacio Orogen (1.41 - 1.3 Ga). The Santa Helena Orogen (1.48 - 1.42 Ga) could be divided into a syn-accretion stage and a syn-collisional stage. The syn-accretion stage occurred at 1.48 Ga and generated calc-alkaline igneous and meta-igneous rocks including orthogneiss, tonalite and granite emplaced between *c.* 1.47 and 1.44 Ga. The syn-collision stage occurred between *c.* 1.42 and 1.4 Ga and consisted of period of mylonitization and migmatization at *c.* 1.42 to 1.4 Ga. There followed by a pulse of undeformed granite intrusion at 1.47 Ga. These were products of cratonic manifestation associated with the development of the Santa Helena Arc.

Orogenesis continued until the formation of the Alto Jauru Greenstone Belt, and the final bimodal phase of the Rio Branco Suite farther to the E between *c.* 1.456 and 1.423 Ga, which might be considered as back-arc granitoid magmatism outboard from the subduction zone.

According to Geraldes *et al.* (1997, 1999) and Van Schmus *et al.* (1998) the rocks are components of a *c.* 1.45 Ga, NW trending volcano-plutonic arc, that developed over an eastward dipping subduction zone along the western margin of a continent. This arc consisted of 1.8 to 1.7 Ga units of the Alto Jauru Greenstone Belt and the 1.57 - 1.53 Ga Cachoeirinha Orogen. The Sm/Nd data suggest that part of the orogen represented juvenile crust and may also have included a significant contribution from older crust.

The accretion of the Santa Helena Arc (*c.* 1.45 Ga) in the southwestern part of the Mato Grosso was followed in northern Rondônia by a prolonged period of voluminous anorogenic rapakivi plutonism, metamorphism, thermal effects and sedimentation. Geological rifting between 1.42 - 1.41 Ga to produce back-arc basins did not lead to crustal separation. The intrusions were not accompanied by continental margin processes, which suggest an intracratonic environment. Four distinct intraplate rapakivi magmatic age groups intruding the Rio Negro-Juruena (1.75 - 1.53 Ga) crust are revealed by the following intrusive suites: Santo Antônio (1.406 Ga), Teotônio (1.387 Ga), Alto Candeias (1.346 and 1.338 Ga), São Lourenço/Caripunas (1.314 and 1.309 Ga) and probably the Igarapé Preto Intrusive Suite (Rb/Sr age 1.195 Ga). Farther to the NW the Canama and Guariba alkaline plutons probably belong to this period.

In spite of contradictory opinions, it is reasonable to suppose that the 1.41 - 1.3 Ga intermitent intraplate bimodal magmatism accompanied by sedimentation in rift-related settings (Palmeiral and Prosperança sediments) could be regarded as the distal manifestation inboard of rifting and environmentally related to the development of the Santa Helena and San Ignacio orogenies.

The anorogenic episodes have age correlation in eastern Laurentia represented by the Michael Gabbro (1.426 Ga); Mealy dykes (1.38 Ga, Rb/Sr age); AMCG plutons of the Nain Plutonic Suite (1.35 - 1.29 Ga); the alkaline and peralkaline Red Wine Intrusive Suite (1.33 Ga), the genetically related Letitia Lake Formation and the mafic volcanism represented by the Seal Lake Group (1.25 - 1.23 Ga); and dyke emplacement which ended the Elsonian event (Gower, 1996). Coeval anorogenic magmatism was also on-going in Baltica, the eastern proto-margin of Amazoniania, at 1.41 Ga and 1.38 - 1.36 Ga expressed in the southwestern Scandinavian Domain by gabbro-anortosite-granite plutons and dolerite associations (Åhäll and Conelly, 1996).

Events between 1.29 - 1.19 Ga

Events between 1.29 - 1.25 Ga in the southwestern part of the Amazonian Craton are geochronologically poorly known, apart from the initial period of sedimentation. This period refers to the Sunsas and Aguapeí cycles, constrained at 1.215 ± 0.02 Ga (U/Pb SHRIMP detrital zircon maximum age for deposition of the sediments of the Nova Brasilândia Group) by Rizzoto *et al.*, (1999c).

This period overlaps with the Elsonian event (1.46 - 1.23 Ga) on the northern margin of the eastern part of the Grenville Province. It consists of mafic volcanism and dyke emplacement at 1.25 and 1.23 Ga, with the termination of anorogenic events in eastern Laurentia at 1.23 Ga. It coincides with the opening of the Elzevirian Orogenic Cycle (1.29 - 1.19 Ga; Rivers, 1997), with the sedimentation in the Sunsas-Aguapeí Basin, and with the deposition of sediments of the Nova Brasilândia Group.

Events between 1.19 - 1.089 Ga

This was a time of prevailing rifting activity leading to an extensional tectonic environment which permitted substantial sedimentation at *c.* 1.215 ± 0.02 Ga, emplacement of mafic magmatism during the initial extension stage and syn to late-kinematic granitoid emplacement in the compressional stage. Saes and Fragoso Cesar (1996) have considered the Sunsas Belt as resulting from a passive continental margin, and including deep marine turbidite sediments that have undergone subduction under the overthrusted plate. This has been named the San Pablo Terrane which supercedes the term San Pablo Granitoid (Litherland *et al.*, 1986) and it is related to a Sunsas Cycle calc-alkaline magmatic arc (Saes and Fragoso Cesar, 1996). The Aguapeí Aulacogen considered as an interior rift (Saes *et al.*, 1992; Saes, 1999) is represented by shallow marine shelf sediments, and represents the correlatable basin. Relevant units occur in the Nova Brasilândia Terrane and which are linked to the rifting stage include mantle-derived tholeiitic sills, stocks, gabbro and diabase dykes, emplaced at 1.15 Ga. This attests to widespread juvenile accretion (Rizzoto, 1999; Sato and Tassinari, 1997). A U/Pb zircon age of a metagabbro at 1.1 ± 0.015 Ga dates a metamorphic event of the mafic rocks, which is contemporaneous with turbidite beds and the emplacement pre-date the age of the metamorphism.

The Nova Brasilândia units provide an unequivocal link with the correlatable Sunsas Belt and the Aguapeí Thrust, and certainly represent the northwestern extension of the latter. The clastic rocks of the Sunsas-Aguapeí and Nova Brasilândia units represent a foreland assemblage partially backthrusted onto the Amazonian Craton. Suture rocks of coeval age were found in the Central Andes of Colombia and might be hidden below the Cisandine sediments of Bolivia (Sadowski and Bettencourt, 1996).

Granitoid plutons dated at 1.11 ± 0.008 Ga (anatetic leucogranite) as well the first 1.098 ± 0.01 Ga granitoid pulse associated with the compressive stage of the Nova Brasilândia Group are described. Isotope data for these rocks indicate derivation from an older crustal source (Rizzoto, 1999). A metamorphic event between 1.15 - 1.14 Ga (Ar/Ar age in hornblende and biotite) was defined by Bettencourt *et al.* (1996) in northern Rondônia (Jamari area), and another peak at 1.1 Ga (U/Pb SHRIMP zircon) was suggested by Rizzoto (1999) in the Nova Brasilândia Group. These favour the idea of dominance of two metamorphic pulses in the southwestern part to the Amazonian Craton during this interval.

The Sunsas Cycle was conventionally delimited by Litherland *et al.* (1986) between 1.35 - 1.0 Ga. Throughout much of the Grenville Province, the period between 1.3 - 1.0 Ga has been subdivided into the Elzeverian (1.23 - 1.18 Ga), Adirondian (1.18 - 1.08 Ga) and Grenville cycles (1.08 Ga - 970 Ma; Gower, 1996). More recently the Grenville Cycle was subdivided by Rivers (1997) into Elzeverian Orogenic Cycle (1.29 - 1.19 Ga) and the Grenville Orogeny, which comprises the terminal continent-continent collision and the respective pulses, which are the Sawinigan (1.19 - 1.14 Ga), Ottawan (1.08 - 1.02 Ga) and Rigolet (1.0 Ga - 980 Ma).

It seems that the events between 1.19 Ga - 980 Ma recorded in the southwestern part of the Amazonian Craton could be more appropriately compared to the Sveconorwegian (1.17 Ga - 950 Ma) and Adirondian (*c.* 1.18 - 1.08 Ga) extensively documented all over the Grenville Province and in southern Greenland (Gower, 1996). The Adirondian was accompanied by the emplacement of voluminous AMCG suites (1.17 and 1.12 Ga) and partly by compressional tectonic regime (Gower, 1996; Emslie and Hunt, 1990). This magmatic event overlaps with the time of emplacement of the mafic-felsic magmatism (1.15 and 1.098 Ga) and tectonics recorded in the Nova Brasilândia Terrane.

The remaining doubt is whether the 1.15 and 1.098 Ga mafic-felsic magmatism might suggest emplacement inboard of a continental margin over a northward dipping subduction zone and subsequent back-arc rifting or simply an expression of an intracontinental rifting and a proto-ocean expansion. Under these considerations it follows that the concept of the Sunsas Orogeny in the Amazonian Craton needs urgent re-evaluation, as already considered in respect to the Grenville Cycle in eastern Laurentia (Gower, 1996; Rivers, 1997).

Late Sunsas Events (1.08 Ga - 970 Ma)

Cratonic magmatism within this interval is characterized by the substantial addition of within-plate A-type felsic alkalic and mafic rocks dominantly in the Nova Brasilândia Terrane, Rondônia Tin Province, Bolivia and in scattered places in southwestern part of Mato Grosso. These rocks are represented by the Santa Clara Intrusive Suite (1.08 - 1.07 Ga); the Younger Rondônia Granite (1.0 Ga - 970 Ma); the associated Nova Floresta Formation consisting of alkali basalt and mafic dykes (K/Ar *c.* 1.0 Ga - 900 Ma); the Rio Pardo Granite Suite (995 ± 15 Ma), accompanied by foliated charnockite; the Costa Marques Group (Rb/Sr *c.* 1.018 Ga); the Guapé Intrusive Suite (Rb/Sr 950 ± 40 Ma); the late kinematic Sunsas Granitoid; the mafic-ultramafic Rincon del Tigre Complex (Rb/Sr 993 ± 139 Ma); and a number of pegmatites and basic rocks.

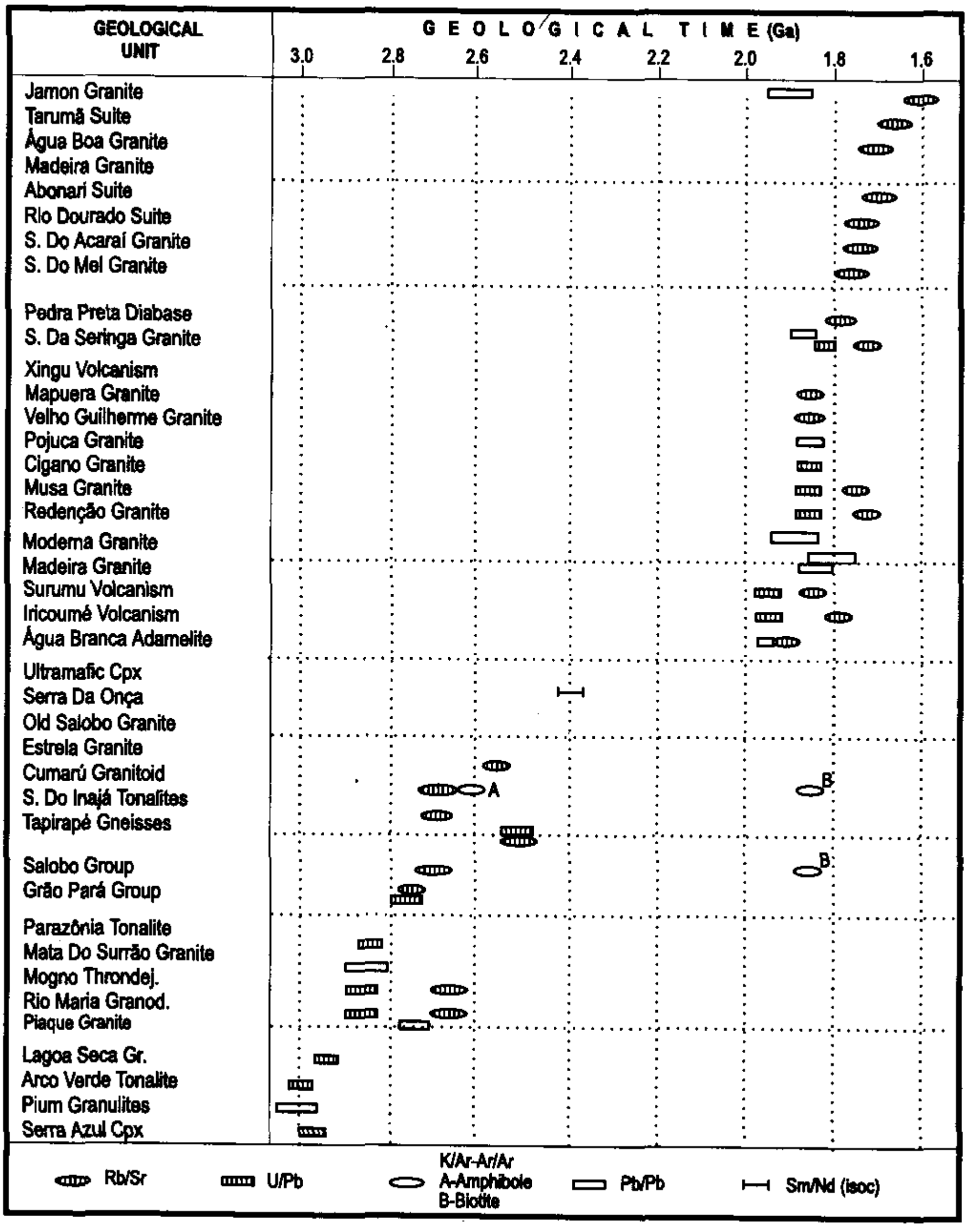
The magmatism is dominated by older sub-alkaline rapakivi granite intruded by younger alkaline rocks and deep source mafic magma. There also occurred the deposition of intracontinental rift sedimentary sequences, represented by the Palmeiral Formation, the Prosperança Formation; the Pacaás Novos, Uopiane and São Lourenço components, and the Sunsas Group (Leite Jr. *et al.*, 1996; Scandolara and Amorim, 1999) which indirectly attest to an extensional tectonic regime and rifting related to the final stage of the Sunsas Orogeny.

The rapakivi granite and associated mafic rocks in the Nova Brasilândia Terrane were most probably formed in a back-arc extension inboard from an active continental margin magmatic arc. The plutons in the Rondônia Tin Province developed in the foreland of the Sunsas Orogen probably involved extension and crustal thickening at *c.* 1.0 Ga followed by granite plutonism in the 990 - 960 Ma interval. This magmatism, deformation and thermal effects (high-grade ductile deformation and metamorphism which calls for re-evaluation) have been interpreted by Sadowski and Bettencourt (1996) as reflecting collisional tectonism, more precisely, continent-continent collision of left lateral transpressional character between Laurentia/Baltica and Amazoniania. The associated transtensional regime responsible for this magmatism also resulted in the collapse of the older cratonic sedimentary cover into the Uopiane, Pacaás Novos grabens in Brazil.

Magmatism associated with the same event in Laurentia has been widely attributed to collisional orogenesis between Laurentia and Amazoniania (Gower, 1996). According to Rivers (1997) the continent-continent Grenville Orogeny took place between *c.* 1.19 Ga and 980 Ma and comprised three distinct pulses of crustal shortening at *c.* 1.19 - 1.14 Ga, 1.08 - 1.02 Ga, and 1.0 Ga - 850 Ma, separated by periods of extension. The last pulse caused northwesterly propagation of the orogen into its foreland. The periods of crustal extension during the Grenville Orogeny were coeval with emplacement of mafic magma and anorthosite complexes implying that large quantities of mantle magma and heat had access to the base of the previously thickened orogenic crust. All these models are compatible with the late Sunsas geological evolution. Corrigan and Hanmer (1997) interpreted the within-plate AMCG-type plutonism in the Grenville Orogen, between 1.08 Ga and 850 Ma, to have formed in response to the delamination of the subcontinental lithosphere, injection of mantle magma and concomitant extension, following crustal thickening.

Also important is the intervening deformation and metamorphism, the latter still unequivocally demonstrated. In the correlatable Grenville Province and Sveconorwegian Orogen (Baltica) tectonic deformation and emplacement of AMCG plutons within the interval 1.1 Ga to 920 Ma are commonly accompanied by high-grade ductile deformation, amphibolite and high-pressure granulite down to greenschist facies metamorphism (Gower, 1996; Corrigan and Hanmer, 1997; Andersson *et al.*, 1999; Larsen, 2000). However the younger Ar/Ar dates on biotite and feldspar recorded in the Rondônia Tin Province by Bettencourt *et al.* (1996) of *c.* 1.001 Ga and 912 Ma, show slow metamorphic cooling rates that are consistent with the K/Ar ages observed in the Younger Rondônia Granite (1.08 Ga - 970 Ma). These ages are readily interpreted as related to cooling as rapakivi magmatism waned during crystallization as stability of the Sunsas Orogen was reached.

Table 1 - Summary of the isotopic ages referred to the rocks of the Central Amazonian Province.



Events between 970 - 920 Ma (Terminal Acivities of the Sunsas Orogen)

The terminal magmatism related to the Sunsas Orogen are the bimodal Guapé Intrusive Suite (Rb/Sr *c.* 950 ± 40 Ma) and the S-type São Domingos Intrusive Suite, dated at 930 ± 19 Ma and 917 ± 5 Ma (U/Pb zircon) by Geraldes (2000). They are related, respectively, to extension and to the Aguapeí thrusting. Post-collisional time-correlatable igneous episodes in the Grenville Province, represented by several granitoid plutons and aplite dykes occurred between *c.* 966 - 956 Ma, following crustal thickening (Tucker and Gower, 1994; Gower, 1996; Wasteneys *et al.*, 1997). In the Sveconorwegian Orogen (1.1 Ga - 900 Ma) (southwestern Sweden and south-southwestern Norway) synchronous post-collisional bimodal rift-related AMCG intrusions and dolerite are recorded at *c.* 966 and 956 Ma. Minor syn-tectonic calc-alkaline magmatism is dated at *c* 1.04 Ga (Bingen *et al.*, 1998; Larsen, 2000). Also marking the end of the tectonic activity, there is the Rogaland AMCG Complex and other norite-anorthosite complexes and related hybrid rocks, which appear to lack Grenville correlatives, (Åhäll and Schöberg, 1996) are recorded in southwestern Sweden.

Table 2 - Summary of the isotopic ages referred to the rocks of the Maroni-Itacaiúnas Province.

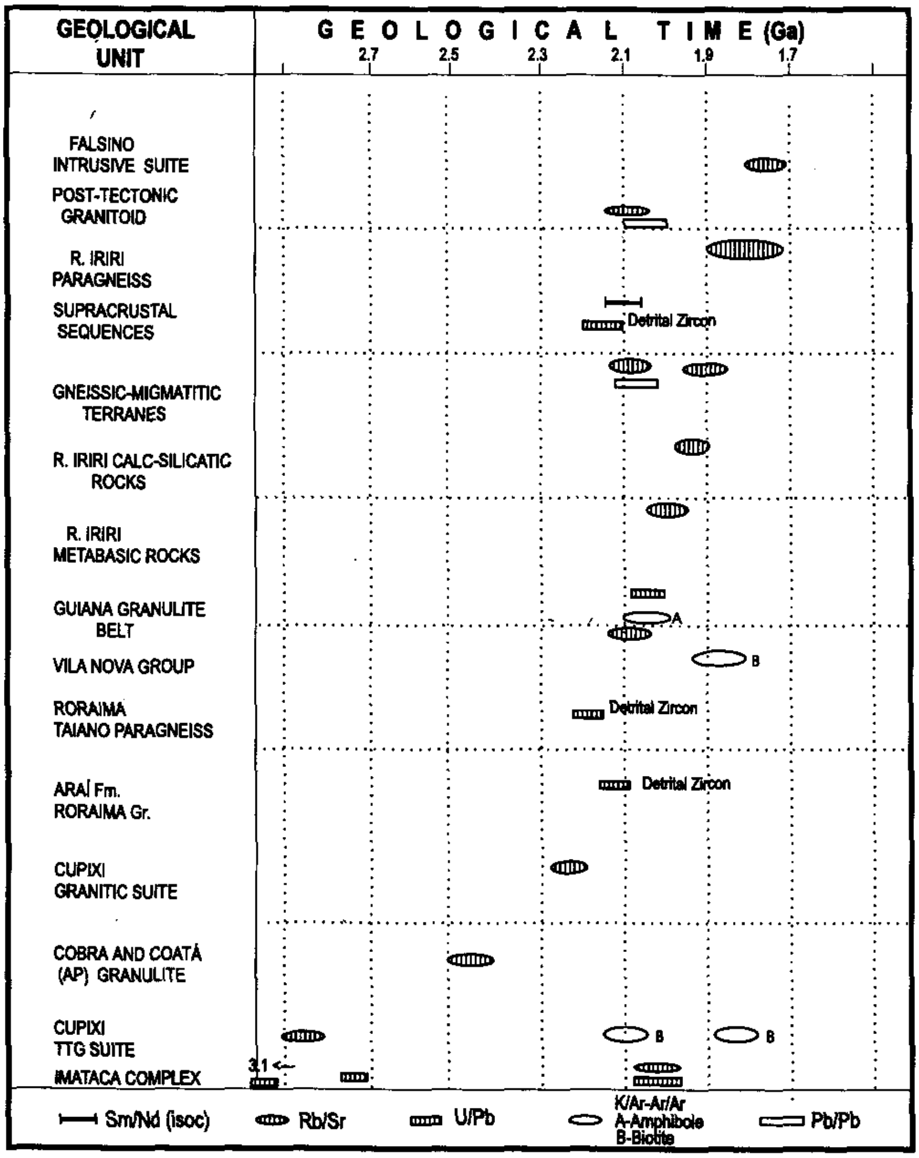


Table 3 - Summary of the isotope ages referred to the rocks of the Ventuari/Tapajós Province.

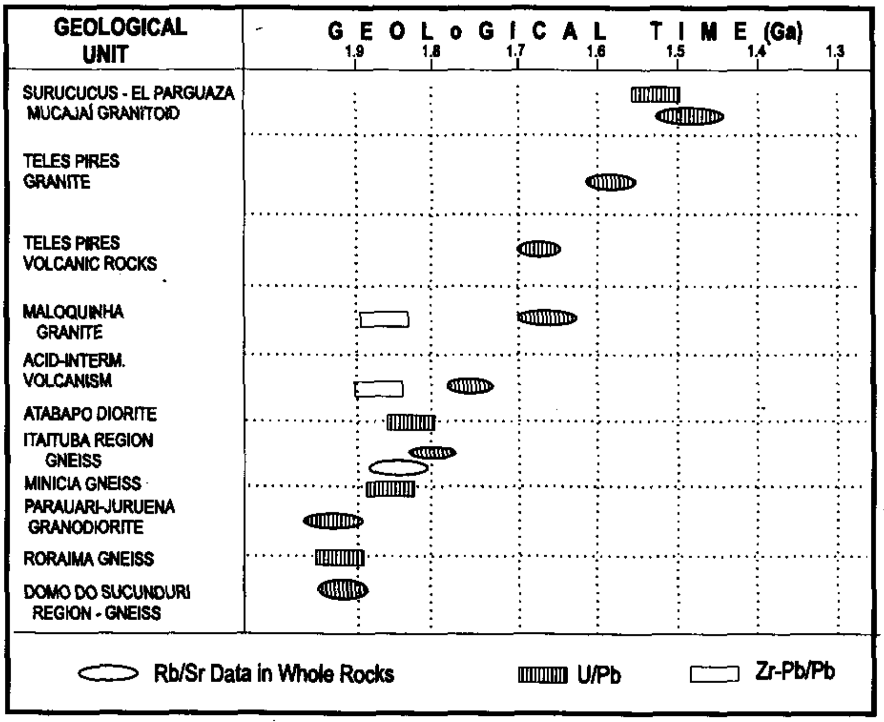


Table 4 - Summary of the isotopic ages referred to the rocks of the Rio Negro/Juruena Province.

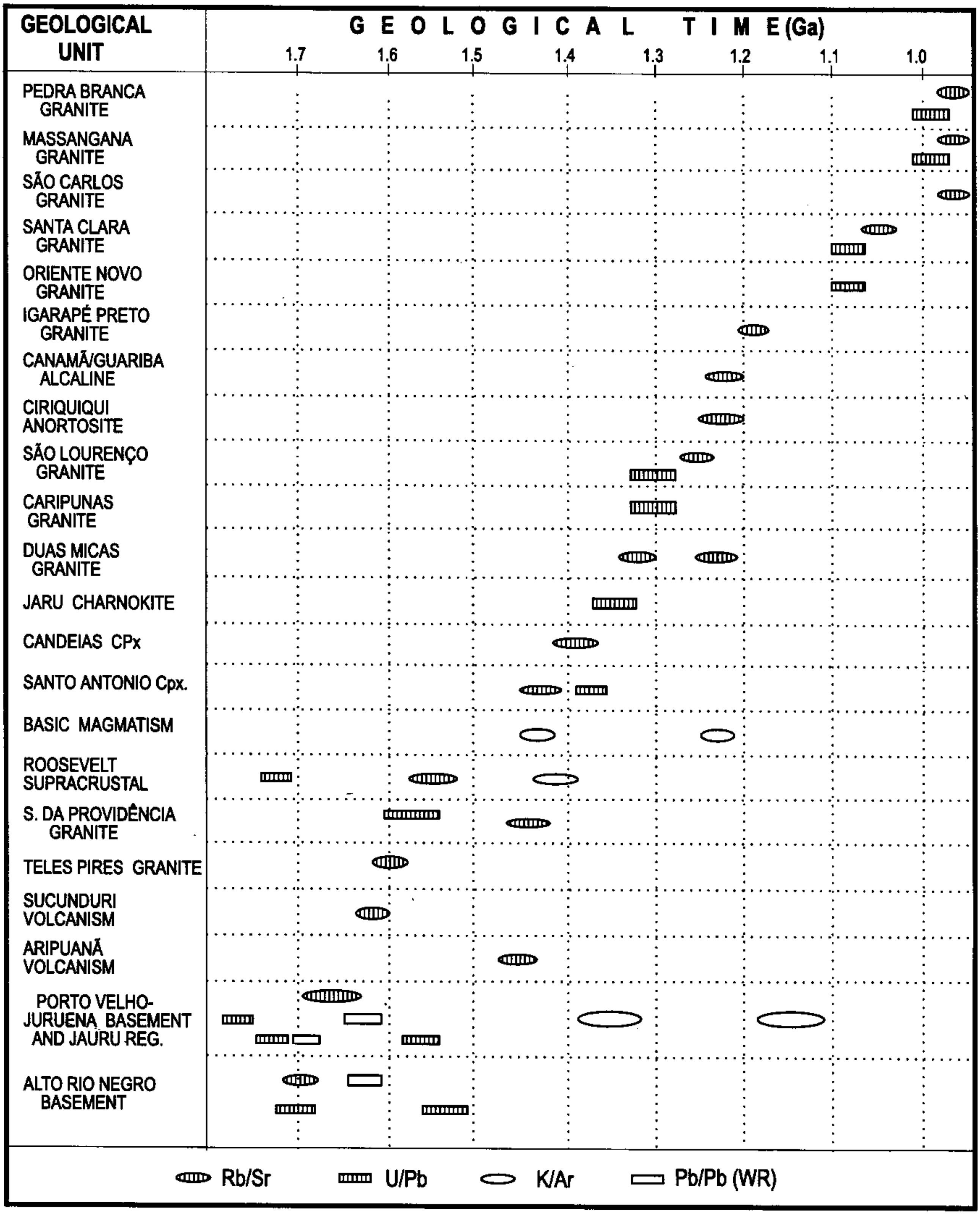


Table 5 - U/Pb and Sm/Nd properties of the Alto Jauru greenstone belt (after Geraldes, 2000).

NUMBER	ROCK DESCRIPTION	U/Pb AGE (Ma)*	$\varepsilon_{Nd}(0)$	$\varepsilon_{Nd}(T)$	T_{DM}
97-131	Cabaçal volcanism	1767±24	-17.7	2.8	1868 Ma
97-133	Pink Gneiss	1795±10	-18.1	2.2	1926 Ma
97-147	São Domingos Gneiss	1746±20	-11.9	2,0	1773 Ma

Table 6 - U/Pb and Sm/Nd properties of the Cachoeirinha calc-alkaline suite (after Geraldes, 2000).

DESCRIPTION	U/Pb (Ma)*	$\varepsilon_{Nd(0)}$	$\varepsilon_{Nd(T)}$	T_{DM}	f
Quatro Marcos Tonalite	1536 ± 11	- 14.2	+ 0.5	1.77	-0.38
Cachoeirinha Tonalite	1549 ± 10	- 14.7	+ 1.0	1.83	-0.40
São Domingos Gneiss	1562 ± 36	- 20.2	+ 0.9	1.79	-0.53
Quatro Marcos Granite	1522 ± 12	- 19.6	+ 0.9	1.78	-0.54
Cachoeirinha Granite	1537 ± 06	- 22.2	+ 0.5	1.75	-0.60
Santa Cruz Granite	1587 ± 04	- 15.0	- 0.8	2.05	-0.36
Água Clara Granodiorite	1485 ± 04	- 5.0	+ 1.7	1.77	-0.50
Araputanga Granite	1440 ± 06	- 20.2	- 0.2	1.74	-0.56
Alvorada Granite (1)	1389 ± 03	- 20.3	- 1.3	1.77	-0.54

*U/Pb zircon ages were made by isotopic dissolution in monocrystal

Table 7 - U/Pb and Sm/Nd properties of the Rio Alegre volcano-sedimentary sequence (after Geraldes, 2000).

NUMBER	ROCK DESCRIPTION	U/Pb AGE (Ma)	$\varepsilon_{Nd(0)}$	$\varepsilon_{Nd(T)}$	T_{DM}
97-122	Metadiorito	1509 ± 10	-2.7	4.3	1.54
97-124	Metadacite	1503 ± 14	-2.4	4.8	1.48
97-134	Metadiorite	1494 ± 11	-11.3	2,5	1.67

Table 8 - U/Pb and Sm/Nd properties of the Santa Helena Suite (after Geraldes, 2000).

SAMPLE	LITHOLOGY	U/Pb (Ma)	$\varepsilon_{Nd(0)}$	$\varepsilon_{Nd(T)}$	T_{DM}	f
97-113	Lavrinha tonalite	1464 ± 25	-13,1	+ 3,8	1,53	-0,45
97-140	Pau-a-Pique tonalite	1481 ± 47	-4,9	+ 4,1	1,50	-0,25
97-108	Guaporé granodiorite	1435 ± 22	-11,8	+ 3,4	1,54	-0,42
97-106	Alto Guaporé gneiss	1424 ± 15	-8,6	+ 4,0	1,49	-0,35
97-102	Triangulo gneiss	1445 ± 04	-15,4	+ 2,9	1,56	-0,51
97-115	Santa Helena granite-gneiss	1433 ± 06	-8,9	+ 3,1	1,62	-0,32
97-141	Maraboa granite	1449 ± 07	-7,1	+ 2,6	1,70	-0,26
97-105	Alto Guaporé augen-gneiss	1424 ± 11	-12,8	+ 2,8	1,57	-0,44
97-120	Cardoso magnetite-granite	1423 ± 15	-11,7	+ 3,6	1,52	-0,33
97-135	Santa Elina granite	1436 ± 06	-10,2	+ 2,7	1,55	-0,38
97-168	Ellus Farm granite	1437 ± 12	-11,1	+ 3,7	1,52	-0,40
97-169	Ellus Mine granite	1444 ± 21	-10,8	+ 3,6	1,51	-0,39

Table 9 - Rb/Sr results of the San Ignacio major units (after Litherland et al., 1986).

STAGE		GRANITOIDS		Rb/Sr AGES (Ma)	$^{87}Sr/^{86}Sr$ INITIAL	K/Ar AGES (Ma)
Syn to Late kinematic		San Rafael		1291 ± 49	0.7003 ± 0.0007	
		San Ramon San Andres Ascension Ibaimini San Javier Marimonos El Puente Refugio Santo Rosario San Pedro Santo Corazon Correreca Tauca Cocalito				
	Pensamiento Complex	Poerto Alegre				
		La Junta		1375 ± 80	0.7052 ± 0.0031	
		Guarayos				1043± 22 (biotite)
		San Martin Compamento				
		Florida				1244 ± 27 1380 ± 19 (biotite)
		Piso Firme		1325 ± 45	0.7044 ± 0.0026	
		Cerro Grande Granophiric Suite	San Simon de Guarayos			
			Cerro Branco			
			Cerro Grande			
Late to Post kinematic		San Cristobal				1296 ± 18 (biotite)
		Porvenir				
		Padre Eterno				1326 ± 19 1268 ± 2 (muscov.)
		Tres Picos				143 ± 14 (biotite)
		Orobayaya		1283 ± 33	0.7058 ± 0.0031	
		Diamantina		1391 ± 70	0.7004 ± 0.0033	
		Discordância				
		El Tigre Alkaline Complex		1286 ± 46		

Table 10 - U/Pb and Sm/Nd properties of the of Rondônia Tin Province granitoid rocks (after Bettencourt et al., 1999a).

EPISODES	RAPAKIVI SUITES	U-Pb AGES (Ma)	Rb-Sr AGES (Ma)	K-AR AGES (Ma)	TECTONIC SETTING
12	Santo Antônio Intrusive Suite	1406±32	1305 (3)		Extensional regime
	Teotônio Intrusive Suite	1387±16	1270 (4)		related to
2	Alto Candeias Intrusive Suite		1358±74 (5)		Rondonian-San Ignacio
	Alto Candeias Batholith	1346±05			Orogenic Cycle or to
		1346±05			opening of the
		1338±05			Grenville Ocean.
3	São Lourenço-Caripunas	1314±13	1268±15 (6)		
	Intrusive Suite	1312±3			
		1309±24			
	Igarapé Preto Intrusive Suite		1195±50 (1)	1195 (7)	
	Ciriquiqui basic and ultrabasic		1300 (7)		

Table 11 - U/Pb and Sm/Nd properties of the Fazenda Reunidas Domain major units (after Geraldes, 2000).

LITHOLOGY	U/Pb (Ma)	$\varepsilon_{Nd(0)}$	$\varepsilon_{Nd(T)}$	T_{DM}	f
Tonalitic Gneiss	1384 ± 40	-4.7	3.6	1.52	-0.24
Carrapato Microgranite	1400 ± 24	-11.2	4.2	1.49	-0.42
Rio Alegre Granodiorite	1412 ± 21	-5.1	3.6	1.58	-0.24
Lajes Granito	1360	-14,8	0.0	1.69	-0.44
Lajes Granito	1606	-14,8	3.4	1.69	-0.44

Table 12 - Rb/Sr results of the Sunsas granitoid (after Litherland et al., 1986).

STAGE	GRANITOIDS	Rb/Sr	$^{86}Sr/^{87}Sr$	K/Ar (Ma)
Pos	Casa de Piedra Granite	1005 ± 12		958 ± 27 (bio)
				911 ± 20 (musc)
Pos	Talcoso			986 ± 27
Pos	Taperas			
Tardi	Limonales			
Tardi	Salinas			
Syn	Espiritu			
Syn	Santa Catalina Zone			
Syn	Las Palmas			
Syn	La Placa			
Syn	San Miguel			
Syn	Motacucito			
Syn	San Pablo			
Syn	El Carmen	972 ± 21		
Pos	Tasseoro			
Pos	Nomoca			991 ± 27
Syn	San Pablo			546 ± 16
Pos	Lucma			
Pos	Rincon del Tigre Complex	993 ± 139		

Table 13 - U/Pb and Sm/Nd properties of the Nova Brasilândia group major units (after Rizzoto, 1999).

SAMPLE	LITHOLOGY	U/Pb (Ma)	$\varepsilon_{Nd(0)}$	$\varepsilon_{Nd(T)}$	T_{DM}
GR-05	Metagabbro	1110 ± 15	+ 4.8	+ 4,3	-
GR-10	Metaturbidite	-	+ 1.0	+ 3.1	-
GR-10A	Biotite Monzogabbro	1098 ± 10	- 10.3	- 0.4	1,63
GR-18	Metagabbro	-	+ 2.3	+ 5.0	-
GR-20	Anatetic leucogranite	1110 ± 08	-12.6	-1.5	1,66
GR-20A	Metaturbidite	-	- 15.3	- 3.9	1,85
GR-20A$_1$	Metaturbidite	-	-14.9	- 3.8	1,85
GR-20C	Calc-silicated gneiss	-	- 15.2	- 4.3	1,91
GR-23	Porphire Monzogranite	995 ± 15	- 10.6	+ 0.5	1,50

Table 14 - Granitoids of Rondônia Tin Province major units U/Pb and Sm/Nd results (after Bettencourt et al., 1999a).

EPISODES	RAPAKIVI SUITES	U-Pb AGES (Ma)	Rb-Sr AGES (Ma)	K-Ar AGES (Ma)	TECTONIC SETTING
1	Santa Clara Intrusive Suite		1052±21 (5)	1035 (8)	Extentional regime related to collisional stage of Sunsas-Aguapei Orogenic Cycle.
	Manteiga Massif	1082±05			
	Santa Clara Massif	1081±50 1074±21			
	Oriente Novo Massif	1080±27 1074±08			
	Costa Marque Group		1018±76 (9)		
2	Younger Granite of Rondônia		956±09 (6)		
	Pedra Branca Massif	996±05		950 (8)	
	São Carlos Massif	995±73 974±06			
	Massangana Massif	991±14		1000 (7)	
	Nova Floresta Formation				

Table 15 - Precambrian geological history of the Amazonian Craton.

MAJOR EPISODES OF CRUST FORMATION	MAIN EVENTS OF ACCRETION OF JUVENILE CRUST	DIVERGENT TECTONISM WITHIN-PLATE MAGMATISM	SEDIMENTS AND ASSOCIATED MAGMATISM IN CONTINENTAL RIFT SYSTEMS
1.28-1.1Ga SP	1.1 Ga SP	1.05-0.9Ga RSIP	1.1 - 1.0Ga RSIP
1.5 -1.3Ga RSIP	1.48-1.42Ga RSIP	1.45 - 1.2Ga RNJP	1.45 - 1.2Ga RNJP
1.8 -1.55Ga RNJP	1.8-1.7 and 1.57-1.53 Ga RNJP	1.75 - 1.5Ga VTP-RNJ	1.65 - 1.4Ga RNJP
1.95 - 1.8Ga VTP	1.95 - 1.8Ga VTP	1.85 - 1.8Ga CAP	1.90 - 1.6Ga CAP-VTP
2.25 - 2.0Ga MIP	2.1 - 1.95Ga 2.25 - 2.1Ga MIP		1.95 - 1.8Ga CAP
2.9 - 2.7Ga CAP	2.98 – 2.87 Ga CAP	2.6 - 2.3Ga CAP	
3.1 - 3.0Ga CAP	>3.0 Ga (?) CAP		

CAP - Central Amazonian Province
VTP - Ventuari-Tapajós Province
RSIP - Rondonian-San Ignácio Province
MIP - Maroni-Itacaiúnas Province
RNJP - Rio Negro-Juruena Province
SP - Sunsás Province

REFERENCES

Åhäll, K.I. and Connelly, J.N. 1996. Proterozoic plate geometry in the North Atlantic region, constrains from persistent 1.51-1.15 Ga anorogenic magmatism in Scandinavia. In: *Proterozoic evolution in the North Atlantic realm* (compiled by C.F. Gower). COPENA-ECSOOT-IBTA Conference, Goose Bay, Labrador, 1996. Program and Abstract, 3-4.

Åhäll, K.I. and Schöberg, H. (1996). West-northwest trending post-collisional intrusions in SW Sweden: Implications for 963 Ma block movements in the Sveconorwegian orogen. In: *Proterozoic Evolution in the North Atlantic Realm* (compiled by C.F. Gower). COPENA-ECSOOT-IBTA Conference, Goose Bay, Labrador, July 29-August 2, 1996. Program and Abstracts, 3-4.

Åhäll, K.I. and Gower, C.F. (1997). The Gothian and Labradorian orogens: variations in accretionary tectonism along a late Paleoproterozoic Laurentia-Baltica margin. *GFF*, **119**, 181-191.

Almeida, F.F.M. de (1978). A evolução dos Crátons Amazônico e do São Francisco comparada com os seus homólogos do hemisfério norte. In: *30 Congresso Brasileiro de Geologia*, **6**, 2393-2407.

Almeida, M.E., Fraga, L.M.B. and Macambira, M.J.B. (1997). New geochronological data of calc-alkaline granitoids of Roraima State, Brazil. *South American Symposium on Isotope Geology. Extended Abstracts*, 34-35.

Althoff, F.J., Dall'Agnol, R. and Souza, Z.S. (1991). Região de Marajoara - SE do Pará, prolongamento dos terrenos arqueanos de Rio Maria ou retrabalhamento? *Simpósio Geológico da Amazoniania, III, Anais. Belém, Brazil*, 130-141.

Amaral, G. (1974). *Geologia pré-cambriana da região Amazônica*. Tese de Livre-Docência, Instituto de Geociências, Universidade de São Paulo, 212 p.

Andersson, S.L. (1988). Interpretations of K-Ar mineral dates from the Greenville orogenic belt. *American Journal Science*, **288**, 701-734.

Andersson, J. Söderlung, U., Cornell, D., Johansson, L., Möller, C. (1999). Sveconorwegian (Grenvillian) deformation, metamorphism and leucosome formation is SW Sweden, SW Baltic Shield: constraints from a Mesoproterozoic granite intrusion. *Precambrian Research*, **98**, 151-171.

Araújo, O.J.B., Maia, R., Jorge João, X.S. and Costa, J.B.S. (1988). A megaestruturação arqueana da Folha Serra dos Carajás. *VII Congresso Latino Americano de Geologia, Extended Abstracts, Anais*, 324-333.

Araújo, O.J.B,. and Maia, R. (1991). *Programa Levantamentos Geológicos Básicos do Brasil. Programa Grande Carajás. Folha SB.22-Z-A Serra dos Carajás*. Belém, CPRM.

Barbosa, A.A,. and Lafon, J.M. (1996). Geocronologia Pb-Pb e Rb-Sr de granitóides arquenos da região de Redenção - Sul do Pará. *Revista Brasileira de Geociências*, **26**, 255-264.

Barrios, F.J. (1983). *Caracterização geocronológica da região amazônica da Venezuela*. Dissertação MSc. Instituto de Geociencias da Universidade de São Paulo, 123 p.

Barros, A.M., Silva, R.H. da, Cardoso, O.R.F.A., Freire, F.A., Souza Jr, J.J. da, Rivetti, M., Luz, D.S. da Palmeira, R.C. de B. and Tassinari, C.C.G. (1982). *Projeto RadamBrasil. Folha SD-21 (Cuiabá)*, 192 p.

Barros, C.E.M., Dall'Agnol, R., Lafon, J.M., Teixeira, N.P. and Ribeiro, J.W., 1992. Geologia e geocronologia Rb-Sr do Gnaisse Estrela, Curionópolis, PA. *Bol. Museu Paraense Emílio Goeldi*, **4**, 83-102.

Basei, M.A.S (1977). *Idade do vulcanismo ácido a intermediário na região Amazônica*. Dissert. de Msc., Instituto de Geociências da Universidade de São Paulo, 133 p.

Beisiegel, V.R., Bernadelli, A.L., Drummond, N.F., Ruff, A.W. and Tremaine, J.W. (1973). Geologia e recursos minerais da Serra dos Carajás. *Revista Brasileira de Geociências*, **3**, 215-242.

Bell, S., Condie, K. and Geraldes, M.C. (1999). Origin of arc-related granitoids from the SW Amazonian craton: Juvenile crustal additions at 1550-1450 Ma. *The Geological Society of America 1999 Annual Meeting. Denver-CO. Abstract with Programs*, 205 p.

Ben Othman, D., Polvé, M. and Allégre, C.J. (1984). Nd-Sr Isotopic composition of granulites and constrains on the evolution of the lower continental crust. *Nature*, **307**, 510-515.

Bettencourt, J.S., Tosdal, R.M., Leite, W.B., Jr,. and Payolla, B.L. (1995). Overview of the rapakivi granites of the Rondônia Tin Province. In: The rapakivi granites of Rondônia tin province and associated mineralization, eds. Bettencourt, J.S. and Dall'Agnol, R. *Symp. Rapakivi Granites and Related Rocks, 6th., Belém, Brazil, 1995. Excursion Guides, UFPA*, p. 5-

Bettencourt, J.S., Onstot, T.C,. and Teixeira, W. (1996). Tectonic interpretation of 40Ar/39Ar ages on country rocks from the central sector of the Rio Negro-Juruena Province, southwest Amazonianian Craton. *International Geology Review*, **38**, 42-56.

Bettencourt, J.S., Leite JR., W.B. and Payolla, B.L. Scandolara, J.E., Muzzolon, R. and Vian, J.A.J. (1997). The rapakivi granites of the Rondônia Tin Province, northern Brazil. In: *International Symposium on Granites and Associated Mineralzations, Salvador, Bahia, Brazil. 24-29 August, 1997, Excursions Guide*. Companhia Bahiana de Pesquisa Mineral e Superintendência de Geologia e Recursos Minerais do Estado da Bahia, Brasil, 3-31.

Bettencourt, J.S., Tosdal, R.M., Leite JR., W.B. and Payolla, B.L. (1999a). Mesoproterozoic rapakivi granites of the Rondônia Tin Province, southwestern border of the Amazonianian craton, Brazil – I. Reconnaissance U-Pb geochronology and regional implications. *Precambrian Research*, **95**, 41-67.

Bettencourt, J.S., Payolla, B.L., Leite Jr., W.B., Tosdal, R.M. and Spiro, B. (1999b). Mesoproterozoic rapakivi granites of Rondônia Tin Province, southwestern border of Amazonianian craton, Brazil: Reconnaissance Nd, Sr, O, Pb isotopic geochemistry and regional implications. In: Barbaran, B. (ed.) *Fourth Hutton Symposium Clermont Ferrand, France. Abstracts. Documents du BRGM* 290, 132 p.

Berrangé J.P. (1977). The Geology of southern Guyana. Inst. Geol. Sciences *Overseas memoir*, **4**, 112 p.

Bingen, B., Boven, A. Punzalan, L. Wijbrans, J.R. and Demaiffe, D. (1998). Hornblende $^{40}Ar/^{39}Ar$ geochronology across terrane boundaries in the Sveconorwegian Province of S. Norway. *Precambriam Research*, **90**, 159-186.

Bosma, W., Kroonenberg, S.B., Maas, K. and De Roever, E.W.F. (1983). Igneous and metamorphic complexes of the Guiana Shield in Suriname. *Geologie en Mijnbouw*, **62**, 241-254.

Carneiro, M.A., Ulbrich, H.H.G.J. and Kawashita, K. (1992). Proterozoic crustal evolution at the southern margin of the Amazonianian craton in the State of Mato Grosso, Brazil: evidence from Rb-Sr and K-Ar data. *Precambriam Research*, **59**, 263-282.

Choubert, B. (1974). *Le Précambrien des Guyanes*. Mém BRGM, Orléans. 81, 213 p.

Choudouri A. (1980). The early proterozoic greenstone belt of htr northern Guiana shield, South America. *Precambriam Research*, **13**, 363-374.

Cordani, U.G., Tassinari, C.C.G., Teixeira, W., Basei, M.A.S. and Kawashita, K. (1979). Evolução Tectônica da Amazônia com base nos dados geocronológicos. *II Congresso Geológico Chileno, Arica, Chile, Actas*, 137-48.

Cordani, U.G. and Brito Neves, B.B. (1982). The geologic evolution of Ssouth America dduring the Archean and Early Proterozoic. *Revista Brasileira de Geociências, São Paulo*, **12**(11-3), 78-88.

Cordani U.G., Tassinari, C.C.G. and Kawashita, K. (1984). A Serra dos Carajás como região limítrofe entre províncias tectônicas. *Ciências da Terra*, **9**, 6-11.

Cordeiro, A.A.C. and Saueressig, R. (1980). Serra das Andorinhas:

geologia e principais ocorrências de ouro. *Congresso Brasileiro de Geologia, XXXI, Resumos, Camboriú, Brazil,* 2, 344.

Corrigan, D. and Hanmer, S. (1997). Anorthosites and related granitoids in the Grenville orogen: a product of convective thinning of the lithosphere? *Geology,* 25, 61-64.

Costa, J.B.S., Araújo, O.J.B., Santos, A., Jorge João, X.S., Macambira, M.J.B. and Lafon, J.M. (1995). A Província Mineral de Carajás: Aspectos tectono-estructurais, estratigráficos e geocronólogicos. *Bol. Museu Paraense Emílio Goeldi,*7, 199-235.

Costa, J.B.S. and Hasui, Y. (1997). Evolução Geológica da Amazônia. In: *Contribuições à Geologia da Amazônia,* eds. Costa, M.L. and Angélica, R.S., pp.15-90. Belém, SBG.

Cunha, B.C.C., Potiguar, L.A.T., Ianhez, A.C., Bezerra, P.E.L., Pitthan, J.H.L., SOUZA JR., J.J., Montalvão, R.M.G., Souza, A.M.S., Hildred, P.R. and Tassinari, C.C.G. (1981). *Levantamentos dos Recursos Naturais.* 22, Folha SC22 Tocantins. Geologia I, NME/SG, 196 p.

Dall'Agnol, R., Lafon, J.M. and Macambira, M.J.B. (1994). Proterozoic anorogenic magmatism in the Central Amazonianian Province, Amazonianian craton: geochronological, petrological and geochemical aspects. *Mineral Petrology,* 50, 113-138.

Dall'Agnol, R., Souza, Z.S., Althoff, F.J., Barros, C.E.M., Leite, A.A.S. and Jorge João, X.S. (1997). General aspects of the granitogenesis of the Carajás Metallogenic Province. *Intern. Symp. Granites Assoc. Mineraliz., II, Excurs. Guide. Salvador, Brazil,* 135-142.

Dall'Agnol, R., Souza, Z.S., Althoff, F.J., Macambira, M.J.B. and Leite, A.A.S. (1996). Geology and geochemistry of the Archaean Rio Maria granite-greenstone terranes, Carajás Province, Amazonian Craton. *Symp. Archaean terranes of the S. America plataform, Extended Abstract Brasília, Brazil,* 29-30

Dall'Agnol, R., Ramo, O.T., Magalhães, M.S. and Macambira, M.J.B. (1999a). Petrology of the anorogenic, oxidised Jamon and Musa granites, Amazonianian Craton: implications for the genesis of Proterozoic A-type granites. *Lithos,* 46, 431-462.

Dall'Agnol, R., Silva, C.M.G. and Scheller, T. (1999b). Fayalite-hedembergite rhyolites of the Iriri Formation, Tapajós gold province, Amazonianian Craton: implications for the Uatumã volcanism. *Simp. Vulc. Amb. Assoc., I, Gramado, Brazil,* 31.

Dall'Agnol, R. and Macambira, M.J.B. (1992). Titanita-Biotita Granitos do Baixo Rio Uaupés, Província Rio Negro, Amazonianas. Parte I: Geologia, Petrografia, e Geocornologia. *Revista Brasileira de Geociências,* 22-1, 3-14.

Darbyshire, D.P.F. (1979). Results of the age determinations programme. Rep. E. Bolivia Miner. Explor. Proj. (Projecto Precámbrico), Phase I, 9 Interim report available on open file in Bolivia (GEOBOL) and the United Kingdom -IGS (umpublished).

Dardenne, M.A., Ferreira Filho, C.F. and Meirelles, M.A. (1988). The role of shoshonite and calc-alkaline suites in tectonic evolution of the Carajás District, Brazil. *Journal of South American Earth Sciences,* 1, 363-372.

Deckart, K., Féraud, G. and Bertrand, H. (1997). Age of jurassic continental tholeiites of French Guyana, Surinam and Guinea: Implications for the initial opening of the central Atlantic ocean. *Earth and Planetary Sciences Letters,* 150, 205-220.

Delor, C., Rossi, Ph., Cocherie, A., Capdevilla, R., Peucat, J.J. and Vidal, M. (1998). The French Guyana basement revisited: New petrostructural and geochronological results and correlations with the West African shield. In: *Congresso Brasileiro de Geologia, 40, Belo Horizonte, Anais* p.49.

De Vletter D.R., Aleva, G.J.J. and Kroonenberg, S.B. (1998). Research into the Precambrian of Suriname. In: *The history of earth sciences in Suriname. Royal Netherlands Academy of Arts and Sciences,* eds. Wong, Th.E., de Vletter, D.R., Krook, L., Zonneveld, J.I.S. Van Loon, A.J., pp. 15-64. Netherlands Institute of Applied Geoscience TNO, Amsterdam.

DOCEGEO (1988). Revisão litoestratigráfica da Província Mineral de Carajás. *XXV Congresso Brasileiro de Geologia Anexo aos Anais,* 11-56.

Egal, E., Milési, J.P., Ledru, P., Cautru, J.P., Freyssinet, P., Thiéblemont, D. and Vernhet, Y. (1994). *Ressources minèrales et évolution lithostructurale de la Guyane. Carte thématique minière d 1/100 000, Feuille Cayenne.* Rapport BRGM, R 38019, 59 p.

Emslie, R.F. and Hunt, P.A. (1990). Ages and petrogenetic significance of igneous mangerite-charnockites suites associated with massif anorthosites, Grenville province. *Journal of Geology,* 98, 213-231.

Faria, M.S.G., Luzardo, R. and Pinheiro, S.S. (1999). Litoquímica e petrogênese do Granito Igarapé Azul - Sudeste de Roraima. *Simpósio Geológico da Amazoniania, VI, Boletim Resumos Expandidos Belém, Brazil,* 577-580.

Ferreira Filho, C.F. and Bizzi, L.A. (1985). Caracterização dos cumulatos máficos granulitizados do Rio Alegre, M.T. *II Simp. de Geol. do Centro Oeste. Goiânia,* 239-246.

Figueiras, A.J.M. and Villas, R.N.N. (1982). Estudo petrológico e sedimentológico da sequência clástica (pós-grupo Grão Pará) da Serra dos Carajás, Estado do Pará. *Congresso Brasileiro de Geologia, XXXII, Anais. Salvador, Brazil,* 2, 832-846.

Figueiredo, A.J., Rodrigues, A.P., Pimentel, G. de B., Reischl, J.L., Rezende Filho, S.T. de and Ribeiro Filho, W. (1974). *Projeto Alto Guaporé.* Goiania. DNPM/CPRM. Relatório Final, IV, 35 p.

Fraga, L.M and Reis, N.J. (1995). The rapakivi granite-anorthosite association of Mucajaí region- Roraima State, Brazil. *Simposium on Rapakivi Granites and related rocks, Extended Abstract Bulletin,* 31-32

Fraga, L.M., Almeida, M.E. and Macambira, M.J.B. (1997). First lead-lead zircon ages of charnockitic rocks from Central Guiana Belt (CGB) in the state of Roraima, Brazil. *South American Symposium on Isotope Geology, Actas, Extended Abstracts,* 115-117.

Fuck, R.A., Pimentel, M.M., Machado, N. and Daoud, W.K. (1993). Idade U-Pb do Granito Madeira, Pitinga (AM). *4 Congresso Brasileiro de Geoquimica,* 246-248.

Galvis, V.J., Huguetti, G.A., Ruge, T.P. (1979). Geologia de la Amazoniania Colombiana. *Boletim Geologico Ingeominas,* XXII, 3, 1-153.

Gaudette, H.E., Hurley, P.M., Fairbairn, H.W., Espejo, A. and Dahlberg, E.H., 1976. Older Guiana basement south of the Imataca complex in Venezuela and in Suriname. 24th Progr. Rept., 1974-1976. MIT Geochronol. Labor. R.M. 54-1122, 26-34.

Gaudette, H.E. and Olszewski, W.J., Jr. (1981). Geochronology of the basement rocks, Amazonianas Territory , Venezuela. *I Symposium Amazonianico.* pp. 24-25.

Gaudette, H.E. and Olszewski, W.J., Jr. (1985). Geochronology of the basement rocks, Amazonianas territory, Venezuela and the tectonic evolution of the western Guyana shield. *Geologie en Mijnbouw,* 64, 131-144.

Gaudette, H.E., Olszewski, W.J., Jr. and Santos, J.O.S (1996). Geochronology of Precambrian rocks from the northern part of the Guiana shield, state of Roraima, Brazil. *Journal of South America Earth Sciences,* 9(3/4), 183-196.

Geraldes, M.C. (1996). *Estudos geoquímicos e isotópicos das mineralizações auríferas e rochas associadas da região de Pontes e Lacerda, (MT).* Dissertação de MSc., Instituto de Geociências da Universidade de Campinas, 104 p.

Geraldes, M.C. (2000). *Geocronologia e Geoquímica do Plutonismo Mesoproterozóico do SW do Estado de Mato Grosso (SW do Cráton Amazônico).* Doctored Dissertation. USP- São Paulo, 187 p.

Geraldes, M.C., Kozuch, M., Teixeira, W. and Van Schmus, W.R. (1997). U/Pb constraints on the origin of Mesoproterozoic granites of Pontes de Lacerda region, SW of Amazonian Craton.

South American Symposium on Isotope Geology. Campos do Jordão, Brazil, 120 p.

Geraldes, M.C., Matos, J., Ruiz, A., Fetter, A.H., Kozuch, M., Van Schmus, W.R., Tassinari, C.C.G. and Teixeira, W. (1999). U/Pb Constrains on Proterozoic Magmatic Arcs in SW Amazônia Craton, Brazil. *II South American Symposium on Isotope Geology, Argentina*, 143 p.

Geraldes, M.C., Teixeira, W. and Van Schmus, W.R. (2000) Rio Alegre Terrane in SW Amazonianian craton: a suture zone recording a mesoproterozoic collision. *2000 GSA Southwestern Meeting, Charlleston, South Caroline*, USA (submitted).

Gibbs, A.K. and Olszewski, W.J., Jr. (1982). Zircon U-Pb ages of Guyana greenstone-gneiss terrane. *Precambriam Research*, **17**, 199-214.

Gibbs, A.K. and Barron, C.N. (1983). The Guyana Shied reviewed. *Episodes*, **6-2**, 7-14.

Gibbs, A.K. and Barron, C.N. (1993). The Geology of the Guiana Shield. Oxford University Press (New York), Clarendon Press (Oxford). *Oxford monographs on geology and geophysics*, **22**, 246 p.

Gibbs, A.G., Wirth, K.R., Hirata, W.K. and Olszewski Jr., W.J. (1986). Age and composition of the Grão-Pará Group volcanics, Serra dos Carajás. *Revista Brasileira de Geociências*, **16**, 201-211.

Gibbs, A.K. (1980). *Geology of Barama-Mazaruni Supergroup of Guyana*. PhD Dissertation, Harvard Univ. Microfilms Intern. Ann. Aarbor MI, 81022054, 387 p.

Gomes, C.B., Cordani, U.G. and Basei, M.A.S. (1975). Radiometric ages from Serra dos Carajás area northern Brazil. *Geological Society of America Bulletin*, **86**, 55-89.

Ghosh, S.K. (1981). Geology of the Roraima Group and Its Correlation. *Symposium Amazonianico, Actas*, 26-27.

Gower, C.F. (1996). The evolution of the Grenville province in eastern Labrador, Canada. In: Precambrian crustal evolution in the North Atlantic region, ed. Brewer, T.S. *Geological Society, Special Paper*, **112**, 197-218.

Gower, C.F. and Tucker, R.D. (1994). Distribuition of pre 1400 Ma crust in the Grenville province: Implications for rifting in Laurentia-Baltica during geon 14. *Geology*, **22**, 827-830.

Gruau, G., Martin, H., Leveque, B. and Capdevila, R. (1985). Rb/Sr and Sm/Nd Geochronology of lower Proterozoic granite-greenstone terrains in French Guiana, South America. *Precambriam Research*, **30**, 63-80.

Hasui, Y., Haraly, N.L.E. and Schobbenhauss, C. (1984). Elementos geofísicos e geológicos da região amazônica: Subsídios para o modelo geotectônico. *II Symposium Amazônico, Anais*, 129-148.

Haapala, I. and Rämö, O.T. (1995). Geology of the Rapakivi Granites – A Review. *Symposium on Rapakivi Granites and Related Rocks. Abstracts Volume*, 4.

Hebeda, E.H, Boelrijk, N.A.I.M., Priem, H.N.A., Verdurmen, E.A.Th. and Verschure, R.H. (1973). Excess radiogenic argon in Precambrian Avanaver dolerite in western suriname (South America). *Earth and Planetary Sciences Letters*, **20**, 189-200.

Hirata, W.K., Rigon, J.C., Kadekaru, K., Cordeiro, A.A.C. and Meireles, E.M. (1982). Geologia regional da Província Mineral de Carajás. *Simpósio Geológico da Amazoniania, I, Anais. Belém, Brazil*, 1,100-108.

Hoffman, P.F. (1992). Global Grenville kinematics and fusion of the Neoproterozoic supercontinent Rodinia. *Geological Association of Canada, Program with Abstract*, **17**, 49.

Huhn, S.R.B. (1991). Controle estrutural dos depósitos e ocorrências auríferas nos terrenos granito-greenstones da região de Rio Maria. *Simpósio Geológico da Amazoniania, III, Anais. Belém, Brazil*, 211-219.

Huhn, S.B., Macambira, M.J.B. and Dall'Agnol, R. (1999). Geologia e geocronologia Pb/Pb do Granito Alcalino Arqueano Planalto, região da Serra do Rabo, Carajás, PA., 1999. *Simpósio Geológico da Amazoniania, VI, Boletim Res. Expandidos Belém, Brazil*, 463-466.

Hurley, P.M., Melcher, G.C., Pinson, W.H. and Fairbairn, H.W. (1968). Some orogenic episodes in South America by K-Ar and whole-rock Rb-Sr dating. *Canadian Journal of Earth Sciences*, **5**, 633-638.

Isotta, C.A.L., Carneiro, J.M., Kato, H.T. and Barros, R.J.L. (1978). *Projeto Província Estanífera de Rondônia. Porto Velho*. DNPM/CPRM. Vols 1,2, e 3, Relatório Final.

Iwanuch, W. (1999). Evolução geológica com base em dados geocronológicos da porção sudeste do Estado do Amazonianas e do trato contíguo do norte do Estado de Mato Grosso na região do baixo e médio Rio Juruena. *VI Simpósio de Geologia da amazonia, Ext. Abstract Bull.*, 467-470.

João, X.S.J. et al. (1979). *Geologia da região do sudoeste do Amapá e norte do Pará. Projeto Sudoeste do Amapá*. DNPM-CPRM. Serie Geol. n×10. Sec. Geol. Bás. n×7 125 p.

João, X.S.J. and Marinho, P.A.C. (1982a). Catametamorfitos arqueanos da região centro-leste do território Federal do Amapá. In: *Simpósio Geológico da Amazoniania 1, Belém. Anais... Belém, SBG*, **2**, 207-228.

João X.J.S. and Marinho, P.A.C. (1982b). Granitóides sódicos da região centro leste do Território Federal do Amapá. In: *Simpósio Geológico da Amazoniania 1, Belém. Anais... Belém, SBG*, **2**, 229-252.

João, X.S., Santos, C.A. and Provost, A. (1985). Magmatismo adamelítico Água Branca (Folha Rio Mapuera - NW do Estado do Pará). *Simpósio Geológico da Amazoniania, II, Anais. Belém, Brazil*, **2**, 93-109.

Kroonenberg, S.B. (1976). Amphibolite-facies and granulite-facies metamorphism in the Coeroeni-Lucie area, southwestern Surinam. PhD Thesis. Univ. Amsterdam. *Geologie en Mijnbouw Dienst. Sur.*, **25**, 109-289.

Kroonenberg, S.B. (1982). A Grenvillian granulite belt in the Colombian Andes and its relation to the Guiana shield. *Geologie en Mijnbouw*, **61**, 325-333.

Lafon, J.M., Rodrigues, E.M.S. and Duarte, K.D. (1994). Le granite Mata Surrão: un magmatisme monzogranitique contemporain des associations tonalitiques-trondhjemitiques-granodioritiques archénnes de la région de Rio Maria (Amazonianie orientale, Brésil). *Comptes Rendus de l'Académie des Sciences, Paris*, **318**, 643-648.

Lafon, J.M. and Scheller, T. (1994). Geocronologia Pb/Pb em zircão do Granodiorito Cumaru, Serra dos Gradaús, PA. *Simpósio Geológico da Amazoniania, IV, Anais, Belém, Brazil*, 321-324.

Lafon, J.M., Rossi, P., Delor, C., Avelar, V.G. and Faraco, M.T.L. (1998). Novas testemunhas de relíquias arqueanas na crosta continental paleoproterozóica da Província Maroni-Itacaiúnas (Sudeste do Escudo das Guianas). In: *Congresso Brasileiro de Geologia, 40, Belo Horizonte, Anais*, 64.

Laitakari, R., Suominen, V.N., Stepanov, K. and Amantova, A. (1996). Subjotnian: Rapakivi granites and related rocks in the surroundings of the Gulf of Finland. In: Koistinen, T.J. (ed.), Explanation to the Map of Precambrian basement of the Gulf of Finland and surrounding area 1: 1 mill. *Geological Survey of Finland Special Paper*, **21**, 59-97.

Lamarão, C.N., Dall'Agnol, R., Lafon, J.M. and Lima, E.F. (1999). As associações vulcânicas e plutônicas de Vila Riozinho e Morais Almeida, Província aurífera do Tapajós, SW do Estado do Pará. *Simp. Vulc. Amb. Assoc., I, Gramado, Brazil, 93.*

Larsen, S.Å. (2000). The Sveconorwegian tectonic cycle reviewed. *IGC 2000, Rio de Janeiro*. (Submited).

Leal, J.W.L., Silva, G.H., Abreu, A.S. and Lima, M.I.C. (1976). Granito Serra da Providência. In: Congr. Bras. Geol., 29, Ouro Preto, Brasil, 1976. Anais... Ouro Preto. *Sociedade Brasielira de Geociências*, **4**, 59-74.

Leal, J.W.L., Silva, G.H., Santos, D.B., Teixeira, W., Lima, M.I.C., Fernandes, C.A.C. and Pinto, A.C. (1978). Geologia. *Projeto Radambrasil. Folha SC.20. Porto Velho. Rio de Janeiro, MME/DNPM*, 19-184.

Ledru, P., Lasserre, J.L., Manier, E. and Mercier, D. (1991). Le Protérozoïque inférieur nord guyanais: révision de la lithologie, tectonique transcurrente et dynamique des bassins sédimentaires. *Bulletin de la Société Géologique de France*, **162-4**, 627-636.

Ledru, P., Johan. V., Milési, J.P. and Tegyey, M. (1994). Markers of the last stages of the Paleoproterozoic collision: evidence for a 2 Ga continent involving circum-South Atlantic provinces. *Precambriam Research*, **69**, 169-191.

Leite, J.A.D. (1989). *Contexto geológico e geoquímica das lavas máficas da Sequência Vulcano-Sedimentar Quatro Meninas, Município de Indiavaí, MT. Porto Alegre. UFRGS.* Dissertação de Mestrado, 82 p.

Leite, J.A.D., Saes, G.S. and Ruiz, S. (1986). Anatomia e Interpretação das Lavas Basálticas subaquosas da Sequência Vulcano-Sedimentar Quatro Meninas, Araputanga, M.T. In: *34 Cong. Bras. de Geol., Bol. de Res. e Brev. Com.*, 99 p.

Leite Jr., W.B., Bettencourt, J.S. and Payolla, B.L. (1996). The alkaline complex suite of the Younger rapakivi granites of Rondônia, Brazil. In: Haapala, I., Rämö, O.T. and Kosumen, P. (eds.). *Symposium on Rapakivi Granites and Related Rocks, Abstaract Volume, University of Helsinki, University Press, Helsinki, Finland*, 49-50.

Lima, M.I.C., Montalvão, R.M.G., Issler, R.S., Oliveira, A.S., Basei, M.A.S., Araújo, J.F.V. and Silva, G.G. (1974). Geologia. In: *Projeto RADAM. Folha NA/NB 22 Macapá. Rio de Janeiro, I/* 120 p. (levantamentos de recursos naturais, 6).

Lima, M.I.C., Oliveira, E.P. and Tassinari, C.C.G. (1982). Cinturões granulíticos da porção setentrional do Craton Amazônico. In: *Simpósio Geológico da Amazoniania 1, Belém*, **1**, 147-162.

Lima, M.I.C., Santos, J.O.S and Siga Jr., O. (1986). Os terrenos de alto grau do Craton Amazonianico. *Congresso Brasileiro de Geologia*, **2**, 751-765.

Litherland, M. and Bloomfield, K. (1981). The Proterozoic History of eastern Bolivia. *Precambriam Research*, **15**, 157-179.

Litherland, M., co-ordinador (1986). The geology and mineral resources of the Bolivian Precambrian Shield. *Overseas Memoir Br. Geol. Survey*, **9**, 153.

Litherland, M. et al. (1989). The Proterozoic of eastern Bolivia and its relationship to the Andean Mobile Belt. *Precambriam Research*, **43**, 157-174.

Lindenmayer, Z.G., Laux, J.H. and Viero, A.C. (1995). O papel da alteração hidrotermal nas rochas da bacia Carajás. *Boletim Museu Paraense Emilio Goeldi*, **7**, 125-145.

Lopes Jr., I, Pizzato, L.G., Menezes, R.G. and Silva, L.C. (1992). Geoquímica do Granito Santa Helena na Folha Pontes e Lacerda M.T. *Anais do 37 Congresso Brasileiro de Geologia*, 52-65.

Macambira, M.J.B. and Lafon, J.M. (1995). Geocronologia da Província Mineral de Carajás: Síntese dos dados e novos desafios. *Boletim Museu Paraense Emílio Goeldi*, 263-288.

Macambira, M.J.B., Lafon, J.M., Dall'Agnol, R., João, X.S.J. and Costi, H.T. (1990). Geocronologia da granitogênese da Província Amazônia Central brasileira: Uma revisão. *Revista Brasileira de Geociências*, **20**, 258-266.

Macambira, M.J.B., Lafon, J.M. and Pidgeon, R. (1998). Crescimento crustral arqueano registrado em zircões de sedimentos da região de Rio Maria, Província de Carajás, Pará. *XL Congresso Brasileiro de Geologia Abstract bull.*, 55 p.

Macambira, E.M.B. (1996). Aspectos geológicos e potencial mineral do Complexo Máfico-ultratramáfico da Serra da Onça - Sul do Pará. *Simpósio Geológico da Amazoniania, V, Anais. Belém, Brazil*, 79-82.

Macambira, E.M.B. and Vale, A.G. (1997). *Programa Levantamentos Geológicos Básicos do Brasil.* Programa Grande Carajás. Folha SB.22-Y-B São F. do Xingu. Belém, CPRM.

Macambira, J.B., Kotschoubey, B., Santos, M.D., Moura, C.A.V. and Ramos, J.F.F. (1986). Estratigrafia e mineralizações primárias de ouro na aba sul do sinclinório de Gradaús. *Congresso Brasileiro de Geologia, XXXIV, Anais. Goiânia, Brazil*, **5**, 1956-1968.

Macambira, M.J.B. (1992). *Chronologie U-Pb, Rb-Sr, K-Ar et croissance de la croûte continentale dans l'Amazonianie du sud-est, exemple de la région de Rio Maria, Province de Carajás, Brésil.* Montpellier, Université de Montpellier II, PhD. Thesis, 212 p.

Machado, N., Lindenmayer, Z., Krogh, T.E. and Lindenmayer, D. (1991). U-Pb geochronology of archaean magmatism and basement reactivation in the Carajás area, Amazonian shield, Brazil. *Precambriam Research*, **49**, 329-354.

Maclelland, J.M. (1989). Crustal growth associated with anorogenic, Mid-Proterozoic, anortosite massifs, in northeastern North America. *Tectonophysics*, **161**, 331-341.

Matos, J.B. (1992). *Contribuição à geologia da parte meridional do Cráton Amazônico: Região do Rio Alegre.* M.T. Master Dissertation. IG-USP, 108 p.

Matos, J.B., Sousa, M.Z.A., Ruiz, A.S., Silva, C.H. and Sousa, F.J. (1986). Caracterização preliminar do Granodiorito Água Clara do Distrito de Farinópolis (Araputanga-MT). *XXXIX Congresso Brasileiro de Geologia. Anais*, **2**, 62-64.

Matos, J.B. and Schorscher, J.H.D. (1997a). Tendências geoquímicas da Sequência Vulcano-sediemntar do Rio Alegre-MT. *Anais do VI Simpósio de Geologia do Centro Oeste. Cuiabá, Outubro de 1997*, 26-28.

Matos , J.B. and Schorscher, J.H.D. (1997b). Geologia da região do Rio Alegre-MT. *Anais do VI Simpósio de Geologia do Centro Oeste. Cuiabá, Outubro de 1997*, 85-87.

Marot, A. (1988). *Guyane Sud. Carte Géologique de la France a 1/ 500000.* Département de la Guyane. Ed. BRGM. Notice explicative, 86 p.

McReath, I. and Faraco, M.T.L. (1997). Sm-Nd and Rb-Sr systems in part of the Vila Nova metamorphic suite, northern Brazil. In: *South American Symposium on Isotpoe Geology, 1, Campo de Jordão. Extended abstracts*, 194-196.

Medeiros, H., Gastal, M.C.P., Dall'Agnol, R. and Souza, Z.S. (1987). Geology of the Rio Maria area (eastern Amazonianian region - Brazil): an example of Archaean granite-greenstone terrane intruded by anorogenic granites of Middle Proterozoic ages. *Final Meeting 204-Project, IUGS-UNESCO. Extended Abstract, Carajás, Brazil*, 97-109.

Medeiros Filho, C.A. and Meireles, E.M. (1985). Dados preliminares sobre a ocorrência de cromita na área Luanga. *Simpósio Geológico da Amazoniania, II, Atas. Belém, Brazil*, **3**, 90-96.

Menezes, R. G. de, Lopes, I. and Bezerra, J.R.L. (1993). *Folha Pontes e Lacerda 1:100.000. Carta Geológica e Texto Explicativo.* Programa de Levantamentos Básicos. CPRM-DNPM. 176 p.

Milési, J.P., Egal. E.E., Ledru, P., Vernhet, Y., Thièblemont, D., Cocherie, A. Tegyey, M., Martell-Jantin, B. and Lagny, P. (1995). Les mineralisations du Nord de la Guyane Française, dans leur cadre geologique. *Chronique de la Recherche Minière*, **518**, 5-58.

Montalvão, R.M.G., de and Tassinari, C.C.G. (1984). Geocronologia pré-cambriana do Território Federal do Amapá (Brasil). In: *Simp. Amaz., 2, Manaus. Anais. Manaus, MME-DNPM*, 54-57.

Monteiro, H., Macedo, P.M., Silva, M.D., Moraes, A.A. and Marcheto, C.M.L. (1986). O greenstone belt do Alto Jaurú. *XXXIV Congresso Brasileiro de Geologia*, **2**, 630-646.

Montgomery, C.W. (1979). Uranium-lead geochronology of the Archean Imataca Series, Venezuelan Guyana shield. *Contribuition to Mineralogy and Petrology*, **69**, 167-176.

Montgomery, C.W. and Hurley, P.M. (1978). Total rock U-Pb and Rb-sr systematics in Imataca Series, guyana Shield, Venezuela. *Earth and Planetary Science Letters*, **39**, 281-290.

Mougeot, R., Respaut, J.P., Briqueau, L., Ledru, P., Milesi, J.P., Macambira, M.J.B. and Huhn, S.B. (1996). Geochronological constraints for the age of the Aguas Claras formation (Carajás Province, Pará, Brazil). *Congresso Brasileiro de Geologia, XXXIX, Boletim Res. Expandidos Salvador, Brazil*, **6**, 579-581.

Moura, C.A.V., Gorayeb, P.S. and Matsuda, N.S. (1999). Geocronologia Pb-Pb em zircão do riolito Vila Raiol, formação iriri, Sudoeste do Pará. *VI Simpósio de Geologia da Amazoniania, Extended Abstract Bulletin*, 475-477.

Nogueira, A.C.R., Truckenbrodt, W. and Pinheiro, R.V.L. (1995). Formação Águas Claras, Pré-Cambriano da Serra dos Carajás. Redescrição e redefinição. *Boletim Museu Paraense Emílio Goeldi*, 7, 177-197.

Norcross, C.E., Davis, D.W. and Spooner, E.T.C. (1998). U-Pb geochronology of the Omai intrusion hosted Au-quartz vein deposit and host rocks, Guyana, South America. In: *GSA Annual Meeting, Toronto, Ontario*, p. A-127.

Oberthur, TH., Angermeier, H.O., Hirdes, H., Hopnner, M., Schmidt Mumm, A., Vetter, U. and Weiser, TH. (1994). Gold Mineralization in the Ashanti belt of Ghana and it is relation to the crustal evolution of the terrain. *Workshop "Metallogeneis of gold in Africa", technical cooperation, project 832063 27, BGR Report n 112049*, 50-57.

Oliva, L.A. (1979). *Ocorrências Minerais na Folha Cuiabá (SD.21).* Relatório de Viagem. Goiânia, DNPM, 18 p.

Oliveira, J.R., Silva Neto, C.S. and Costa, E.J.S. (1994). *Programa Levantamentos Geológicos Básicos do Brasil.* Programa Grande Carajás. Folha SB.22-X-C Serra Pelada. Belém, CPRM.

Olszweski, W.J., Gaudette, H.E. and Santos, J.O.S. (1989). Isotopic age results from the shield areas of western Brazil. *Geol. Soc. America Meeting. Abstract with Programs*, 20 p.

Payolla, B.L., Kozuch, M., Leite, W.B. Jr., Fetter, A., Bettencourt, J.S. and Van Schmus, W.R. (1998). *Proterozoic geological evolution of the central-eastern part of the Rondônia Tin Province (Brazil), infered from U/Pb and Sm/Nd isotopic data.* University of Wisconsin. Open File Report 1998-10, 165 p.

Pessoa, M.R., Santiago, A.F., Andrade, A.F., Nascimento, J.O., Santos J.O., Oliveira, J.R., Lopes, R.C. and Prazeres, W.V. (1977). *Projeto Jamanxin*. Relatorio Final, CPRM/DNPM, 9 volumes.

Pidgeon, R., Macambira, M.J.B. and Lafon, J.M. (1998). Datação U-Pb de estruturas primárias e secundárias de zircões de granulitos do Complexo Pium, Província de Carajás, Pará. *XL Congresso Brasileiro de Geologia, Abstract bull.*, 56 p.

Pimentel, M.M. and Machado, N. (1994). Geocronologia U-Pb dos terrenos granito-greenstone de Rio Maria, Pará. *XXXVIII Congresso Brasileiro de Geologia Ext. Abstract*, **2**, 390-391.

Pinheiro, S.S., Fernandes, P.E.C.A., Pereira E.R., Vasconcelos, E.G., Pinto, A.C., Montalvão R.M.G., Issler, R.S., Dall'Agnol R., Teixeira, W. and Fernandes, C.A.C. (1976). Geologia. In: *Levantamento de Recursos Naturais, Folha NA-19 Pico da Neblina, Projeto RADAM, DNPM/MME*, 11, 19-137.

Pinho, M.A. de S.B. (1990). *Geologia, Petrologia, e Geoquímica das Rochas Ocorrentes ao Longo do Rio Aguapeí- Sudoeste do Cráton Amazônico-Pontes e Lacerda-MT. Porto Alegre, URGS.* Dissertação de mestrado, 199 p.

Pinho, M.B. and Fyfe, W.S. (1994). A evolução do conhecimento geoquímico das vulcânicas máficas-ultramáficas tholeiíticas da formação Mata Preta, greenston belt do Alto Jauru-Mato Grosso. In: *38 Congresso Brasileiro de Geologia*, **3**, 141-143.

Pinho, M.B., Fyfe, W.S. and Pinho, M.A.S.B. (1997). Early Proterozoic evolution of the Alto Jauru greenstone belt, southern Amazonianian craton, Brazil. *International Geology Review*, **39**, 220-229.

Priem, H.N.A., Bon, E.H., Verdumen, E.A.Th. and Bettencourt, J.S. (1989). Rb-Sr chronology of Precambrian crustal evolution in Rondônia, western margin Brasilian craton. *Journal of South America Earth Sciences*, **2**, 163-170.

Priem, H.N.A., Boelrijk, N.A.I.M., Hebeda, E.H., Kuyper, R.P., De Roever, E.W.P., Verdurmen, E.A.Th., Verschure, R.H. and Wilens, J.B. (1978). How old are the supposedly Archean charnockitic granulites in the guyana Shield basement of western Suriname (South America)? *USGS, Open File rep.* **78-701**, 341-343.

Priem, H.N.A., Hebeda, E.H., Boelrijk, N.A.I.M. and Verschure, R.H. (1968). Isotopic age determinations on surinam rocks, 3. Proterozoic and Permo-Triassic basalt magmatism in the Guiana shield. *Geologie en Mijnbouw*, **47**, 17-20.

Priem, H.N.A., Boelrijk, N.A.I.M., Hebeda, E.H., Kroonenberg, S.B., Verdurmen E.A.Th. and Verschure, R.H. (1973). Age of the Roraima formation in northeastern South America: evidence from isotopic dating of Roraima pyroclastic volcanic rocks in Suriname. *Geological Society of America Bulletin*, **84**, 1677-1680.

Priem, H.N.A., Boelrijk, N.A.I.M., Hebeda, E.H., Verdurmen E.A.Th. and Verschure, R.H. (1977). Isotopic ages in the high-grade metamorphic Coeroeni group, southwestern Suriname. *Geologie en Mijnbouw*, **56**, 155-160.

Priem H.N.A., de Roever, E.W.F. and Bosma, W. (1980). A note on the age of the Paramaka metavolcanics in northeastern Suriname. *Geologie en Mijnbouw*, **59-2**, 171-173.

Ramos, J.F.F., Moura, C.A.V., Melo, C.F., Pereira, J.L., Serique, J.S.C.B. and Rodrigues, R.M. (1984). Uma discussão sobre sequências sedimentares tidas como Formação Rio Fresco, Sudeste do Pará. *Congresso Brasileiro de Geologia, XXXIII, Anais. R. Janeiro, Brazil*, 862-872

Rämö, O.T. and Haapala, I. (1995). One hundred years of rapakivi granite. *Mineralogy and Petrology*, **52**, 129-185.

Rämö, O.T., Huhma, H. and Kirs, J. (1996). Radiogenic isotopes of the Estonian and Latvian rapakivi granites suites: new data from the cancealed Precambriam of the East European Craton. *Precambriam Research*, **79**, 209-226.

Rivers, T. (1997). Lithotectonic elements of the Grenville Province: review and tectonic implications. *Precambrian Research*, **86**, 117-154.

Rizzoto, G.J. (1999). *Petrologia e Ambiente Tectônico do Grupo Nova Brasilândia-RO.* Master Thesis, UFRS, 136 p.

Rizzotto, G.J., Scandolara, J.E. and Quadros, L.E.S.M. (1996). Aspectos gerais da associação mangerito-charnockito-granito, MCG da porção oriental do Estado de Rondônia. In: *Congr. Bras. Geol., 39, Salvador, Brasil, 1996. Anais...Salvador, Sociedade Brasielira de Geociências*, 1, 35-37.

Rizzoto, G.J., Lima, E.F., Chemale Jr., F. and Luft, J.L. (1999a). Caracterização do Cinturão transpressivo Rio Branco do sudoeste de Rondônia. *VI Simpósio da Geologia da Amazônia, Manaus. Anais*, 315-318.

Rizzoto, G.J., Lima, E.F., Chemale Jr., F. and Luft, J.L. (1999b). Acresção continental do Esteniano no sudeste de Rondônia: Implicações tectônicas. *VI Simpósio da Geologia da Amazônia, Manaus. Anais*, 319-322.

Rizzotto, G.L., Chemale Jr., F., Lima. E.F., Van Schmuss, R. and Fetter, A. (1999c). Dados isotópicos Sm-Nd e U-Pb das rochas da Sequência metaplutono-vulcanossedimentar Nova Brailândia (SMNB) - RO, (1999). *VI Simpósio de Geologia da amazonia, Ext. Abstract Bull.*, 490-493.

Rodrigues, E. S., Lafon, J.M. and Scheller, T. (1992). Geocronologia Pb-Pb da Província Mineral de Carajás: primeiros resultados. *Congresso Brasileiro de Geologia, XXXVII, Res. S. Paulo, Brazil*, **1**, 183-184.

Ruiz, A.S. (1992). *Contribuição à Geologia do Distrito da Cachoeirinha, MT.* São Paulo. IG-USP, Master Dissertation, 98 p.

Saes, G.S. (1999). *Evolução Tectônica e Paleogeográfica do Aulacógeno*

Aguapeí (1.2-1.0 Ga) e dos Terrenos do seu Embasamento na Porção Sul do Cráton Amazônico. Doctored Dissertation, USP, São Paulo, 135 p.

Saes, G.S., Leite J.A.D. and Weska, R.K. (1984). Geologia da Folha Jauru (SD-21-Y-C-III): Uma Síntese dos Conhecimentos. In: *33 Congr. Bras. de Geol. Rio de Janeiro*, **5**, 2193-2204.

Saes, G.S., Leite, J. and Alvarenga, C.J.S. (1992). Evolução tectono-sedimentar do Grupo Aguapeí, Proterozóico Médio na porção meridional do Cráton Amazônico: Mato Grosso e oriente Boliviano. *Revista Brasileira de Geociências*, **23**, 31-37.

Saes, G.S. and Fragoso Cesar, A.R.S. (1996). Acresção de terrenos mesoproterozóicos no SW da Amazônia. In: *39 Congresso Brasileiro de Geologia*, 348 p.

Sadowski, G.R. and Bettencourt, J.S. (1996). Mesoproterozoic tectonic correlations between eastern Laurentia and the western border of the Amazonianian Craton. *Precambriam Research*, **76**, 213-227.

Santos, J.O.S. (1982). Granitos proterozóicos da plataforma Amazônica. *Cong. Latino America Geol., II, Actas. B. Aires, Argentina*, 99-112.

Santos, J.O.S. and Reis Neto, J.M. (1982). Algumas idades de rochas graníticas do Cráton Amazônico. *Congresso Brasileiro de Geologia, XXXII, Res. Salvador, Brazil*, **1**, 339-348.

Santos, J.O.S., Hartmann, L.A. and Gaudette, H.E. (1997). Reconnaissance U/Pb in zircon, Pb/Pb in sulfides and review of Rb/Sr geochronology in the Tapajós gold Province, Pará/ Amazonianas States, Brazil. *South American Symposium on Isotope Geology. Ext. Abstract*, 280-282.

Santos, M.V., Tassinari, C.C.G., Souza Filho, E.E., Teixeira, W., Ribeiro, A.C.O., Payolla, B. and Vasconi, A. (1988). Litoestratigrafia das rochas precambrianas na Bacia do medio Rio Xingu, Altamira, Para. *VII Congresso Latino Americano de Geologia, Extended Abstracts*, 363-377.

Santos, J.O.S., Reis, N.J., Hartmann, L.A., MacNaugthon, N.J. and Fletcher, I.R. (1999). Associação anortosito-charnockito-rapakivi do Calimiano do norte do Craton Amazônico, Estado de Roraima: Evidências obtidas por geocronologia U-Pb (SHRIMP) em zircão e baddeleyta. *VI Simpósio de Geologia da amazonia, Ext. Abstract Bull.*, 502-505.

Sato, K. (1998). *Evolução Crustal da Plataforma Sul americana, com base na geoquimica isotopica Sm-Nd*. Doctoral thesis, Institut of Geosciences, University of São Paulo.

Sato, K. and Tassinari, C.C.G. (1997). Principais eventos de acreção continental no Cráton Amazônico baseados em idade modelo Sm-Nd, calculada em evoluções de estágio único e estágio duplo. In: *Contribuições à Geologia da Amazônia*, eds. Costa, M.L. and Angélica, R.S., pp. 91-142. Belém, Sociedade Brasileira de Geociências.

Scandolara, Amorim, J.L.J. (1999). A Faixa Móvel Guaporé, sua definição e inserção no contexto geotectônico do SW do Cráton Amazônico. *VI Simpósio da Geologia da Amazônia, Manaus, Anais*, 278-281.

Scandolara, J.. Rizzoto, G.J. and Amorim, J.L. (1999b). Evolução Proterozóica de Rondônia: Cronologia dos eventos tectônicos. *VII Simposium Nacional de Estudos Tectônicos (SNET)*, 24-27.

Scandolara, Amorim, J.L J. Rizzoto, G.J., Quadros, M.L.E.S. and Bahia, R.B.C. (1999c). Compartimentação tectônica-estratigráfica pré-Cambriana de Rondônia: Subsídios para modelos evolutivos. *VI Simpósio da Geologia da Amazônia, Manaus, Anais*, 282-285.

Schobbenhaus, C., Campos, D.A., Derze, G.R. and Asmus, H.E. (1981). *Geological Map of Brazil, scale 1:2 500 000*. DGM/ DNPM, Brasilia.

Schobbenhaus, C., Hoppe, A., Lork, A. and Baumann, A. (1994). Idade U/Pb do magmatismo Uatumã no norte do Cráton Amazônico, Escudo da Guianas (Brasil): primeiros resultados. *XXXVIII Congresso Brasileiro de Geologia Extended Abstract*, **2**, 395-397.

Silva, C.R., Bahia, R.B.C. and Silva, L.C. (1992). Geologia da região de Rolim de Moura- sudeste de Rondônia. In: *Congr. Bras. de Geol., 37, São Paulo, 1992, Boletim de resumos expandidos, São Paulo, SBG*, **1**, 152-153.

Silva, G.G., Lima, M.J.C., Andrade, A.R.F., Issler, R.S. and Guimarães, (1974). Geologia, geomorfologia, pedologia, vegetação e uso potencial da terra. In: *Projeto RADAM-BRASIL. Folhas SB-22 Araguaia e parte da SC-22 Tocantins. Rio de Janeiro, DNPM/MME*, 143 p.

Silva, G.H., Leal J.W.L., Montalvão, R.M.G., Bezerra, P.E.L., Pimenta, O.N.S., Tassinari, C.C.G. and Fernandes, C.A.C. (1980). Geologia. In: *Projeto RADAMBRASIL. Folha SC-21 Juruena. Rio de Janeiro, DNPM/MME*, 21-116.

Snelling, N.J. and McConnell, R.B. (1969). The geochronology of Guyana. *Geologie en Mijnbouw*, **48-2**, 201-213.

Souza, A, E. P. and Hildred, P.R. (1980). Contribuição ao estudo da geologia do Grupo Aguapeí, Mato Grosso. In: *31 Congr. Bras. de Geol.*, **2**, 587-598.

Souza, Z.S. (1994). *Geologia e petrogênese do "Greenstone Belt" de Identidade, Implicações sobre a evolução geodinâmica do terreno granito-greenstone de Rio Maria, SE do Pará*. Belém Univ. Fed. Pará. Brazil. PhD Thesis, 621 p.

Souza, Z.S., Dall'Agnol, R., Oliveira, C.G. and Huhn, S.R.B. (1997). Geochemistry and petrogenesis of metavolcanic rocks from Archean greenstone belts: Rio Maria region (Southeast Pará, Brazil). *Revista Brasileira de Geociências*, **27**, 169-180.

Spooner, C.M., Berrangé, J.P. and Fairbairn, H.W. (1971). Rb-Sr whole-rock age of the Kanuku complex, Guyana. *Geological Society of America Bulletin*, **82**, 207-210.

Sudblad, K. and Ahl, M. (1997). Contrasting metallogenetic features of Palaeproterozoic granitoids in the Svecofennian domain, northern Europe. Second International Symposium on Granites and associated Mineralizations. In: Ferreira, V.P. and Sial, (eds). *A.N. ISGAM II, Salvador, Bahia, Brazil, 1997. Extended Abstract and Program*, 89-90.

Tassinari, C.C.G. (1981). *Evolução tectônica da Província Rio Negro-Juruena na região Amazônica*. Dissertação de MSc., Instituto de Geocências, Universidade de São Paulo, São Paulo, 2 vol., 99 p.

Tassinari, C.C.G. (1996). *O Mapa Geocronológico do Craton Amazônico no Brasil: Revisão dos Dados Isotópicos*. Tese de Livre Docência. Instituto de Geociências, Universidade de São Paulo. São Paulo, Pp. 139.

Tassinari, C.C.G. (1984). A porção ocidental do craton Amazônico: evidências isotópicas de acreção continental no Proterozóico Médio. In: *Symp. Amazônico, 2, Manaus, Brasil, 1984. Atas...Manaus, DNPM/MME*, 439-446.

Tassinari, C.C.G., Siga Jr., O. and Teixeira, W. (1984). Épocas metalogenéticas relacionadas a granitogênese do craton Amazônico. In: *Congr. Bras. Geol., 33, Rio de Janeiro, 1984. Anais... Rio de Janeiro, Sociedade Brasileira de Geociências*, **6**, 2963-2977.

Tassinari, C.C.G, Teixeira, W. and Siga Jr., O. (1978). Considerações crono-estratigráficas da região das Chapadas do Cachimbo e Dardanelos. *XXX Congresso Brasileiro de Geologia*, **1**, 477-490.

Tassinari, C.C.G., Cordani, U.G., Nutman, A.P., Van Schmus, W.R., Bettencourt, J.S. and Taylor, P.N. (1996). Geochronological systematics on basement rocks from the Rio Negro - Juruena Province (Amazonianian Craton) and tectonic implications. *International Geology Review*, **38**(2), 1161-175.

Tassinari, C.C.G. and Macambira, M.J.B (1999). Geochronological Provinces of the Amazonianian Craton. Episodes, **22**(3), 174-182.

Tassinari, C.C.G., Cordani, U.G., Correia, C.T., Nutman, A.P., Kinny, P., Dias Neto, C. (1999). Dating of granulites by shrimp U-Pb systematics in Brazil: constraints for the age of the metamorphism of Proterozoic Orogenies. *Anais do II South American Symposium on Isotope Geology*, 234-238.

Teixeira, J.B.G. and Eaggler, D.H. (1994). Petrology, geochemistry and tectonic setting of Archean basaltic and dioritic rocks from the N4 iron deposit, Serra dos Carajás, Pará, Brazil. *Acta Geológica Leopoldinense*, **40**, 71-114.

Teixeira, N.P., Bettencourt, J.S., Moura, C.A.V. and Dall'Agnol, R. (1998). Pb-Pb and Sm-Nd constraints of the Velho Guilherme Intrusive Suite and volcanic rocks of the Uatumã Group, South-Southeastern Pará, Brazil. *ICGP 462, Field and Proc. Vol.*, 178-180.

Teixeira, W., Kawashita, K., Taylor, P.N., Ojima, S.J. and Viera, A.G. (1985). Reconhecimento geocronológico da Guyana francesa: novos dados, integração e implicações tectônicas. *Simpósio Geológico da Amazoniania, 2. Belém. Anais. Belém. Sociedade Brasileira de Geociências*, **1**, 194-207.

Teixeira, W. (1978). Significação tectônica do magmatismo básico e alcalino na região Amazônica. *Congresso Brasileiro de Geologia*, **1**, 477-490.

Teixeira, W. and Tassinari, C.C.G. (1984). Caracterização geocronológica da Província Rondoniana e suas implicações geotectônicas. *2 Symposium Amazônico, Actas*, 87-102

Teixeira, W., Ojima, S.K. and Kawashita, K. (1984). A evolução geocronológica de rochas metamórficas e ígneas da faixa móvel Maroni-Itacaiúnas na Guiana Francesa. *2 Symposium Amazônico, Actas*, 75-86.

Teixeira, W., Tassinari, C.C.G., Cordani, U.G. and Kawashita, K. (1989). A review of the geochronology of the Amazonianian Craton: Tectonic Implications. *Precambrian Research*, **42**, 213-27.

Teixeira, W., Tassinari, C.C.G., Szabó, G.J., Mondim, M., Sato, K., Santos, A.P., Siso, C.S. (1999). Sm-Nd Constrains on protolith age of the Archean Imataca Complex, Venezuela. *II South American Symposium on Isotope Geology, Actas*, 136-138.

Tosdal, R.M. (1996). The Amazonian-Laurentian connection as viwed from the Middle Proterozoic rocks in the Central Andes, western Bolivia and northern Chile. *Tectonics*, **15** (4), 827-842.

Trendall, A.F., Basei, M.A.S., De Laeter, J.R. and Nelson, D.R. (1998). SHRIMP zircon U-Pb constraints on the age of the Carajás formation, Grão Pará Group, Amazonian Craton. *Journal of South America Earth Science*, **11**, 265-277.

Tucker, R.D. and Gower, C.F. (1994). Anological framework for the Pinware Terrane, Grenville Province, southern Labrador. *The Journal of Geology*, **102**, 67-78.

Van Schmus, W.R Geraldes, M.C., Kozuch, M., Fetter, A.H., Tassinari, C.C.G. and Teixeira, W. (1998). U/Pb and Sm/Nd constraints on the age and origin of proterozoic crust in southwestern Mato Grosso, Brazil: Evidence for a 1450 Ma magmatic arc in SW Amazoniania. *Internation Symposium on Tectonics. Ouro Preto-MG. Abstract Volume*, 121-125.

Van Schmus, W.R Geraldes, M.C., Kozuch, M., Fetter, A.H., Tassinari, C.C.G. and Teixeira, W. (1999). Jauru Terrane: A late Paleoproterozoic orogen in SW Amazonian craton, Mato grosso State, Brazil. *European Union Geology bianual Meeting. Streisburg, France. Abstract Volume*, 56 p.

Vanderhaeghe, O., Ledru, P., Thiéblemont, D., Egal, E. Cocherie, A., Tegyey, M. and Milési, J.P. (1998). Contrating mechanism of crustal growth. Geodynamic evolution of the Paleoproterozoic granite-greenstone belts of French Guiana. *Precambriam Research*, **92**, 165-193.

Vasquez, M.L., Klein, E.L., Quadros, M.L.E., Bahia, R.B.C., Santos, A., Ricci, P., Sachett, C.R., Silva, C.M.G. and Macambira, M.J.B. (1999). Magmatismo Uatumã, na Província Tapajós – Novos dados geocronológicos. *VI Simpósio de Geologia da amazonia, Ext. Abstract Bull.*, 471-474.

Vilzeuf, D., Clemens, J.D., Pin, C. and Moinet, E. (1990). Granites, granulites and crustal differentiation. In: *Granulites and Crustal Evolution*, eds. D. Vilzeuf and Vidal Ph. Kluwer, Dordrech, 59-85.

Vignol, L.M. (1987). *Etudes geochimiques des granulites du Bresil et de la zone D'Ivree: Les elements (K, Rb, Sr, Sm, Nd) et les isotopes radiogeniques (Sr et Nd).* Diplome d'Études Approfondies de Geochimie – Universite Paris VII. 23 p.

Wirth, K.R., Gibbs, A.K. and Olszewski, W.J. (1986). U-Pb ages of zircons from the Grão-Pará group and Serra dos Carajás Granite, Pará, Brazil. *Revista Brasileira de Geociências*, **16**, 195-200.

Wasteneys, H.A., Kamo, S.L., Moser, D., Krogh, T.E., Gower, C.F. and Owen, J.V. (1997). U-Pb geochronological constraints on the geological evolution of the Pinware terrane and adjacent areas, Grenville Province, southeast Labrador, Canadá. *Precambrian Research*, **81**, 101-128.

The São Luís Craton and the Gurupi Fold Belt

Georg R. Sadowski

The present day, erosional northern border of the Paleozoic Parnaíba Basin is represented by a crystalline basement high, which constitute the Urbano Ferrer Arch (Almeida, 1967). The crystalline rocks are considered to be the probable representatives of the rim of a reminiscent piece of the West African Craton (WAC) and its marginal Neoproterozoic mobile belt. Presently, one of the main questions to be answered concerns the relationship of the Gurupi Belt with its West-African counterparts: the Pharusian and Rokel River belts (Fig.1).

The first published accounts of the basement rocks of Gurupi Belt date from 1871 (*apud* Hasui *et al.*, 1984), but since then little work has been done to unravel the stratigraphy and geological evolution of the area due mainly to the poor quality of the exposures.

Gold deposits, known from colonial times (Lisboa, 1935; Souza, 1938), attracted the attention of the Brazilian Mines Bureau. Moura (1934) published one of the first formal reports on surveys in that region. It describes outcrops along the Gurupi River valley that cut through the marginal Gurupi Fold Belt and its neighbouring cratonic area. About one decade later, Oliveira and Leonardos (1943), in their classical book on Brazilian Geology, suggested a correlation between the Gurupi Belt and similar belts on the African continent, mainly the Pharusian Belt in Togoland. Based on K/Ar and some Rb/Sr dates obtained as the result of a large scale geochronological reconnaissance survey of Northeastern Brazil and Bullard's pre-drift continental reconstruction, Hurley *et al.* (1967) suggested a correlation between the São Luis Craton with the Liberian and Eburnean provinces of the West African Craton .

At approximately the same time, in a first published geochronological subdivision of the Precambrian of South America, Cordani *et al.* (1967) defined the São Luís Cratonic Nucleus. Later, Cordani and Almaraz (1969), and Cordani *et al.* (1974), improved on the local geochronological information based on samples collected by one of the authors along the margins of the Gurupi River. Cordani *et al.* (1974) speculated on the possible correlation of the Gurupi Belt with the Rockel River Series of Sierra Leone as defined by Allen (1969).

Further studies and surveys conducted by private enterprises and the universities, as well as systematic mapping at 1:250 000 scale by the CPRM - Geological Survey of Brazil in 1994 and 1995, refined the geological knowledge of this region. Some papers deal with local controversies or speculations about the Archean age of the rocks in the region (that has not yet been proven). Others deal with the tectonic nature of the Gurupi Belt (a shear zone to some, and the remains of an old orogen to others). Still others have concentrated on the large scale geophysical signatures of the main tectonic units (Costa *et al.*, 1996; Lesquer *et al.*, 1984; Hasui *et al.*, 1984).

Tectonic framework

Cratonic area

The whole region is relatively flat, with the exception of some hills next to the coastal area or isolated on the mainland plain. Weathering is deep, and Meso-Cenozoic sedimentary rocks usually occur at the surface. The northern part of the exposed basement (Fig.1) corresponds to the cratonic fragment, composed of an assemblage of accreted calc-alkaline plutons consisting of tonalite and trondhjemite with greenstone inliers intruded by late K-granites. Cordani and Sadowski (personal communication, 1974), presented K/Ar and Rb/Sr ages between 2.2 and 2.1 Ga mainly from hornblende derived from tonalitic granitoid plutons from the cratonic area (Fig. 1) and related them to the time of emplacement. By comparison, Panafrican cooling ages obtained in mica and whole-rock were restricted to the gneiss of the bordering Gurupi Belt. Adjacent to the coast, a 2.054 ± 0.064 Ga ($\lambda tot = 0.530 \times 10^{-9} a^{-1}$) whole-rock K/Ar cooling age was obtained in phyllite collected from an oil exploration borehole drilled near from Vizeu. This borehole revealed the presence of Mesoproterozoic low-grade metamorphites and caused speculation about the geographical extent and age of the Gurupi rocks. However, recent surface mapping of similar rocks in the Viseu region has shown that these metamorphites are related to the consolidation of the craton, and to gold-bearing greenstones in Nigeria. From the stratigraphic point of view, the cratonic tonalitic-trodhjemitic granitoid plutons and metamorphites were initially assigned to the Maracaçumé Complex and to the Archean, whereas the greenstone belt rocks and low-grade metamorphites were assigned to the Viseu and Igarapé de Areia formations. The K-rich plutons, which intrude the complex, were called the Tromaí Granitoid Suite and initially considered to be anorogenic. Some of the Rosário tonalitic granitoid bodies exposed near to the City of São Luís gave Pb/Pb ages in zircon c. 2.1 Ga (Gaudette *et al.*, 1996). All the cited age determinations as well as those illustrated in the Figure 1, are compatible with the 2.1 - 2.0 Ga plutonic accretion period of the West African Craton proposed by some authors.

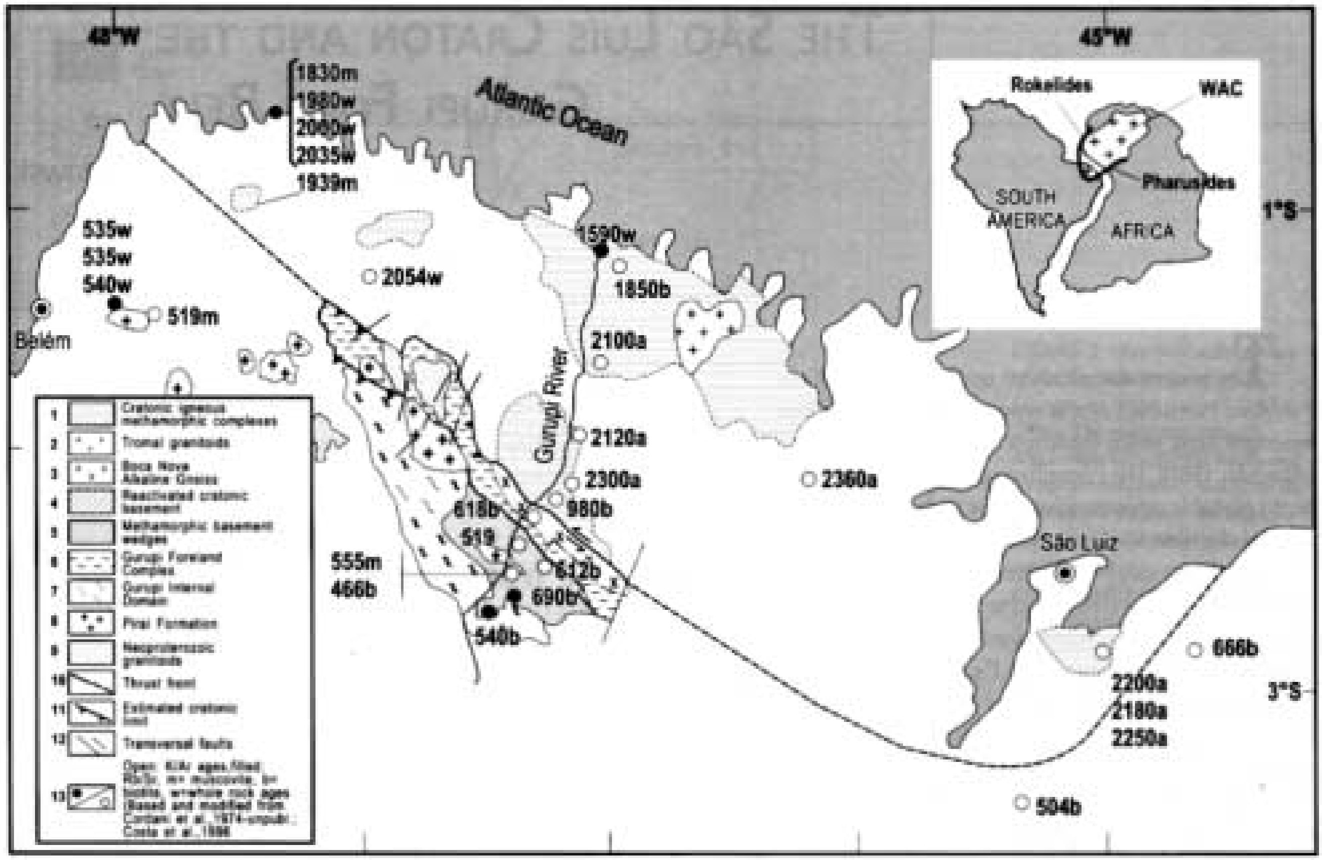


FIGURE 1 - Geological outline of the São Luis Craton and the Gurupi Belt with indication of Rb/Sr and K/Ar ages.

The Gurupi Mobile Belt

The southwestern border of the craton was apparently affected by left-lateral shearing of which the main zone of influence was initially known as the Chega-Tudo Formation and later, the Tentugal Shear Zone by Hasui *et al.* (1984).

The general trend of the Gurupi Belt is NW-SE in which there occur complex internal stuctures, but the real significance of them still remains unclear. The straightening zone which constitutes part of the cratonic border was considered to be originally left lateral (CPRM, 1995), and appears to have transpressional thrust components. The belt is composed of an assemblage of low to high-grade metamorphites formerly assigned to the Gurupi Group. It contains different types of gneiss (some of them are probably of the two-mica variety), collisional granite, remobilized cratonic fragments, kinzigite, schist, iron formation units, phyllite, quartzite and metavolcanic rocks. These units are still informally designated as Gurupi Schist, Santa Luzia Schist, Marajupema Kinzigite, Itamoari Cataclastic Gneiss, and Maria Suprema Gneiss. Molasse-like layers of the Piriá Formation consisting of greenish meta-arkose, some greywake, slate, sandstone and phyllite overlie the craton or are tectonically intercalated in the belt (Pitororó Graben). They are broadly folded along NNW-SSE axes.

Based on the metamorphic grade and the structural style of these units, Costa *et al.* (1996) proposed the subdivision of the Gurupi Mobile Belt into an internal domain or thrust belt and an external foreland belt.

Plutonic activity

Small batholiths and stocks of Brasiliano-Panafrican granitoid occur rarely at Bragança, Ney-Peixoto and Ourém with Rb/Sr ages *c.* 580 Ma and K/Ar cooling ages *c.* 520 Ma. Some of these appear to be syn-tectonic, whereas others are clearly late to post-orogenic. Moura (1934) mentioned the presence of volcanic rocks (dacite) in the region of the Gurupi Belt, which was subsequently confirmed.

A deformed alkaline pluton (Boca Nova) gave a Rb/Sr isochron age *c.* 723 Ma (Villas, 1982). This has been related to an extensional phase predating the deposition and deformation of at least some of the lithostratigraphic units assigned to the Gurupi Belt.

Discussion

The remnant cratonic fragment clearly illustrates that tectonic heritage is not a necessary parameter in the break-up and drift of continental masses. The correlation of the Gurupi Belt with the rocks of the Rokelides and Pharusides formations remains open. Stratigraphic surveys are incomplete on the Brazilian side mainly due to the highly weathered nature of the exposures. However, some similarities may be made with the Rokel River Series. The main deformation trends and metamorphic polarities and also the vergence of the Gurupi Mobile Belt seem to combine with the Rockelides, although strongly affected by a superimposed lateral shearing.

Andalusite, sillimanite and cordierite gneiss of the Rokelide Kasila Group may correspond to the Marajupema Kinzigites in the southern part of the Gurupi Belt. Post-metamorphic 546 to 505 Ma K/Ar cooling ages for the muscovite of the Marampa Formation were reported by Dallmeyer (1989). The tillite of the Tabe Formation at the base of the Rokelide Group and some clays associated with glaucogenic deposits give Rb/Sr ages between 630 and 595 Ma that apparently constrain the lower age of deposition of this belt. The age assemblage indicates the possible absence of the Panafrican I event (*c.* 650 Ma) in both belts, which seems not to be the case with the Pharusides where the deformation of the Tiririne Formation and syn-tectonic batholith emplacement occurred *c.* 660 Ma (Caby, 1989).

In the Rokelides rifting would be coincident with an intrusion of a syenite between 700 and 680 Ma, followed by a calc-alkaline arc related subduction at 680 to 650 Ma. A later collision probably took place in Senegal between 660 and 650 Ma (Ar^{39}/Ar^{40}), and also between 575 and 550 Ma (Dallmeyer and Lécorché, 1990; Villeneuve and Dallmeyer, 1988). So the opening of the Rokelides-Gurupi basins probably succeeded the first collision of the Pharusian terrane followed by their closure up to the end of the Lower Paleozoic.

The present picture, taking vergence, plutonism and the internal tectonic division of the belt into account seems to indicate an origin related to Neoproterozoic subduction directed to the NW with a left lateral collisional component.

REFERENCES

Allen, P.M. (1969). The Geology of part of an orogenic belt in western Sierra leone, West Africa. *Geologiesche Rundschau*, **58**(2), 588-620.

Almeida, F.F.M. (1967). *Origem e Evolução da Plataforma Brasileira.* Bol. 241 DGM/DNPM. Rio de Janeiro, 37 p.

Caby, R. (1989). Precambrian Terranes of Benin-Nigeria and Northeast Brazil and the Late Proterozoic south Atlantic fit. *Geological Society of America, Special Paper,* **230**, 145-158.

Cordani, U.G. Melcher, G.C. and. Almeida, F.F.M. de (1967). Outline of precambrian Geochronology of South America. *Canadian Journal of Earth Sciences*, **5**, 629-632.

Cordani, U.G. and Almaraz, J.S.U. (1969) Delimitação entre as províncias geocronológicas precambrianas ao longo do rio Gurupi. In: *Congr. Brasileiro de Geologia 23. Salvador. Resumo das conferências e comunicações. SBG. Bol. Especial* 1.

Cordani, U.G., Sadowski, G.R. and Almaraz, J.S.U. (1974). Estudo Geocronológico do Craton de São Luíz. *Anais da Academia Brasileira de Ciências*, **46** (3/4), 702.

Costa, J.L. da, Almeida, H.G. de, Ricci, P.S. dos S.F. (1996). Compartimentação Tectonica do Orógeno Gurupi e das Janelas Pré-Cambrianas da Província Parnaíba. *V Simpósio de Geologia da Amazonia, Resumos Expandidos Belém, SBG/NO*, 290-293.

CPRM (1977). Projeto Gurupi. Relatório Final by Costa, J.L., Villas Boas, J.M., Pastana, J.M.N, Wanderley, V.J.R., Araujo, A.A., Frizzo, S.J. Convênio DNPM/CPRM. Belém. Report (unpublished).

CPRM (1994). *Folhas São Luiz SA23-Z-A/Curupupu SA23-X-C, Estado do Maranhão.* Programa de levantamentos Geológicos Básicos do Brasil. Escala 1:250.000. Org. Rodrigues et al. MME. 114 p.

CPRM (1995). *Folhas Turiaçú SA23-X-D/ Pinheiro SA23-Y-B.* Estados do Pará e Maranhão. Escala 1:250.000 Programa de Levantamentos Geológicos Básicos do Brasil. Coord. Pastana. J.M. MME. 205 p.

Dallmeyer, R.D. (1989). A tectonic linkage between the Rodelide Orogen (Sierra Leone) and the St. Lucie metamorphic complex in the Florida subsurface. *Journal of Geology*, **97**(2), 183-195

Dallmeyer, R.D. and Lécorché, J.P. (1990). $^{40}Ar/^{39}Ar$ polyorogenic mineral age record within the southern Mauritanide orogen (M'Bout-Bakel Region) West Africa. *American journal of Sciences*, **290**, 1136-1168.

Gaudette, H., Moura, C.A.V., Abreu, F.A.M. de, Gorayeb, R.S.S. (1996). Dados geocronológicos Pb-Pb em zircão de granitoides da Suite Rosário. *Anais Congr. Bras. de Geologia, Salvador.* Vol. 2.

Hasui, Y., Matos de Abreu, F. de A. and Villas, R.N. (1984). Província Parnaíba In: *O Pré-Cambriano do Brasil*, eds. Almeida F.F.M. and Hasui, Y. Ed. Edgard Blücher, 378 p.

Hurley, P.M., Almeida, F.F.M. de, Melcher, G.C., Cordani, U.G., Rand, J.R., Kawashita, K., Vandoros, P., Pinson Jr., W.H. and Fairbain, H.W. (1967). Test of Continental Drift by means of radiometric ages. *Science*, **157**(3788), 495-500.

Lesquer, A., Beltrão, J.F. and Abreu, F.A.M. (1984). Proterozoic links between northeastern Brazil and West Africa, a plate tectonic model based on gravity data. *Tectonophysics*, **110**(1/2), 9-26.

Lisboa, M.A. (1935). *A bacia do Gurupi e as suas minas de ouro.* Bol. 7 Serv. Fom. Prod. Min. - DNPM 1938-1940. Rio de Janeiro.

Moura, P. (1934). *Rio Gurupy.* Serv. Geológico Mineralógico/DNPM 78. 44 p.

Oliveira, A.I. and Leonardos, O.H. (1943). *Geologia do Brasil.* Série Didática 2. Serviço de Informação Agrícola, Rio de Janeiro, Brasil.

Souza, H.C.A. de (1938). *Rochas da região de Bragança e Turiaçu (Pará\Maranhão).* Bol. 12. Serv. Fom. Prod. Min. Rio de Janeiro, 1-36.

Villas, R.N.N. (1982). Geocronologia de intrusões ígneas na Bacia do rio Guamá. Nordeste do Estado do Pará. *An. 1º Simp. Geol. da Amazonia, Belém, SBG*, **1**, 233-247.

Villeneuve, M. and Dallmeyer, R.D. (1988). Geodynamic evolution of the Mauritanide, Bessaride and Rokelide Orogens (West Africa). *Precambrian Research*, 37, 19-28.

ARCHEAN AND PALEOPROTEROZOIC TECTONIC EVOLUTION OF THE SÃO FRANCISCO CRATON, BRAZIL

W. Teixeira, P. Sabaté, J. Barbosa, C. M. Noce, and M. A. Carneiro

This synthesis aims to identify the tectonic environments and dynamics of the Archean and Paleoproterozoic terranes of the São Francisco Craton - SFC (Almeida, 1977), through the structural, kinematic, petrological and isotopic markers available in the areas studied. The information compiled from both the northern and southern parts of the Craton suggests that major events that took place during crustal evolution, resulted from juvenile accretion, as well as intercontinental and intracontinental collision. The work was co-ordinated by W. Teixeira, and he also wrote, with C.M. Noce and M.A. Carneiro, the section on the southern part of the S. Francisco Craton. P. Sabaté and J. Barbosa were responsible for the manuscript on the northern sector of the SFC.

The São Francisco Craton, situated in the central-eastern part of South America, is the best exposed and the most easily accessible unit of the Precambrian Brazilian Shield. Considering the craton's boundaries on geophysical evidence (Motta *et al.*, 1981; Ussami, 1992), and the fold belts resulting from the Brasiliano/Pan-African Orogeny (*c.* 680 - 550 Ma) that completely surround it, the SFC covers nearly the entire State of Bahia and a large part of the State of Minas Gerais. The Archean and Paleoproterozoic basement exposures along the central part of the Craton were tectonically affected by a rift-thrust belt (1.8 to 1.2 Ga) with N-S elongated grabens and basins that became the site of deposition of the Paleo to Mesoproterozoic Espinhaço Supergroup. Along the NNW-SSE axis of the Craton the *c.* 1.75 Ga Paramirim polyphase province (Cordani *et al.*, 1992) is distinguished by its unusual Lagoa Real uranium-associated plutonism,

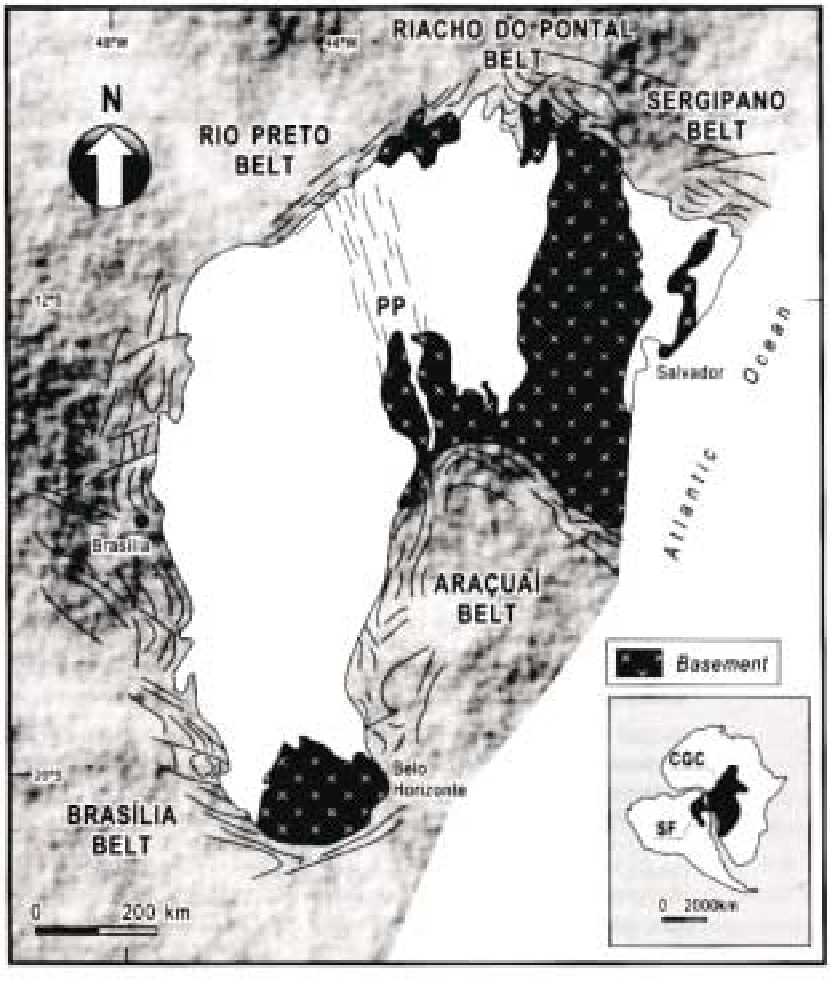


***FIGURE 1** - Sketch map showing the boundaries and the major structural units of the São Francisco Craton. Keys: Archean and Paleoproterozoic basement (crosses) and the structural framework of the surrouding Brasiliano belts. Also shown the Neoproterozoic and Phanerozoic covers (unpatterned). PP = Paramirim Province.*

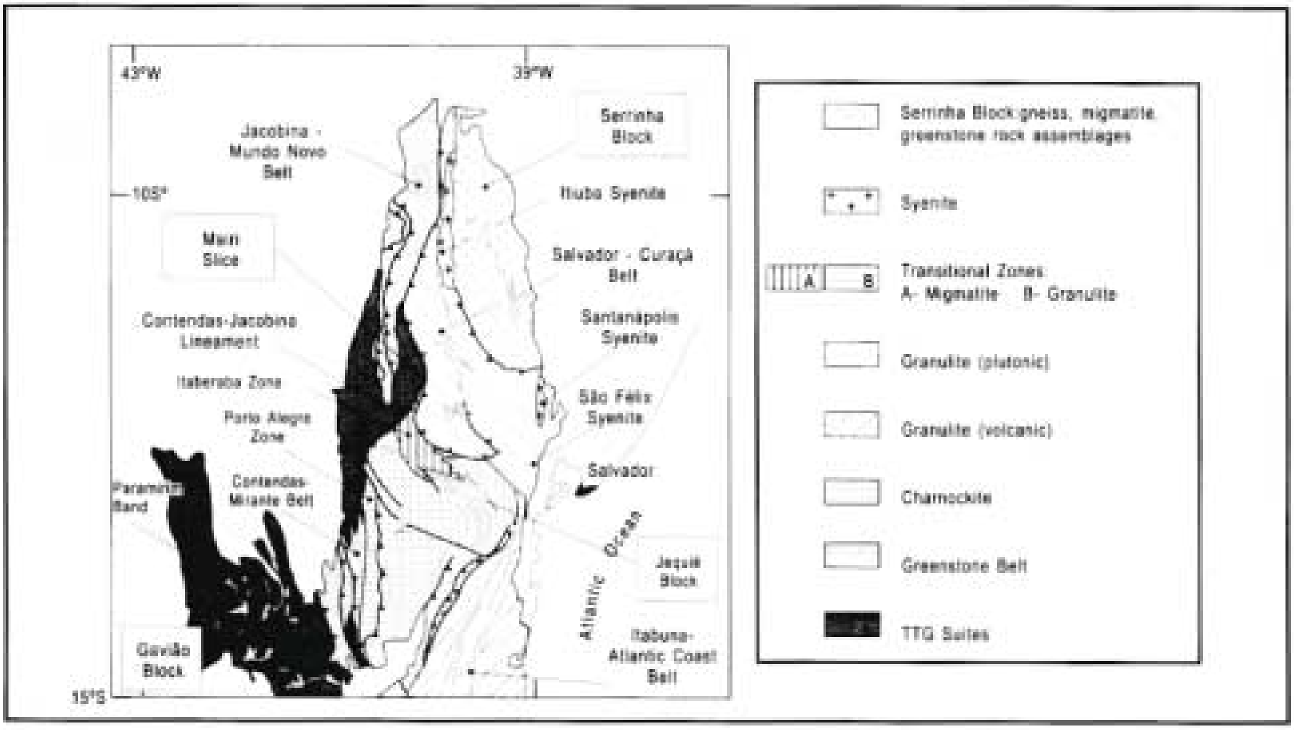


FIGURE 2 - Structural outline of the northern part of the SFC (after Sabaté and Cunha, 1998, unpublished). Keys: Transitional zones of the Jequié Block (A=Itaberaba region; B= retrometamorphosed charnockite of the Porto Alegre band). The charnockite field (Jequié) includes granulitic supracrustals. Plutonic and volcanic series refers to the Salvador-Curaça and Itabuna-Atlantic coastal belts, respectively. The greenstone belt field includes volcano-sedimentary sequences (Jacobina and Mundo Novo). See text for details.

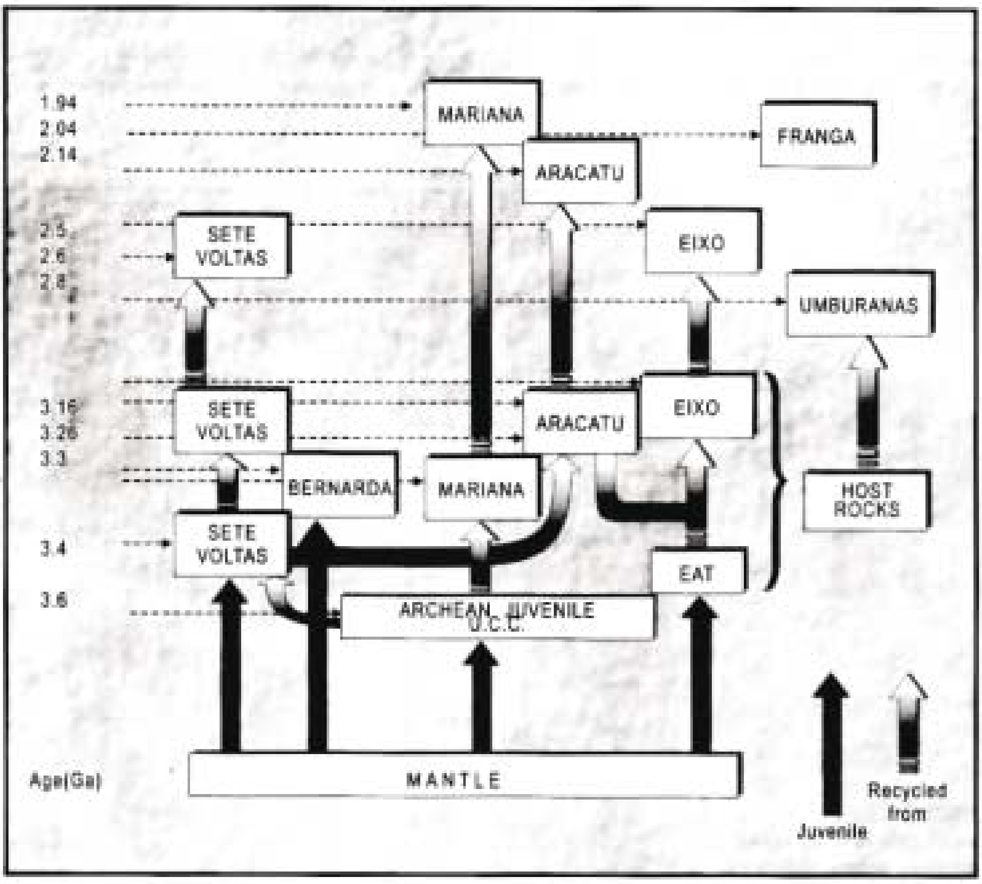


FIGURE 3 - Sketch diagram showing the genetic trend of plutonic TTG and calc-alkaline associations through geochemical modelling. See text for details. Black arrows correspond to the juvenile path from mantle sources to plutonism through a two-stage model. White-grading arrows correspond to palyngenetic path through recycling models by partial melting of the previous plutonic associations or either an Archean upper continental crust (U.C.C. - mean values from Condie 1993) or early Archean tholeiitic rock sources (E.A.T. - mean values from Condie,1981), or also by partial melting of the host rocks of the proper plutonic bodies. For example the 3.4 Ga old TTG of the Sete Voltas Massif participates as a solid source producing the melt generating the 3.1 Ga young TTG suite, this latter corresponding to the source of the 2.6 Ga grey granite. Analogous sources may produce also the Aracatu TTG Massif (compiled from Martin et al., 1991, 1997; Santos-Pinto et al., 1998).

as well as by the gneissic rocks and supracrustal sequences (northern part of the Espinhaço Supergroup), which were isotopically reset and deformed during the Brasiliano Orogeny (Cordani *et al.*, 1985; Alkmim *et al.*, 1993). Finally, the crystalline basement is partially covered elsewhere by Neoproterozoic (Bambuí Group) and Phanerozoic sedimentary rocks (Fig. 1).

The Archean and Paleoproterozoic medium to high-grade metamorphic terranes and granite-greenstone associations in the SFC crop out in two geographically distinct areas: the first and larger one to the N and NE in the State of Bahia, and the second in the southern part of the Craton in the State of Minas Gerais (Fig. 1). Geological correlation between the two regions is imprecise, and hence separate descriptions will be necessary (see sections 1 and 2) before making lithostructural, chronological and geodynamic comparisons; in particular, the northern part of the SFC, taking into account the evolutionary context of Western Gondwana that is tectonically associated with the West Congo-Gabon Craton of África, as supported by the comparable Archean and Paleo to Mesoproterozoic geological units. Therefore, the crustal evolution of the terranes will be herein briefly described in terms of their African counterparts.

The Northern part of the São Francisco Craton

Through the various studies carried out in the past in the northern part of the SFC, a great number of lithological units, as well as geological formations and groups have been defined for the Archean and Paleoproterozoic units. The foremost synthesis appears in the first edition of the geological map of Bahia (Inda and Barbosa, 1978), but the structural constraints and most of the radiometric data lead to confidence in the geodynamic models only at the end of the decade of the 1980s. The second edition of the Geological Map of Bahia (Barbosa and Dominguez, 1996) is the most recent base on which our geological knowledge of the northern part of the SFC had been placed.

This northern part of the SFC resulted from the accretion of Archean and Paleoproterozoic terranes through tectonic processes essentially achieved during the Transamazonian orogenic collision (2.14 - 1.94 Ga). The distribution of lithotypes, the high-grade nature of most rocks, the difficulty to establish lithostratigraphic and chronological correlation, lead us to the identification of lithotectonic units separated by tectonic discontinuities.

The resulting mosaic is articulated on a major convergent structure: the Contendas-Jacobina Lineament (Sabaté, 1991, 1996). This corresponds to a *c.* 800 km N-S tectonic imbrication, which assembled several previous (Archean to Paleoproterozoic) continental segments and contemporaneous Paleoproterozoic belts. It establishes the junction between two different domains: a western domain corresponding to the Gavião Block and an eastern domain that includes the Jequié and Serrinha blocks, and the Salvador-Curaçá and Itabuna-Atlantic belts (Fig. 2). Two volcano-sedimentary belts are included between the two domains along this junction zone: the Jacobina Belt to the N, and the Contendas-Mirante Belt to the S. In spite of the presence of tectonic slices (Mairí slice) of older terranes imbricated in these belts, the sedimentation in both took place during the Paleoproterozoic. In addition, both belts showed the successive tectonic increments of the Transamazonian orogenic collision, mirrored by the westward horizontal tectonics and by syn to late kinematic peraluminous granitic intrusions.

The Gavião Block, to the W, is mainly composed of typical tonalite-trondhjemite-granodiorite (TTG) plutonism associated with Archean greenstone belts. It represents the oldest well-dated and well-preserved Archean crust of South America. The Jequié Block, to the SE of the Contendas-Jacobina Lineament, demonstrates a diachronic behaviour and a distinct geotectonic environment compared to the TTG suites. On one hand it consists mainly of charnockite and enderbite corresponding to plutonism with strong calc-alkaline affinities, and on the other hand of granulite facies supracrustal rocks. If the protoliths have preserved an Archean age, the tectono-metamorphic history relates to the Transamazonian Orogeny.

The Salvador-Curaçá Belt, to the NE of the Contendas-Jacobina Lineament, shows tectonic contacts with a tectonic slice of the western domain (Mairí slice), and the Jacobina Belt. Some Archean inliers may be found within the belt, but it comprises mainly plutonic rocks developed during the Transamazonian shearing and, probably, pull-apart mechanisms. The Salvador-Curaçá Belt rocks are essentially metamorphosed in the granulite facies, but exhibit high amphibolite facies only in its northwestern part. The Itabuna-Atlantic Coast Belt, to the E of the Jequié Block, seems to prolong southwards the Salvador-Curaçá Belt. It is a high-grade metamorphic shear belt, the protoliths being of magmatic origin. The emplacement of the protoliths seems to have occurred earlier than the granulite metamorphism related to the Transamazonian Orogeny.

The Serrinha Block occurs to the E of the Salvador-Curaçá Belt (Fig. 2), and it consists of amphibolite facies orthogneiss and migmatite. These rocks constitute the basement of both the Itapicuru River and the Rio Capim greenstone belts. Finally, beyond the main features of these major lithostructural units we will present their diachronic evolution and their main chronological, petrogenetic and tectono-thermal constraints. Although most of these units had a polyphase history and although, independently of the age of their formation, all were involved in the Transamazonian collision, we summarize here the most relevant parameters of the Archean and Paleoproterozoic behaviour.

The Archean Crust

The Gavião Block

Under the Gavião Block, we associate the lithotectonic sub-unit described as the Paramirim Block (not shown in Fig. 2), in the northern part of the SFC. This sub-unit is the northwestern extension of the Gavião Block, thrust to the W over the Espinhaço Supergroup sequences, during the Brasiliano Orogeny. The Gavião Block crops out discontinuously on up to 500 km to the W of the Contendas-Jacobina Lineament from Jacobina to the Vitória da Conquista region. It is nearly 200 km wide in the S and appears largely hidden to the N by Mesoproterozoic and Neoproterozoic cover. Tectonic slices of the Gavião Block (Fig. 4) are imbricated into both the Contendas-Mirante (Sete Voltas and Mata Verde massifs) and the Jacobina Belts (Mairí slice).

Lithotypes and Petrogenesis

The Gavião Block is composed of TTG associations, remnants of supracrustal rocks (Umburanas-Brumado greenstone belt), and medium-grade gneiss-leptynite-amphibolite rocks.

Tonalite-trondhjemite-granodiorite associations

The oldest recognized TTG association is the Sete Voltas Massif (Marinho, 1991; Cordani *et al.*, 1985; Martin *et al.*, 1991, 1997) which represents a tectonic slice uplifted onto the Contendas-Mirante Belt (see above) due to the Transamazonian collision. It appears as a composite segment built in at least three successive accretional events dated by Rb/Sr, $^{207}Pb/^{206}Pb$ single-zircon evaporation and U/Pb SHRIMP methods, as follow: (i) Plutonic precursors of the old grey gneiss emplaced at *c.* 3.4 Ga. They show the composition of Archean TTG suites, and the geochemical modelling indicate they resulted from partial melting of an Archean tholeiite, leaving a hornblende garnet residue. (ii) Younger grey gneiss and porphyritic granodiorite intruded the old grey gneiss between 3.17 and 3.15 Ga. Their composition is intermediate between the TTG and calc-alkaline trends. Geochemical modelling shows that they are the products of partial melting of an older continental crust. (iii) Grey granite (granitic dykes) emplaced at *c.* 2.6 Ga corresponds to a late-stage magmatic event in the Sete Voltas Massif.

The Sete Voltas Massif rocks yield homogeneous Sm/Nd T_{DM} ages between *c.* 3.7 - 3.6 Ga, which are interpreted as the protolith age, or rather their contamination by even older pre-existing continental crust (Martin *et al.*, 1997). In the Brumado region to the W of the Sete Voltas Massif the TTG suites were dated at 3.3 Ga. According to the Nd isotope evidence these rocks originated from different protoliths (3.36 - 3.26 Ga; 3.42 Ga). Therefore, distinct accretionary processes participated in the evolution of the Gavião Block (Bastos Leal *et al.*, 1996, 1997).

In the Aracatu region (Fig. 2), the Gavião Block has been divided into three domains on the basis of $^{207}Pb/^{206}Pb$ dating on single zircon and monazite crystals, supported by Sr and Nd isotopic data and geochemical modelling (Santos Pinto *et al.*, 1998). Firstly, an Archean juvenile-type terrane represented by the Bernarda Massif grey gneiss, the mantle extraction of which took place around 3.3 Ga ago. Secondly, an Archean domain dated between 3.24 and 3.16 Ga, characterized either by recycling or by juvenile accretion processes (Fig. 3). This domain consists of trondhjemitic grey gneiss (Aracatu Massif), and K-rich calc-alkaline granitoid plutons such as the Mariana and Serra do Eixo massifs. The third domain is distinguished by the Paleoproterozoic tectonic overprint, exemplified by the case of the Umburanas Pluton. The U/Pb ages between 3.1 - 2.8 Ga (inherited zircon) indicate the Archean protholith, whereas a monazite age of *c.* 2.0 Ga is supposed to date the reworking of the pluton during the Transamazonian Orogeny.

Despite the few direct radiometric dates available on the 3.4 Ga TTG rocks their distribution from the Jacobina region (Mougeot, 1996) to the southeastern and southwestern parts of the Gavião Block suggests their large paleogeographical extent. Besides the Nd and Sr isotope evidence (Martin *et al.*, 1997), a $^{207}Pb/^{206}Pb$ age of 3.473 ± 0.008 Ga, in single zircon core of the Sete Voltas Massif has been reported (Nutman and Cordani, 1993). Therefore a significant primitive continental segment corresponding in size at least to the whole present-day Gavião Block can be postulated. Moreover, recent Nd isotopic mapping (Sato, 1998) points to the existence of a continental crust older than 3.4 Ga in northern part of the SFC.

At least two migmatitic events took place in the Gavião Block (Santos Pinto *et al.*, 1998; Bastos Leal *et al.*, 1996, 1997). The first affected the old grey gneiss at *c.* 3.2 Ga before the emplacement of the Mariana and Serra do Eixo (type 1) massifs, whereas the second affected the younger grey gneiss at 2.91 Ga, preceding emplacement of the Neoarchean grey granite. Finally, the Transamazonian Orogeny was also responsible for anatectic partial melting processes dated at 2.1 Ga.

The younger grey gneiss derived from their precursor magmas by melting of pre-existing crust, as supported from petrogenetic modeling, is similar as to the old grey gneiss (Martin *et al.*, 1997). Consequently, the calculated mineral assemblage (quartz+hornblende+plagioclase) in equilibrium with the magmatic liquid implies P-T conditions achieved in the interval 10 to 15 kbar and 800 to 700 °C, for H_2O <5% to 15% respectively (Johnson and Wyllie, 1988). This suggests that the hydrous melting of the old grey gneiss occurred at depth of *c.* 30 - 45 km (Martin *et al.*, 1997).

Supracrustal units and Mesoarchean intrusions

Greenstone belts, granitoid intrusions and medium-grade gneiss-leptynite-amphibolite supracrustal associations are common within the Gavião Block. The latter give ages of 2.7 - 2.6 Ga (Marinho *et al.*, 1979; Brito Neves *et al.*, 1980; Cordani *et al.*, 1985). Granitoid bodies, either high-K calc-alkaline in composition such as the Lagoa do Macambira and Malhada de Pedras plutons to the W (Bastos Leal *et al.*, 1996) or peraluminous plutons such as Serra dos Pombos Massif to the E (Marinho 1991; Marinho *et al.*, 1994a, b) emplaced at 2.9 - 2.8 Ga. The Archean volcano-sedimentary sequences such as the lower unit of the Contendas-Mirante Belt (Marinho, 1991), the Umburanas and Ibitiara-Brumado rock assemblages (Cunha and Fróes, 1994), as well as the Bate-Pé unit are now recognized to be greenstone belts. They associate komatiite, basalt, felsic calc-alkaline volcanic units, as well as some chemical and detrital sediment. SHRIMP U/Pb and $^{207}Pb/^{206}Pb$ isotope data on zircon indicate ages of 3.3 and 3.0 Ga for the mafic units in these belts, whereas the felsic units yield ages of 2.75 and 2.50 Ga (Umburanas and Contendas-Mirante belts, respectively; Marinho 1991; Cunha *et al.*, 1996; Bastos Leal *et al.*, 1997). These units mark the occurrence of mantle derived volcanism following individualization of the TTG continental segments within the greenstone belts, as well as a Mesoarchean tectono-magmatic episode that generated felsic volcanics and/or plutons. In particular, the basal mafic sequence of the Contendas Mirante possibly represents ocean floor remnants trapped during a 3.0 Ga continental growth. However, the tectonic scenario of this basin also includes Paleoproterozoic assemblages (Marinho *et al.*, 1979).

Tectonic slices of the grey gneiss of the Gavião Block (Sete Voltas, Boa Vista-Mata Verde) are imbricated into the Contendas-Mirante Belt. In spite of the lack of typical komatiite within this belt, the rock association of the Archean basal unit resembles other greenstone associations that crop out in the Gavião Block (Cunha and Fróes, 1994). The geological inferences support the view that huge Early Archean (3.3 - 3.0 Ga) tectonic slices related to the Gavião Block played an important role in the formation of the Contendas-Mirante Basin. Chemically, its lower mafic and intermediate rocks have a contaminated tholeiitic signature, and the trends are in good agreement with modern island arc series, and the high-Ti values differ appreciably from the tholeiitic rocks of the Paleoproterozoic Itabuna-Atlantic Coast Belt (Marinho *et al.*, 1994b). The whole geochemical patterns and especially the behaviour of Nb, P and Ti are compatible with continental tholeiites (Marinho, 1991), but the existence of at least some oceanic tholeiite remnants cannot be ruled out (Sabaté and Marinho, 1982). In the same way, the differentiated acid subvolcanic rocks have intermediate geochemical characteristics.

Attempts to date the tholeiitic metavolcanics of the Contendas Mirante greenstone belt have not been successful and the Sm/Nd (T_{DM}) model ages are most powerful tool. Marinho (1991) defined two age groups, corresponding to the recognized geochemical evolutionary trends: between 3.0 - 2.3 Ga and at 3.3 Ga. With the support of geochemical and isotopic studies, the maximum age for the extrusion of the tholeiitic magma from the mantle occurred at 3.0 Ga (Marinho *et al.*, 1994b). This interpretation is not resolved by the dates on the banded iron formation units intercalated with the tholeiitic lava, that give 3.3 Ga (Pb/Pb), and have T_{DM} ages of *c.* 3.0 Ga. Nevertheless, these model ages conflict with the zircon U/Pb age of the subvolcanic acid rocks (see below) that intrude the above lithotypes (Marinho *et al.*, 1994b). On the other hand, the positive $\varepsilon_{Nd(3.0\,Ga)}$ values are not compatible with an upper crust source, as previously stated.

Whole-rock errorchrons on the acid subvolcanic rocks of the Contendas Mirante Belt yield *c.* 3.0 Ga (Pb/Pb) and *c.* 2.1 Ga (Rb/Sr). These were interpreted to be the ages of the emplacement and of the Transamazonian metamorphic overprint, respectively (Wilson, 1987). However, further Sm/Nd T_{DM} ages of 3.3 and 3.4 Ga ($\varepsilon_{Nd(3.0\,Ga)}$ values between - 0.4 and - 0.9) coupled with an upper concordia intercept U/Pb age of 3.3 Ga are now considered to be the best estimated intrusion age of these subvolcanics (Marinho, 1991).

Situated to the W of the Contendas Mirante Basin, the Umburanas greenstone belt was investigated by SHRIMP U/Pb zircon geochronology. Its basal conglomeratic quartzite beds include two zircon populations with ages clustering at 3.25 Ga and between 3.33 - 3.04 Ga, respectively (Bastos Leal *et al.*, 1996, 1997). The large age spectra is compatible with the crustal evolution of the Gavião Block (see above), implying that the Umburanas basal unit was derived by erosion of distinct pre-existent continental rocks. In addition, $^{207}Pb/^{206}Pb$ zircon evaporation ages on the Umburanas meta-andesite middle unit yielded 2.75 Ga, constraining the basin evolution between 3.04 and 2.75 Ga. In addition, the existence of a 2.6 - 2.5 Ga granitic intrusive into this greenstone belt suggests a probable genetic relationship for the meta-volcanics. These rocks have $\varepsilon_{Nd\,(2.75\,Ga)}$ values from + 1.5 to - 4.5, which compare well with that of the lower metakomatiite unit. Therefore, the evolution of the Umburanas Basin probably combined crustal contamination and juvenile accretion processes, in coherence with either an intraplate or proximal continental setting for the basin.

Structure and tectonics

The present-day structural frame of Archean terranes of the Gavião Block is strongly controlled by the Proterozoic tectonic events, and the Transamazonian deformations in particular, developed into deep ductile conditions, largely penetrative from regional to sample scale. Such a scenario is mirrored by the Sete Voltas, Mairí and Mundo Novo slices that are elongated and imbricated along with Paleoproterozoic supracrustal rocks into the Contendas-Jacobina Lineament, as well as by the mosaic drawn by the outcrop of lithotectonic units, their internal thrust faults and shear-zones. On the other hand, during the Brasiliano Orogeny tectonic discontinuities under a ductile regime were formed within the Gavião Block, such as the westward thrust fault of Caetité (that marks the western limit of this block), as well as the two related eastward backthrusts of the Paramirim band. The Transamazonian shear zones were also reworked by the Brasiliano episodes, as illustrated by the NW-SE shearing on both sides of the Umburanas greenstone belt (Cunha *et al.*, 1994).

Archean structures can also be distinguished in such a complex structural framework. This is seen by the 3.4 Ga grey gneiss (Sete Voltas Massif) that occur as a great enclave into the younger grey gneiss, the former preserving a distinct foliation from that of the host rock. In the same way, the 3.1 Ga porphyritic granodiorite exhibits planar and linear magmatic preferred orientation, marked by undeformed zoned plagioclase phenocrysts and biotite crystals, which are otherwise only slightly folded by the Transamazonian shortening. These flat preferred orientation suggests that horizontal kinematics operated during early emplacement of the plutonic body. In addition, tectonic slices of the Gavião Block (see above) exhibit marginal zones with a similar magmatic texture, but this primary orientation is foliated by a strong sub-vertical superposed ductile deformation, related to the uplift and tectonic transport of the slices onto the Contendas-Mirante Terrane. In the Aracatu region, the 3.2 Ga grey gneiss exhibit such a flat foliation with sheath folds assigned to the Archean (Sabaté *et al.*, 1988). Considering the thickness of the 3.2 Ga crust estimated above (Martin *et al.*, 1997), as well as its intense migmatization and strong (Archean) foliation, the old grey gneiss are interpreted as the product of collisional thickening. For these authors, this interpretation, together with the existence of horizontal tectonics, favours the idea that modern-style plate tectonics operated in the Gavião Block, during the Early Archean.

Finally, the geometry of the southern part of the Umburanas greenstone belt seems to be constrained by the transpressive dextral emplacement of the Serra do Eixo Granitoid (Sabaté *et al.*, 1988) originated at *c.* 3.2 Ga (Santos Pinto *et al.*, 1998). However, further tectonic interference was introduced into the basin by a calc-alkaline plutonism dated at 2.6 - 2.5 Ga (Bastos Leal *et al.*, 1997). The greenstone belt dynamics may have resulted from at least a diapir model regime, although kinematic evidence of vertical downsagging/rising have not been observed yet.

The Jequié Block

The Jequié Block crops out to the SE of the Contendas-Jacobina Lineament, in tectonic contact with the Contendas-Mirante Belt to the W as well as with the Salvador-Curaçá and the Itabuna-Atlantic Coast Belt to the N and E, respectively. The Jequié Block consists of charnockite-enderbites (Barbosa, 1990) which intrude ortho-derived rocks, supracrustals and anatectic granite bodies. All these rocks are metamorphosed in granulite facies. The Ipiau band (Barbosa 1986) which associates gabbro and supracrustals in the amphibolite facies transition to granulite facies occupies the interface between the Jequié Block and the Itabuna-Atlantic Coast Belt.

Lithotypes and Petrogenesis

Charnockite-enderbite

Generally highly deformed, the enderbite-charnockite association is extensively distributed in the Jequié Block but predominates in its eastern part. These rocks comprise a plutonic suite varying from enderbitic to charnockitic terms either hornblende-free or hornblende-bearing enderbite-charnockite, metamorphosed in the granulite facies but which underwent varying degrees of retrometamorphism (Barbosa *et al.*, 1998). Intrusive charnockite bodies (Brejões Dome) are observed.

Geochemical signatures (Fornari, 1992) show that these rocks belong to a calc-alkaline magmatic series. However, it is possible that they may contain a contribution from a crustal source, as for example, in the case of the Brejões Dome. Pb/Pb zircon evaporation ages on this body give 2.7 Ga and 2.1 Ga, tentatively interpreted as the emplacement/crystallization time of the charnockite and as the granulite metamorphic overprint, respectively (Barbosa *et al.*, 1998). At the present neither recycling of an older TTG crust has been identified geochemically in the plutonic rocks of the Jequié Block, nor a typical TTG suites have been observed.

A SHRIMP U/Pb age of 2.69 Ga (Alibert and Barbosa, 1992), similar to a previous Rb/Sr isochron age (Wilson, 1987), indicates the crystallization age for the hornblende-bearing enderbite-charnockite, whereas a T_{DM} age of 3.0 Ga ($\varepsilon_{Nd(3.0\,Ga)} = +1.8$) constrains the magmatic extraction time and the mantle source isotopic signature. The hornblende-free enderbite-charnockite yields a distinct U/Pb SHRIMP age of 2.81 Ga (Alibert and Barbosa, 1992).

In the Maracás region, in western part of the Jequié Block, the charnockitic rocks yield a Pb/Pb whole-rock isochron age of 2.66 Ga, and the T_{DM} model ages are between 3.4 - 3.2 Ga (Marinho 1991; Marinho *et al.*, 1994b). These results suggest a diachronic history of plutonic accretion in the Jequié Block and for source heterogeneities, with probable recycling of a previous continental crust.

Heterogeneous granulite

Heterogeneous dark green granulite with charnockitic characteristics constitutes a distinct lithological unit of the Jequié Block. They contain fine to medium-grain basic enclaves and/or disrupted dykes as well as quartz-feldspathic bands and supracrustal xenoliths. Wilson (1987) reported Pb/Pb a whole-rock isochron age of 2.9 Ga (Pb/Pb whole-rock isochron), whereas the T_{DM} ages are in the range 2.9 - 2.6 Ga. In the western Jequiriçá region, the same author determined an age of 2.7 Ga (Rb/Sr whole-rock isochron) and a T_{DM} age of *c.* 3.2 Ga. These unique rocks appear to be a result of multiple and non co-magmatic protoliths, which represent the oldest source materials of the Jequié Block accretion.

Supracrustal and related rocks

The Jequié supracrustal rocks consist of intercalated bands of aluminous kinzigitic gneiss, basic to intermediate metavolcanics, iron formation units, quartzite, quartz-feldspathic bands and some graphitic zones. They represent rocks essentially derived from some volcanic and/or volcaniclastic deposit intercalated with chemical sediments, all lithotypes being recrystallized in the granulite facies. Mineralogical and geochemical investigations support a pelitic origin for the aluminous sediments and a tholeiitic affinity for the metavolcanics (Barbosa and Fonteilles, 1989). Undeformed garnet-cordierite bearing charnockite dated at 2.1 Ga is associated with the aluminous kinzigite-gneiss, and are interpreted to be S-type intrusive granites (Barbosa, 1990). The age of this charnockite is similar to that of the granulite metamorphism as shown by the Brejões Dome (see above). It also supports the assessment that the aluminous granitic magma originated during the peak of the high-grade metamorphism (Barbosa *et al.*, 1998).

Additional radiometric data from the Jequié rocks are in the range 2.1 - 2.0 Ga, supporting the relationship between isotope resetting and the granulitic metamorphism (Ledru *et al.*, 1994). According to Barbosa (1990) the granulite event took place, under thermodynamic conditions of intermediate pressure (5 - 7 kbar, 850 - 870 °C). However, on the western border of the Jequié Block, at the contact with the Contendas-Mirante Belt, the narrow band of the Porto Alegre charnockite (Fig. 2) underwent retrograde metamorphism from orthopyroxene to green-hornblende equilibrium, due to dehydration of the Contendas-Mirante volcano-sedimentary rocks under the rocks of the overthrusted Jequié Block.

Structure and tectonics

The structural framework of the Jequié Block, in like manner to the Gavião Block, is controlled by the Transamazonian episodes (Ledru *et al.*, 1994). During this event strong penetrative granulitic foliation and/or banding affected the country rocks. Available data suggests that the block was affected by at least two episodes of ductile deformation (Barbosa, 1986; Barbosa *et al.*, 1994). The first episode created recumbent folds with an approximately N-S horizontal axis related to shear ramps with vergence to the W. The first foliation is refolded with tight isoclinal folds, also with subhorizontal axis but with subvertical axial planes. The latter may produce a new axial plane foliation that locally transposes the previous foliation. Interference patterns of these two deformational episodes may occur, at least at a cartographic scale (Barbosa, 1986). Although the regional mapping and radar remote sensing indicate a good distribution of the dominant foliation trajectories, a systematic study of transport lineation trajectories and kinematic criteria does not exist. Based on the available data, a model of megablocks displaced at depth conforming to a system of frontal and lateral tectonic ramps is assumed (Gomes *et al.*, 1991).

Although the Transamazonian episodes clearly influenced the architecture of the Jequié Block, much older structures have also been suggested (Barbosa, 1986; Marinho *et al.*, 1994a; Ledru *et al.*, 1994), but the existence of 2.7 Ga granulitic metamorphism is still a matter of debate. Indeed, no thermo-barometric data on granulitic mineral assemblage point to previous metamorphism under other (P, T) conditions than the Transamazonian high-grade regime, unless the latter had obliterated it. Nevertheless, recumbent folding, described previously, suggests a pre-existing granulitic foliation. In addition to which, well-preserved sub-horizontal sheath folds associated with an approximately N-S mineral stretching lineation and kinematic patterns indicating a vergence to the N, were recently observed in the Ipiau band, in the eastern part of the Jequié Block (Sabaté, unpublished report; J. C. Cunha, personal communication), clear evidence of horizontal Archean dynamics. These preliminary observations prevent the hasty elimination of the existence of an Archean granulitic event in the Jequié Block.

On the contrary, in the main central and western parts of the Jequié Block, recent mesoscopic observations (P. Sabaté) suggest that penetrative sub-vertical foliation result from pure shear deformation. No apparent stretching lineation is developed. Observed polygonal microscopic textures reinforce this view. In this case, the shortening model held by Choukroune *et al.* (1995) turns to an alternative for the Neoarchean tectonic regime in the Jequié Block. The dome-like structure of the Brejões charnockite was considered to represent a typical dome-basin interference pattern (Barbosa 1986, 1990). The mapping and satellite imagery of the Brejões-type bodies may be interpreted in terms of a diapiric regime that conditioned the plutonic emplacement into the supracrustal rocks, at 2.7 Ga. However, the regionally penetrative Transamazonian structures seem on radar imagery and in the field do not introduce in the shape of the body and in its mesoscopic patterns the expected deformation that it should overprint. From these observations we suggest that the intrusion of the Brejões and neighbouring bodies may be contemporaneous with the Transamazonian tectonics. Therefore the *c.* 2.1 Ga age probably refers to the emplacement of the Brejões Dome, whereas the 2.7 Ga age probably reflects an inherited signature from a solid deep crustal source.

In summary, the studies on the Jequié Block and related units do not lead, for the moment, to consistent conclusions about Neoarchean dynamics, and both horizontal and vertical tectonics may have occurred. Except for the Transamazonian tectonic tensors that produced analogous structural features in both the Gavião and the Jequié blocks (see previous items), no evidence for a common Neoarchean tectonic pattern has been clearly established.

Serrinha Block

This is an elongated N-S segment up to 100 km wide, limited to the W and to the S by a tectonic contact with the Salvador-Curaçá Belt. The Serrinha Block does not have any visible connection, either with the Jequié Block or with any litho-structural unit of the Contendas Jacobina Lineament. It is composed of medium-grade gneiss-migmatitic rocks that constitute the basement of the Rio Itapicuru granite-greenstone belt, as well as its equivalent to the NW, the Rio Capim greenstone belt (Schrank and Silva, 1993). Syn to late-tectonic granitoid plutons were emplaced in the greenstone belt and mark the steps of the Transamazonian evolution within this block.

The few markers of the Archean evolution are the migmatitic xenoliths emplaced in the Rio Itapicuru greenstone belt, which yield a U/Pb zircon age of 2.93 Ga (Gaál *et al.*, 1987). These xenoliths are probable crustal remnants of granite sources of the Ambrósio Dome and related plutons.

New SHRIMP U/Pb data on zircon from a granodiorite intrusive in the Ambrósio Dome yielded two age groups (3.11 and 2.94 Ga), and two younger ages clustering around 2.16 and 2.08 Ga. The surrounding migmatitic gneiss were also dated (3.15 and 3.09 Ga), in agreement with the Archean ages of the granodiorite dome (Mello *et al.*, 1999b). These results suggest an autochtonous palyngenetic evolution of the Ambrosio Dome, from its early Archean gneiss (at 3.11 Ga) to granodiorite resulting from possible successive partial melting of the former at 2.93 and 2.16 - 2.08 Ga, and inheriting the zircon markers from its sources. Also the tonalitic gneiss near Uauá (northern part of the Serrinha Block) have SHRIMP U/Pb zircon ages between 3.13 and 3.05 Ga (Cordani *et al.*, 1999). This tonalitic gneiss is therefore clearly older than the Rio Capim greenstone sequence. They may also constitute one of the detrital sediment sources of the Caldeirão supracrustal rocks that form a narrow belt at the northwestern margin of the Serrinha Block (Mello *et al.*, 1999a, b).

Archean crustal evolution

The northern part of the SFC preserves two Archean blocks (Gavião and Jequié) which record a diachronic evolution and constitute two true continental fragments. Besides the strong age contrast between the blocks, the sources and thermodynamic conditions for the genesis of the respective components, as well as their Archean histories, remain clearly distinct. However, recycling mechanisms of the older crustal components do not permit the establishment a parental link between the primitive rocks of these two blocks.

The Gavião Block was assembled essentially by accretion of Early Archean TTG and typical greenstone belts between 3.4 and 3.0 Ga, in three ways: juvenile TTG plutonism, juvenile TTG plutonism with crustal contamination, and recycled intermediate TTG/calc-alkaline magmatism. Continental accretion occurred periodically as can be determined by the isotope ages which seem to be about 100 Ma. Major and trace element petrogenetic modelling suggest that pre-existing rocks similar to the old grey gneiss may be the solid source, the melting of which producing the precursor magma from which derived the younger trondhjemitic grey gneiss and porphyritic granodiorite. This is supported by the intense migmatization that affected the old grey gneiss before the emplacement of the 3.1 Ga intrusives.

The Jequié Block may be considered as a continental segment built between 2.9 and 2.6 Ga by accretion of plutonic rocks derived from the calc-alkaline magmatic series (charnockite-enderbite), with or without a contribution from an older crustal source (supracrustal granulite), tholeiitic volcanics and chemical sediments. Barbosa *et al.*, (1994) suggested geodynamic mechanisms for oceanic crust subduction to the W underneath the Gavião Block to generate such calc-alkaline plutonism. Although the possible source for some charnockite and charnockitic orthogneiss may be the 3.2-3.4 Ga crust, to date, neither typical TTG suites have been found, nor has any geochemical evidence for recycling of an older TTG crust been observed in the plutons of the Jequié Block. This leads to the suggestion that the two blocks may have had an independent evolution and may represent two separated protocontinents during the Late Archean. This interpretation is also in agreement with the contrasting Nd isotope signatures between the Jequié and Gavião blocks (Bastos Leal *et al.*, 1996, 1997).

The two crustal segments were juxtaposed due to the Transamazonian collision. In this evolutionary context, the Contendas-Jacobina Lineament constitutes a major suture separating the two continents. Both were affected by shortening and strike-slip dynamics, which obliterated most of the Archean tectonic markers. The early tectonic history of the Gavião Block is better preserved and shows tangential horizontal mechanisms. These, combined with crustal thickness, constrained by the grey gneiss melting parameters, argue for modern style plate tectonics in the early Archean, as already seen. Although such model is not fully accurate for the Jequié Block, it agrees with the on-going studies.

Finally, in the light of available data, the Serrinha Block represents an older continental segment extensively reworked during its Paleoproterozoic evolution. It may be speculatively considered either as a third Archean continental block, containing gneiss-migmatitic rocks similar to those of the Gavião Block, or as an upper level non-granulitic lateral equivalent of the Jequié Block, separated from it by the Transamazonian Salvador-Curaçá plutonic belt.

The Paleoproterozoic crust

As already discussed, most of the geometric structure in the northern part of the SFC resulted from the Paleoproterozoic history. In addition, new juvenile or recycled magmatic events as well as new deposits participated in the continental growth and collisional processes; most of these lithostructural units being contemporaneous with the Transamazonian Orogeny. A few of them emplaced earlier, at the beginning of the Paleoproterozoic, marking unique Pre-Transamazonian events in the South American Continent. We will show that this early evolution is also diachronic and that these units may represent a response to the sinking (or subducting) processes proposed to explain the Transamazonian tectonic convergence toward the Contendas-Jacobina axis.

Pre-Transamazonian units

Rio Jacaré Sill

The Rio Jacaré Sill comprises a stratified mafic-ultramafic body, composed of gabbro, which forms a lower zone as well as a stratified upper zone where alternating gabbro, pyroxenite and magnetitite occur (Brito, 1984). Magnetite-rich layers form a vanadium deposit. Plutonic textural relicts may be locally preserved at microscopic scale, but widespread recrystallization related to regional deformation processes developed in the amphibolite facies. Detailed mapping indicated that this plutonic complex forms a tectonic slice imbricated into the Paleoproterozoic clastic metasediments of the Contendas Mirante Belt.

A whole-rock Pb/Pb errorchron (Marinho, 1991) gives an age of 2.47 Ga, and the Sm/Nd T_{DM} model ages are 3.3 and 3.5 Ga. The calculated $\varepsilon_{Nd\ (2.5\ Ga)}$ parameters, all in the negative range (- 3.3 to - 6.6), indicate crustal contamination, which is confirmed by the geochemical trends (Marinho, 1991). Based on isotope and geochemical studies this author argues that the *c.* 2.5 age represents the time of magmatic crystallization of the Rio Jacaré body.

Calc-alkaline metavolcanics

These metavolcanic rocks crop out intercalated like *melanges* into the metapelite beds of the Contendas-Mirante Belt, as well as occur in the form of a continuous layer (Brito, 1984), which lies along the border of the lower discontinuity of the Rio Jacaré tectonic slice. Therefore, the volcanics constitute the tectonic interface between the Rio Jacaré mafic-ultramafic complex and the host metasediments of the Contendas-Mirante Belt. The volcanic rocks show calk-alkaline afinities and are composed of foliated metabasalt and andesite, recrystallized in the amphibolite facies.

A Pb/Pb whole-rock isochron age for these rocks is in the range 2.52 - 2.49 Ga, in agreement within error with the 2.47 Ga age of the Rio Jacaré Sill, probably indicating the extrusion date of these metavolcanic rocks. Additional T_{DM} ages are between 3.45 and 3.25 Ga, and the calculated $\varepsilon_{Nd(2.5\,Ga)}$ values vary from - 4.5 to - 6.5. One sample gave a 3.0 Ga T_{DM} model age with a positive $\varepsilon_{Nd(2.5\,Ga)}$ of + 1.8 (Marinho, 1991). These data suggest an upper continental crust source for the metavolcanic rocks (Marinho, 1991; Marinho *et al.*, 1994a), which included contamination from different sources of the Gavião Block.

From the above these calc-alkaline rocks may represent a volcanic component of a 2.5 Ga extrusive magmatism proximal to the margin of the Gavião Block, nearby, and contemporaneous with the deeper emplacement Rio Jacaré mafic-ultramafic body. All together the magmatism marks an extensional tectonic episode at the Archean-Paleoproterozoic transition, in the northern part of the SFC.

Pé de Serra Pluton

The Pé de Serra Massif represents a stretched and N-S elongated belt (*c.* 100 x 5 km), situated between the Jequié Block and the northeastern part of the Contendas-Mirante Belt. It is composed of sub-alkaline granite with granoblastic texture, as well as alkaline granitic and syenitic intrusions. The former contains Mg-hastingsite hornblende and titanite, whereas the latter contains aegirine and andradite. The sub-alkaline granite bodies appear strongly deformed and recrystallized by a E-W shortening responsible for foliation and/or local banding and also for tight centimetric to decimetric upright similar folds. The alkaline granite intrusives are clearly less deformed and may have been emplaced after the sub-alkaline ones.

Geochemically, the plutonic rocks show a clear metaluminous character, high K_2O content and high REE fractionation with moderately negative Eu anomaly. Their REE patterns are comparable to those of the charnockitic rocks of the Maracás region (Marinho, 1991). The sub-alkaline granite intrusives give ages of 2.56 - 2.55 Ga (whole-rock Pb/Pb errorchron and Rb/Sr isochron), interpreted to be the emplacement time of the plutonism. The T_{DM} ages are between 3.2 and 3.1 Ga and the $\varepsilon_{Nd(2.5\,Ga)}$ in the negative range - 3.9 to - 4.4. According to Marinho (1991) the latter values are consistent with the other geochemical parameters and attest to crustal contribution from sources that might correspond with the Jequié charnockite-enderbite. For the alkaline plutons, a Pb/Pb whole-rock isochron yield *c.* 2.3 Ga, considered to be the time of crystallization of these rocks, which were probably derived from analogous charnockite-enderbite sources.

This unique sub-alkaline and alkaline plutonism, coupled with its geological location and geometry argue for an intra-continental rift setting at the border of the Jequié Block. This rift started by emplacement of the *c.* 2.56 - 2.55 Ga sub-alkaline granite, which took place just before the extension regime was aborted to give place to the compressive/transpressive conditions of the Transamazonian convergence and collision. Further extension may have occurred at *c.* 2.3 Ga to allow for the intrusion of alkaline magma, which may have been conditioned to small extensional pull-apart structures along the previous rift faults, during the succeeding transpressive event.

Transamazonian magmatic belts

Salvador-Curaçá Belt

The Salvador-Curaçá Belt (Santos and Souza, 1983) constitutes an elongated accretion prism, which was involved in the Transamazonian collision, together with the Jacobina-Mundo Novo Belt and the Mairí tectonic slice (Fig. 3). The Salvador-Curaçá Belt comprises three main lithological units (Caraíba, São José do Jacuipe, and Ipirá complexes), in addition to several plutonic pulses associated with the Transamazonian Orogeny.

The Caraíba Complex (Figueiredo, 1981), situated in the northern branch of the belt consists of granulite facies orthogneiss, but exhibits clear plutonic characteristics (Teixeira and Melo 1990; Melo, 1991). Geochemical studies (Teixeira, 1997) revealed sodic orthogneiss, analogous in composition to the trondhjemite suites, and soda-potassic orthogneiss of calc-alkaline affinity. The former occur in continuous bands at the eastern and western borders of the belt, as well as in its north-central part where they form two parallel narrow and discontinuous belts. The latter, and more significant, occupy the central and eastern parts of the belt. The distribution of the plutonic terranes suggests a rough symmetry from the axial part of the belt. The trondhjemitic orthogneiss reveal a juvenile origin through a two stage petrogenetic modelling (Martin, 1994; Teixeira, 1997).

Pb/Pb determinations on magmatic idiomorphic single zircon crystals as well as on metamorphic zircon overgrowths from the orthogneiss of the Caraíba Complex gave similar ages of 2.1 Ga (Sabaté *et al.*, 1994). This metamorphic age supports the view that the reported older Rb/Sr whole-rock isochron ages of 2.35 and 2.30 Ga (Brito Neves *et al.*, 1980; Pereira 1992) may correspond to the intrusion age of the Caraíba rocks in thermodynamic conditions with the host rocks, under granulite facies conditions. Additional T_{DM} model ages in the range 3.4 - 2.5 Ga (Sato, 1998) indicate that different sources participated in the evolution of the Caraíba Complex.

The São José do Jacuipe Suite forms discontinuous bands and lenses tectonically intercalated into the Complex close to its western border. It is composed of mafic and ultramafic rocks derived from tholeiitic magma, and shows slight crustal contamination. This suite is considered to be an old oceanic crust remnant, similar to modern oceanic floor found at passive margins (Teixeira, 1997), but isotope data are needed to confirm this interpretation.

The Ipirá Complex (Melo, 1991), probably a Paleoproterozoic platform cover, groups the supracustal rocks of the Salvador-Curaçá Belt. It is mainly composed of kinzigitic and garnet-bearing gneiss, calc-silicate rock, chert and banded iron formation units, as well as subordinate bands of basic rocks. At the western border of this belt, near its thrust contact over the lithotectonic imbrication of the Contendas-Jacobina Lineament, intercalated aluminous gneiss have the mineral assemblage orthopyroxene+ garnet+sapphirine (Leite, unpublished manuscript) indicating the higher P-T conditions.

At least five successive intrusive magmatic episodes were identified in the Salvador-Curaça Belt, according to the structural phases and radiometric data (Padilha and Melo, 1991). The geochemical trends of the granitoid plutons vary from the earlier, monzonitic (Teixeira, 1991, 1997), to the later, shoshonitic variety. One of the earliest plutonic episodes has a Rb/Sr isochron age of 2.01 Ga. The last, post-kinematic phase, gives a Rb/Sr isochron age of *c.* 1.9 Ga (Melo, 1991). The Itiúba Massif to the N of the belt corresponds to a syenitic batholith (180x15 km). Its smaller equivalents to the S are the Santanápolis and São Felix bodies. These intrusions form a N-S elongated "syenitic line" up to 800 km long near the eastern border of the belt, related to a lithosphere-scale shear discontinuity (Conceição *et al.*, 1989). They correspond to an alkaline to high-K saturated magmatism and are composed of a monotonous mantle-derived cumulate syenite or more diversified rock associations (Conceição *et al.*, 1999). In spite of shearing contemporary to the emplacement, the Itiúba Massif preserved magmatic structures in its core, whereas its borders display viscous to ductile and the ductile conditions.

The emplacement of these syenite intrusives took place between 2.14 Ga and 2.06 Ga, as supported by the Rb/Sr data from the Itiúba and Santanápolois bodies, respectively (Conceição, 1998), in agreement with a U/Pb zircon age (2.1 Ga) for the latter (Conceição *et al.*, 1999). A T_{DM} model age for the Itiuba syenite yielded 2.6 Ga (Sato, 1998).

The structural frame is managed by a transpressive regime, which existed from the beginning of pluton emplacement to the youngest intrusion of the Salvador-Curaçá Belt. This regime is responsible for a strong near E-W shortening and N-S stretching compensated by continuous sinistral shear bands contemporary to successive plutonic emplacement. The shearing behaved from viscous magmatic to rigid conditions through sub-solidus and ductile conditions and accompanied the progressive thickening of the belt. Later intrusions were emplaced at the transition between granulite and amphibolite facies. They deform into viscous conditions and slightly, or locally, into ductile conditions. The latest intrusion presents only magmatic deformations and was emplaced due to late ductile strike-slip faults. Besides, the westward motion of both the Salvador-Curaçá Belt and the Archean Serrinha Block, we suggest that the latter probably provided higher kinematic energy (or a higher mass inertia gradient) to explain the resulting overthrusting of the belt onto both the Serrinha Block to the E and a western domain, comprising the Jacobina-Mundo Novo Belt and the Mairí tectonic slice. The two-fold vergence of the resulting frame was evidence for a positive flower tectonic feature (Padilha and Melo, 1991) and interpreted as an oblique collision between the Mairí slice and the Serrinha Block.

Guanambi-Urandi Batholith

The huge Guanambi-Urandi Batholith (Rosa *et al.*, 1996), previously named the Correntina-Guanambi Block (Brito Neves *et al.*, 1980) forms a N-S inlier situated to the W of the Gavião Block, partly covered by the Neoproterozoic Bambuí Group. Remnants of the intruded country rock constitute a small narrow band to the E of the batholith, and correspond to the Santa Isabel Complex for which the Rb/Sr whole-rock isochron ages are between 2.7 and 2.6 Ga (Mascarenhas and Garcia, 1989). This Complex consists of gneiss and migmatite metamorphosed in the granulite to the high-amphibolite facies, appearing as a tectonic slice under the Espinhaço Belt.

The Guanambi-Urandi Batholith consists of multiple intrusions (Rosa *et al.*, 1996). These plutons are composed of monzonite, syenite and granite, having mineralogy, textures and a high-K geochemical signature that are comparable to those of the Itiuba, Santanápolis and São Felix bodies. However, they are less deformed than the eastern equivalents of the syenitic line, in spite of their proximity of both to the Espinhaço Belt and the Santa Isabel tectonic slice. These host rocks seem to have absorbed the Brasiliano tectonics, which otherwise suggests that the metamorphism of the Santa Isabel Complex could be related to the Brasiliano event as well.

Pb/Pb data on single zircon crystals give ages of 2.06 to 2.0 Ga, which correspond to the crystallization age of the batholith, in agreement within error with the previous Rb/Sr isochron ages (Bastos Leal *et al.*, 1996; Leahy *et al.*, 1998). Two T_{DM} ages on granitoid plutons near Guanambi yielded 3.1 and 2.42 Ga (Sato, 1998) and suggest an isotope signature similar to that of Salvador-Curaçá Belt rocks. Geochemical and isotopic arguments indicate an enriched mantle source for these high-K plutons (Paim, 1998), but additional Sr and Nd evidence (Barreto dos Santos *et al.*, 1999; Leahy *et al.*, 1999; Rosa *et al.*, 1999) suggests the participation of crustal components in the magma genesis.

Itabuna-Atlantic Coast Belt

The rocks in the Itabuna-Atlantic coastal belt were derived either from volcanic or plutonic protoliths. Completely recrystallized in granulite facies conditions. In like manner to the neighbouring rocks of the Jequié Block and the Salvador-Curaça Belt, these rocks have lost all their previous mineralogical assemblages and textures. Geochemically, they show, from E to W successively, arc-tholeiite, low-K calc-alkaline and shoshonitic affinities (Barbosa, 1986). With the chemical characteristics and the interpreted tectonic setting, the Itabuna-Atlantic Coast Belt resembles either a modern volcanic arc or an active continental margin magmatic association like that of Japan or Chile, respectively (Figueiredo, 1989; Barbosa, 1990). On the other hand, considering that the magmatic belt is matched with the Jequié Block, it underwent at least the same three deformation episodes under thermodynamically analogous conditions (Barbosa, 1990), *i.e.*, intermediate pressure 5 - 7 kbar and high temperatures of 850 - 870 °C.

The Rb/Sr whole-rock isochrons yield ages from 2.3 Ga to 2.0 Ga (Mascarenhas and Garcia, 1989) that were tentatively interpreted in terms of Transamazonian reworking of Archean crust. A U/Pb zircon age of 2.45 Ga in felsic gneiss (Delhal and Demaiffe, 1985) can be considered to date the crystallization age. According to the available T_{DM} ages the magmatic protoliths of the belt are constrained between 2.46 - 2.28 and 2.81 - 2.6 Ga (Alibert and Barbosa, 1992; Sato, 1998). A single zircon (Pb/Pb evaporation) age of 2.07 Ga defines the time of the granulitic metamorphism (Ledru *et al.*, 1994), whereas a Sm/Nd mineral isochron on the high-grade supracrustals (Alibert and Barbosa, 1992, *apud* Barbosa *et al.*, 1998) yields 2.0 Ga. This age corresponds to the isotopic closure of the metamorphic mineral assemblage.

Tectonically, the Itabuna-Atlantic Coast Belt has been interpreted to be an island arc related to a westward subducting Archean/Paleoproterozoic oceanic crust, underneath the Jequié Block (Figueiredo 1989; Barbosa 1990). The model also postulates the existence of a back-arc basin between the magmatic arc and the Jequié continental fragment. The corresponding rocks should be the supracrustal series overthrusted onto the Jequié Block together with the major part of the magmatic belt during a possible arc/continent collision (Barbosa *et al.*, 1998).

Recent observations and radar imagery analysis of the southern part of the Itabuna-Atlantic Coast Belt indicate the existence of an early tectonic episode reflected by E-W flat foliation and sheath-folds with kinematic markers suggesting NNE horizontal transport (Sabaté 1996; Sabaté and Cruz, 1998; Bordini *et al.*, 1999; Sabaté *et al.*, in preparation.). Strong horizontal shearing mostly sinistral produced large N-S to NE-SW vertical foliated bands indicating transpressive to transcurrent regime in granulitic facies conditions. These shear-bands either reworked or at least inflected tangential structures. The coexistence of tangential and transcurrent dynamics is possible, and should be related to a single horizontal transpressive mechanism. This mechanism governs the emplacement of several gabbro-anorthosite and syenite plutons, related to the Transamazonian Orogeny (Sabaté and Cruz 1998; Bordini *et al.*, 1999). The new structural data favour the hypothesis of lithosphere-scale shear mechanisms that would explain the successive emplacement of the various described magmatic suites.

Supracrustal belts

The tectonic evolution of the Jacobina-Mundo Novo and Contendas-Mirante supracrustal belts constitutes the key to understand the successive increments of the Transamazonian collision. These two belts crop out along with the N-S Contendas-Jacobina Lineament (Fig. 4), which constitutes the junction of convergent blocks during the collision and associates these supracrustal belts and the tectonic slices of the Gavião Block. The axial zone of the lineament, as well as several granitic bodies, mainly peraluminous (Sabaté *et al.*, 1990), intrude the contacts of the belts with their adjacent lithotectonic units. The Itapicuru greenstone belt, on the other hand, crops out some distance from the main convergence zone, and has a distinct history in relation to the Transamazonian collision, given by diachronic magmatism from that of the Contendas Jacobina Lineament.

Contendas-Mirante Belt

In addition to an Archean basal volcanic unit the Contendas-Mirante Belt contains a Paleoproterozoic unit, consisting of two members: i) the lower member which associates a thick flysch sequence and metavolcanic rocks, and ii) the upper clastic member, the largest outcroping unit of the belt. This consists of greywacke, pelite and argillaceous rocks with conglomerate layers, metamorphosed in the greenschist to the amphibolite facies. The volcanic units of the lower member consist of metabasalt and andesite of calc-alkaline affinity, and are intercalated within the pelitic rocks. The metarkose of the upper member corresponds to continental fluvio-deltaic platform deposits.

The metasediments have T_{DM} crustal residence ages between 3.5 and 2.39 Ga (Sato, 1998), revealing the heterogeneity of the sediment sources. However, some of the $\varepsilon_{Nd(2.5\,Ga)}$ values suggest a short time of crustal residence for the source material (Marinho *et al.*, 1994b). It is noteworthy that the metapelite yields ages close to 2.5 Ga, which agree within the error with the T_{DM} age of the underlying metavolcanic member (Marinho, 1991; Marinho *et al.*, 1994b). Additional SHRIMP U/Pb ages on three detrital zircon populations from the upper member at Contendas Mirante are between 2.67 - 2.61 Ga and 2.38 - 2.32 Ga, and at 2.17 Ga. The oldest ages found in rounded brown zircon may be related to charnockitic-granulitic sources from the Jequié Block, whereas the 2.3 Ga zircons (magmatic-shaped crystals) may be related to alkaline magmatic rocks sources such as the Pé de Serra Granitoid.

The youngest SHRIMP age (2.17 Ga) is the most representative in the population, referring to idiomorphic zoned zircons (Nutman *et al.*, 1994). These zircons appear to be related to distinct collision plutonism, the products of which were further eroded. If this interpretation is correct, then Transamazonian tectonics have been active since 2.2 Ga at least in the southern part of the Contendas-Jacobina Lineament. Moreover, *in situ* granitic mobilizates derived from partial melting of the Contendas Mirante metapelite beds has given an accurate Rb/Sr isochron age of 2.0 Ga, supporting the connection between the Transamazonian Orogeny and the metamorphic/anatectic mechanisms. The SHRIMP data also suggests that the Jequié Block rocks may constitute the main source of the uppermost member of the Contendas Mirante Basin, which also implies uplift of this block during the Transamazonian Orogeny.

The Contendas Mirante Belt makes up a huge N-S synform branched into smaller belts along its northern and southern edges (Marinho and Sabaté, 1982). Internally, the synform presents a succession of imbricated second order antiforms, complicated by thrust and shear surfaces. The interference of two main co-axial folding episodes is the most evident feature related with coeval shear structures (Sabaté *et al.*, 1980). Moreover, the pervasive and continuous deformation along the belt resulted from E-W shortening, pinched within the zone between the underthrusted Gavião Block and the overthrusted Jequié Block.

The older main deformational episode (D_1) affected the Archean volcanic unit and at least, the lower member of the Paleoproterozoic unit. Both are marked by tight penetrative schistosity, recumbent folds, sheath folds, thrusting shear surfaces, and the kinematic markers indicate a westward vergence of transport related to this episode. On the other hand, the D_2 deformational episode caused the cartographic framework of all belt units, and refolded the D_1 structures. It develops non-cylindrical folds associated with N-S transpressive shear-bands on its edges, as well as N-S axis folds with an upright crenulation schistosity. A westward transport already exists, but remains limited to the D_2 fold of the western part of the belt. In the axial and eastern part this transport and the shortening are compensated by dextral and sinistral strike-slip along N-S shear-bands. The eastern contact of the Contendas Mirante Belt corresponds to the overthrusted Jequié granulitic terranes. It appears to have a low angle dip in the southern part of the belt, with northwestern trajectories of stretching lineations, and upright to the N, with horizontal strike-slip displacement markers.

The Jacobina Belt

These supracrustal rocks are somewhat similar to the clastic sediments of the Contendas Mirante Belt. As in the latter, two Paleoproterozoic members may be distinguished.

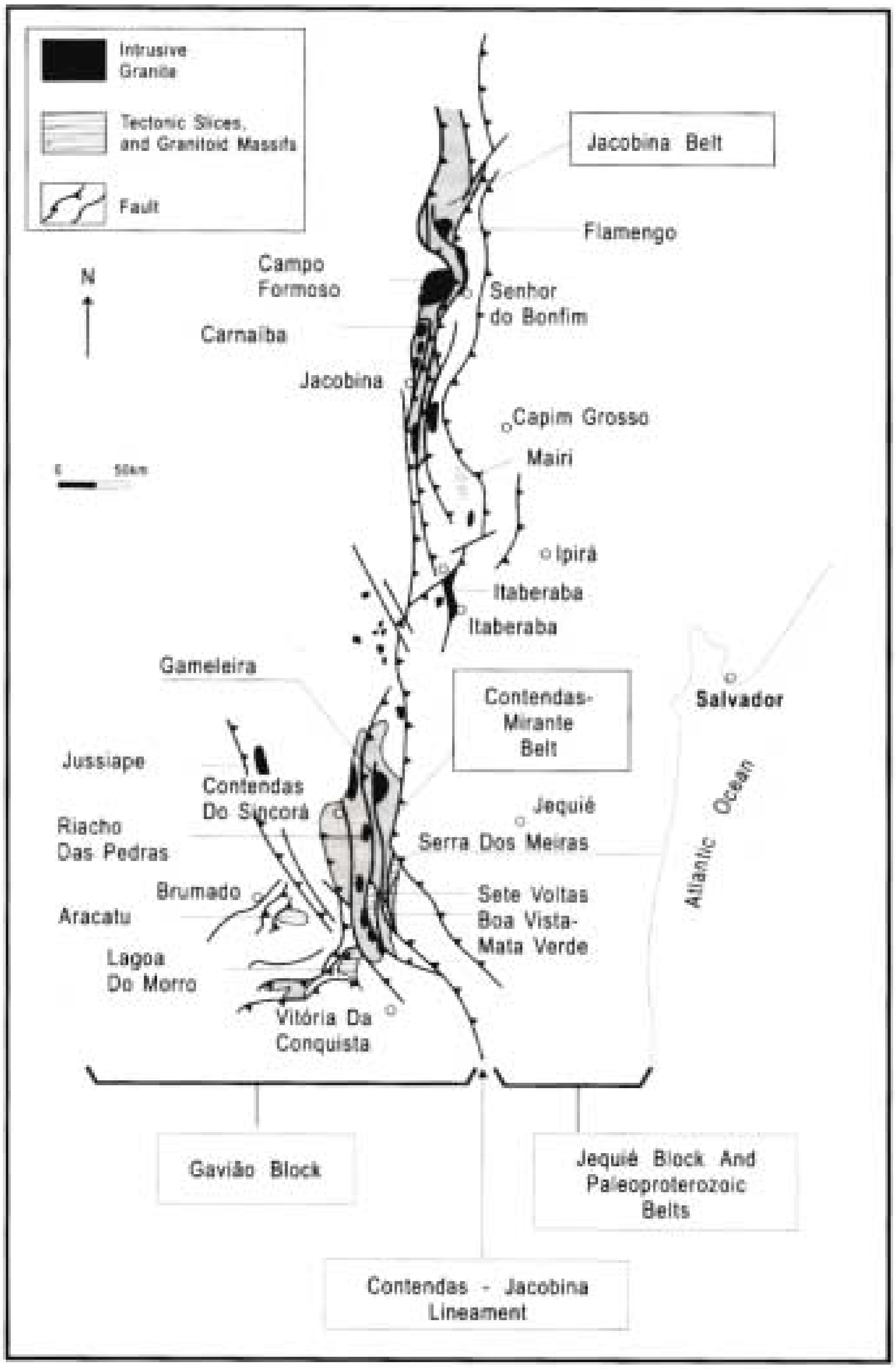


FIGURE 4 - Sketch structural map of the Contendas-Jacobina Lineament (after Sabaté, 1993, 1996). The main tectonic discontinuities (thrust faults and transcurrent faults) makes up the suture zone between the western domain (Gavião Block) and the eastern domain (mainly the Jequié Block and Salvador-Curaçá and Itabuna-Atlantic coastal belts). Also shown are the Jacobina and Contendas-Mirante belts (grey colour) pinched between the two domains, the Gavião Block slices (hatched pattern) imbricated due to the collisional process, as well as the main granitic intrusions (black) distributed along the suture zone.

FIGURE 5 - Sample of the Bouguer anomaly map (Gomes et al., 1994). Area corresponding to the junction between the western domain (Gavião Block with the strongly negative values to the W) and the eastern domain (Jequié Block, Itabuna-Atlantic Coast and Salvador-Curaçá belts, with positive values to E). The roughly N-S contrasting limit (c. - 45 mgal) corresponds to the suture zone (Contendas-Jacobina Lineament) between the two domains. To the E a NE-SW slight gravimetric contrast it is also seen, corresponding to the limit between the Jequié Block and the Itabuna-Atlantic Coast Belt.

A lower member, grouping various formations, is composed of immature conglomerate grading to quartzite, metapelite and metasiltstone and Fe-Mn deposits. Calc-alkaline metatuff and volcanic rocks appear also, locally intercalated into these sediments. An upper member corresponding to the Jacobina Group, includes two detrital formations: a thick quartzite and monomict quartz-pebble conglomerate (Serra do Córrego Formation), well known because of its gold deposits; this is overlain by the thicker quartzitic Rio do Ouro Formation. Several stratigraphic interpretations for the belt, sometimes controversial, where proposed (Cox, 1967, Molinari, 1983, Scarpelli, 1991). The sedimentary characteristics of the upper member are compatible within fluvio-deltaic basins, whereas the lower member contains more mature deposits indicating a marine influence. A new model for the evolution of the Jacobina Basin (Ledru *et al.*, 1997) associating lithological, structural and metamorphic approaches leads to the consideration that the lower formation, according to Couto *et al.* (1978), is clearly older than the detrital Jacobina Group.

Sedimentation of the Jacobina conglomerate units took place in the Paleoproterozoic, as supported by U/Pb zircon data (Mougeot, 1996) that yield a *c.* 2.09 Ga age. A second zircon population from these conglomerate units gives U/Pb ages of 3.4 Ga. This indicates the participation of early Archean sources in the basin deposits, probably the grey gneiss of the Gavião Block.

Located at the contact with the gneiss at the base of the Jacobina sediments, a huge layered meta-ultramafic body mineralized with chromite occurs (Duarte and Fontes, 1986). Other meta-ultrabasic occurrences corresponding to tholeiitic magmas are also intercalated in the detrital formations. The $^{40}Ar/^{39}Ar$ determinations on the main body permit the extrapolation that the metamorphic cooling age of the Jacobina Mundo-Novo Belt took place between 1.98 and 1.93 Ga (Cheilletz *et al.*, 1991).

In like manner to the Contendas Mirante Belt, an overlapping of elongated tectonic lenses constitutes the structural framework of the Jacobina-Mundo Novo Belt. The Mundo Novo lithotectonic unit, identified as a greentone belt remnant (Mascarenhas and Silva, 1994), corresponds, in our view, to an intercalated tectonic slice, which may be related to the Gavião Block. Structural analysis indicates a tectonic stacking of the successive lithostructural units. These pile-up in a reverse order, from the younger Jacobina Group to the W geometrically under the older to the E. We propose that this reverse order may be repeated in the Jacobina Group. The whole belt is marked by a transpressive regime within the external N-S major strike-slip sinistral shear bands limiting the belt. The resulting shortening is responsible for internal secondary order strike-slip and folding as well as the westward transport of the lithotectonic units through a system of thrust zones. Hence the Jacobina Group has been thrusted over the terranes of the Gavião Block, and the older deposits of the lower detrital member are thrust over the Jacobina Group. As in the Contendas-Mirante Belt, the lower member of the Jacobina-Mundo Novo Belt underwent two main phases of deformation D_1 and D_2 with the analogous respective styles. The younger Jacobina Group underwent only the S_2 foliation and its related lineations.

The tectonic evolution of the Jacobina Belt is contemporary with the sedimentary evolution (Ledru *et al.*, 1997), and at least five successive stages of basin development are defined. They correspond to lithotectonic units bounded by thrust faults. The last stage thrusting events are dated in the range 1.94 - 1.92 Ga ($^{40}Ar/^{39}Ar$, Ledru *et al.*, 1997). The sedimentary processes from E to W, with reworking of older deposits, were accompanied by the westward progression of the tectonic evolution. To the E, the thickening results from the subducting continental slab under the Jequié Block added to the shortening of the latter. The metamorphic conditions reached equilibrium near the P-T triple point of alumina-silicates (500 - 600 °C and 4.5 - 5 kbar), whereas sapphirine in the overthrusted granulite to the E was recently found (Leite, personal communication).

In both the Jacobina-Mundo Novo and Contendas Mirante belts, successive tectonic prisms were expelled from the subducting zone and piled up to form the tectonic stacking of the belts. The resulting thickening, increased by D_2 shortening, provides an explanation for the preferential location of the peraluminous granitic intrusions.

Finally, the evolution of the Jacobina supracrustal rocks from the state of basin to the state of belt, as for the Contendas-Mirante Basin is compatible with foreland basins, and gets the role of healing the suture zone between two continental segments. The suture signature is supported by a strong gravimetric contrast (Fig. 5). The bouguer anomaly (Gomes *et al.*, 1994) shows a narrow line (in the range - 44 / - 46 mgal) separating the positive values (+ 10 mgal) of the granulite terranes (Jequié Block and granulite belts) to the E from the strong negative values (- 110 mgal) of the Gavião Block and its Middle and Neoproterozoic cover to the W. This limit follows faithfully the Contendas-Jacobina Lineament (Fig. 3).

Peraluminous plutonism

A string of granitoid plutons crops out along the axial zone of the Contendas Mirante and Jacobina-Mundo Novo belts, delineating up to 500 km of the Contendas-Jacobina Lineament. They either intrude the supracrustal rocks or follow the contacts between both the belts and the overthrusting granulite terranes, or the Archean tectonic slices imbricated onto the supracrustal rocks (Fig. 4).

These plutons are peraluminous two-mica leucogranites. Some of these intrusions are garnet-bearing and a few of them contain cordierite. Their geochemical and isotope signatures indicate a crustal source which may be found either in the supracrustal rocks or in the Archean gneiss (Cuney *et al.*, 1990, Sabaté *et al.*, 1990). According to these authors, the geochemical trends of the granitoid plutons (in spite of a lower peraluminous content compared to the Phanerozoic equivalents) are similar to those of the Hercynian Belt of the French Central Massif and the Himalayan leucogranites. On the other hand, hydrothermal activity related to these peraluminous granitoid plutons may have given origin to metasomatic emerald deposits, as occurs within the Campo Formoso Granite (Rudowski *et al.*, 1987).

The granite emplacements are closely associated with the tectonic evolution of the Contendas Mirante and Jacobina belts. The early bodies emplaced during the collision processes (late-D_1, syn-D_2), and emplacement continued till the termination of the last transcurrent increments (late to post-D_2). Therefore the late bodies mark the end of the Transamazonian collision tectono-metamorphic processes. Within the Contendas-Mirante Belt, the granitoid rocks have Rb/Sr isochron ages in the time interval 1.97 - 1.93 Ga from the earliest to the youngest body. The T_{DM} ages are between 3.28 and 2.55 Ga. Tectonically, the older granites are pre to syn-kinematic to the strike-slip related D_2 shearing in the southern part of the belt, whereas the younger bodies are post-tectonic or at least post-ductile relative to the D_2 episode (Sabaté *et al.*, 1990; Marinho, 1991). The granitoid plutons within the Jacobina Belt have Rb/Sr isochron ages between 1.97 - 1.9 Ga (Torquato *et al.*, 1978; Sabaté *et al.*, 1990), and their T_{DM} ages are between 3.27 and 2.66 Ga. The large Campo Formoso Massif gave a U/Pb zircon age of 1.97 Ga (Mougeot, 1996). They were emplaced at the end of the transpressive sinistral regime of the Jacobina Belt (Sabaté, 1996).

The Rio Itapicuru greenstone belt and related plutonism

The Rio Itapicuru Belt (Fig. 2) crops out within the Serrinha Block, in a framework controlled by emplacement of calc-alkaline to alkaline plutons of Paleoproterozoic age. The belt is elongated N-S with structural trends parallel to the axis. However, in the southern part the polyphase evolution produced E-W structures (Reinhardt and Davison, 1990; Teixeira *et al.*, 1990).

Lithologically, the Rio Itapicuru Belt comprises a pile of oceanic basalt units, andesitic and rhyodicitic lavas (Davison *et al.*, 1988; Silva 1992) together with and metarkose and conglomerate (Kishida and Riccio, 1980). The whole pile has been affected by the Transamazonian Orogeny, in like manner to the Serrinha Block. As in the other supracrustal belts, there occurred two deformational events D_1 and D_2 (Chauvet *et al.*, 1997). The first event results from an E-W shortening, marked by low angle foliation, recumbent and sheath folds, NE-SW trending lineations and thrust direction toward the SE. The second originated sinistral strike-slip shear bands. These events which developed under the transition greenschist to amphibolite facies, affected both the Rio Itapucuru supracrustal rocks and the granitic plutons.

An E-W orthogneiss Barrocos Massif in the southern part of the belt is syn-kinematic with the D_1 tectonics. It gives a Pb/Pb zircon evaporation age of 2.13 Ga, interpreted to be the crystallization time. On the other hand, the N-S elongated plutons are contemporaneous with the D_2 phase. These bodies indicate ages at 2.1 Ga by the same method (Chauvet *et al.*, 1997). The tangential D_1 event with its SE vergence is confined in the southeastern part of the belt. The D_2 development and granitic emplacement are closely related to a roughly N-S main shear zone that separates the domains. To the W of the shear zone, the D_2 structures dominate, whereas D_1 structures dominate to the SE (Chauvet *et al.*, 1997).

Granite to granodiorite diapirs of calc-alkaline affinity were emplaced into the Rio Itapicuru greenstone belt at *c.* 2.0 Ga (U/Pb age, Gaál *et al.*, 1987). They generally exhibit a N-S elongated shape, and are foliated at their margins, when the core remains isotropic (Matos and Davison, 1988). Finally, a pendant of banded gneiss occurring in the strongly foliated margin of one of these bodies (Ambrósio Dome), gave a U/Pb age of 2.93 Ga, confirming the existence of Archean sialic basement near the source region of the granite (Gaál *et al.*, 1987; Davison *et al.*, 1988). An analogous interpretation has been recently proposed for the Nordestina Massif. The external zone yields a Pb/Pb zircon age of 2.1 Ga, whereas the core yields an age of 2.0 Ga (Cruz Filho *et al.*, 1999). The supracrustal protoliths are dated at *c.* 2.21 Ga (Brito Neves *et al.*, 1980; Silva, 1987; Silva and Vidal, 1992).

Late emplacement of several plutons involved a change in genetic evolution, from low to high-K calc-alkaline granites (monzonitic trend) to high to ultra-K alkaline (shoshonitic trend). At least, five groups of intrusions where identified (Rios, 1998; Rios *et al.*, 1998). An older Rb/Sr isochron with an age of 2.03 Ga (Sabaté *et al.*, 1990) considered to be the emplacement time of a monzonitic body must be revised taking into account the new single zircon Pb/Pb age of 2.1 Ga (Oliveira *et al.*, 1998), which is probably the best estimate of the crystallization age. This age also constrains the cratonization stage to after the Transamazonian processes within the Serrinha Block where the pluton is exposed. On the other hand, the structure of this body is quite exclusively magmatic, but weakly overprinted by late magmatic shearing. This shearing corresponds to the final phase of penetrative deformation occurring in the Rio Itapicuru greenstone belt (Nascimento, 1996).

Dynamics of the Transamazonian collision

The field observations in conjunction with the structural and radiometric data converge to identify homogeneous tensors at the scale of the northern part of the SFC. For the Serrinha Block, in relation to the neighbouring Salvador-Curaçá Belt, analogous tectonic tensors may be suggested, similar to those observed in all the other lithostructural units involved in the main collision axis that is the Contendas-Jacobina Lineament.

Our model

In the precedent items we have summarized most of the available information in the evolutionary context of the northern part of the SFC terrane: lithology, petrogenesis, geochronology, structure, kinetics, and dynamics. The integrated information and the corresponding dynamics lead to a tectonic model (Table I; Fig. 3) well constrained in the light of present knowledge.

The geological evidence favours the existence of two Archean protocontinents that determined the geodynamic parameters of the succeeding Transamazonian collision. As already shown, each of these fragments has its own diachronic Archean evolution. The crust of the Gavião Block was formed mainly from 3.4 to 3.0 Ga, whereas the continental accretion of the Jequié Block started later at 2.9 Ga. The respective components and thermodynamic genetic conditions of these fragments are clearly distinct. Besides, the Neoarchean history of the Gavião Block records unusual plutonic events, but these are not known to be related to any orogenic process that has been identified. Within the block, juvenile accreted plutonism predominates relative to the supracrustal rocks. The succeeding episodes either include crustal contamination of the juvenile magmas, or their generation through partial melting of the pre-existing crust. Such episode took place at roughly equal intervals *c.* 300 Ma (3.7 Ga, 3.4 Ga, 3.1 Ga) with the possibly secondary sub-periods of 100 Ma. In Paleoproterozoic times, the crustal growth is, on one hand, related to the recycling of the crustal components, and on the other hand, it links principally with the 2.1 Ga Transamazonian collision tectonics, the orogenic closure of which finally led to the consolidation of the northern part of the SFC.

Early Paleoproterozoic episodes

As already seen, the evolution of the northern part of the SFC broadly follows the worldwide geological characteristics of the Archean/Paleoproterozoic transition (Kröner, 1984). Following numerical models for mantle convection (Brunet and Machetel, 1998), we postulate, that between 2.5 Ga and 2.4 Ga, there occurred the progressive downfall of a dense, cold lithosphere, the axis of which roughly corresponds to the actual Contendas-Jacobina Lineament. The dynamics resulted from the convergence between the two existing continental fragments. It is suggested that three main tectonic regimes occurred on both the western and eastern sides of the downfall axis, as follow:

1 - Development of a *c.* 2.5 Ga extensive regime at the nearest eastern margin of the Gavião Block. This regime induced the crustal contamination of both the calc-alkaline volcanism and the contemporary Rio Jacaré mafic-ultramafic plutonism.

2 - Development of a *c.* 2.5 Ga extensive regime at the nearest western margin of the Jequié Block. This regime originated the deep alkaline-subalkaline Pé de Serra plutons, attesting to crustal contribution from the Jequié charnockite-enderbite sources.

3 - Development of a transpressive regime that generated the E-W shortening in the western part of the Jequié Block marked by stretching and folding of the subalkaline plutons. This regime produced transcurrent shearing with pull-apart opening, which allowed emplacement of the alkaline plutons at *c.* 2.3 Ga (Pé de Serra). Compressive tensors in the Jequié Block should have been active since 2.3 Ga.

Pre-collisional episodes

Due to the 2.5 Ga downfall of voluminous oceanic material the two continental fragments tend to converge. The consecutive lack of lithosphere material is compensated by several magmatic intrusions/extrusions that counteract the mass excess sinking into the mantle. The first magmatic products resulted from the convergence emplaced relatively close to the downfall axis. While the geodynamic evolution continued there occurred further emplacements from the axis. This can be seen in the magmatism occurring within the Salvador-Curaçá and Itabuna-Atlantic Coast belts to the E, and the Guanambi-Urandi Batholith to the W. The succeeding lithosphere flexure (at the eastern margin of the Gavião Continent) finally led to formation of the foreland basins, like the Jacobina and the Contendas basins. Due to such a flexure, extensive mechanisms may have also taken place in these basins during an early evolutionary stage, which resulted in basic magmatic intercalations and related chemical deposits, such as the Fe-Mn formations. In the same way, continental margin-type volcanism originated subordinately.

Collisional episodes

The Transamazonian collision was contemporary with the subduction of the western continental slab (Gavião Block) underneath the eastern one (Jequié Block). This stage developed earlier than the late detrital deposits dated at *c.* 2.1 Ga. Subducted metamorphosed sediments of the foreland basins were ejected to form successive tectonic prisms of the supracrustal rocks. Remnants of the subducting continental slab were also cast out; being intercalated as tectonic slices in the growing belt. This mechanism agrees well with the metamorphic reverse zoning of both the Jacobina and the Contendas-Mirante belts. The following step corresponds to an E-W shortening of these belts between the two thickened continental segments. It also embraces the later detrital deposits (since 2.1 Ga), the older subducted sediments uplifted as tectonic prisms, and the intercalated tectonic slices of the subducted Gavião slab.

Outside the basins, the relative movement of the Gavião and Jequié continents depends on the transpressive tensors. The both segments record only a transpressive regime, which is manifest through left-shearing. This regime governs from the beginning of the Transamazonian collision to the emplacement of the late intrusions of the Salvador-Curaçá Belt (2.1 Ga to 1.9 Ga).

A lithosphere-scale shear zone probably remained active for a long period in order to originate the successive magmatism (from trondhjemitic-type to K-rich alkaline) in this belt. An analogous mechanism may be suggested for the granitic intrusives within the Rio Itapicuru Belt. The tectonic evolution of this greenstone belt was completed by 2.1 - 2.06 Ga when the collision began in the Contendas-Jacobina Lineament. This suggests otherwise that the Archean Serrinha Block may represent an exotic segment in relation to the lithotectonic units directly involved in the Transamazonian Orogeny, which built the collisional Jacobina-Contendas chain. All together, this suggests that the Transamazonian Orogeny may be diachronic in the different orogenic chains in the northern part of the SFC.

After tectonic stabilization of the Rio Itapicuru Belt, peraluminous magmatism was restricted to the collision zone, from *c.* 2.0 to 1.9 Ga. Structurally, this latest magmatic phase can be interpreted in terms of a transition regime from transpressive to purely transcurrent. This implies that the new continent attained enough thickness and cohesion to affect the transfer of mobile zones to other regions. This happened effectively in the western part of the Gavião Block, and was mirrored by the tectonic evolution of the Paleo to Mesoproterozoic Espinhaço Belt.

Tectonic correlation with the Congo-Gabon Craton (Africa)

In the paleotectonic scenario of western Gondwana the northern part of the SFC exhibits remarkable geological similarities with the contiguous Congo-Gabon Craton (CGC), particularly for the Archean and Paleoproterozoic evolution. Intracontinental rift basins (Espinhaço Supergroup and Maiumbien Series), also cut the crystalline basement in both cratons, besides being extensively covered by coeval Neoproterozoic sediments. Nevertheless, the geochronological evolution of the CGC units is, in general, not as well known as those of the SFC.

The oldest terranes of the CCG comprise amphibolite and granulite grade rocks (North Gabon and Chaillu blocks) that show radiometric ages between 3.0 and 2.6 Ga. These rocks are partially overlain by supracrustal belts that include volcanic and sedimentary components, folded and metamorphosed during the Eburnian (Transamazonian) Orogeny (Table 2). The scenario has led to tentative comparisons between the CCG basement rocks with at least a part of the Jequié granulitic terrane and the Paleoproterozoic supracrustal belts of the northern part of the SFC (Figueiredo, 1989).

Regarding the evolution, the most relevant geological similarities between the northern part of the SFC and the CCG concerns that which occurred in Paleoproterozoic time. In the CGC, the Paleoproterozoic evolution of the West Central African Belt (Feybesse *et al.*, 1998) supports an accretionary model from 2.5 to 2.0 Ga, which is similar in many aspects to that herein presented for part of the SFC (Table 2). Lithological associations having approximately the same age occur in the two cratons, indicating similar geodynamic conditions. This scenario suggests that the major convergence of Archean continental segments and Paleoproterozoic terranes took place at *c.* 2.0 Ga. Although the Paleoproterozoic evolution within either the SFC or the CGC appears clearly heterochronous, a rough symmetry of structure and dynamics can be observed. The CGC presents an eastward vergence (Ledru *et al.*, 1994; Feybesse *et al.*, 1998), whereas we have shown the general westward vergence of transport within the terranes of the northern part of the SFC (previous items).

The main axial zone of symmetry seems to be the granulite domains of the Salvador-Curaçá and Itabuna-Atlantic Coast belts. The Gavião continental block (northern part of the SFC) and the Archean Chaillu Block (CGC) occupy symmetric and analogous positions in the accretionary build-up, both acting as uplifted crustal segments. Therefore, they induced development of foreland basins also in a symmetric position, as for example, the Jacobina-Contendas and Francevillian basins, respectively. Moreover, subduction of the continental crust led to crustal thickening of both Paleoproterozoic units. The true collision tectonic mechanisms developed from 2.17 to 1.90 Ga within the northern part of the SFC, as already seen, and from 2.15 to 2.04 Ga in the CGC, with analogous but heterochronous progressive evolution of deposits and respectively concomitant tectonics in the foreland basins.

The Southern Part of the São Francisco Craton

The Archean terranes within the southernmost portion of the SFC comprise mainly gneissic partly migmatized rocks, greenstone belts, as well as granitoid plutons and mafic and mafic-ultramafic intrusions (Fig. 6). These terranes are considered to be part of a platform stabilized at 2.6 Ga (Noce *et al.*, 1998). An arc-shaped collisional belt (Mineiro Belt) developed at the margins of the former Archean platform (Teixeira *et al.*, 1996; Alkmin and Marshak, 1998) during the Transamazonian Orogeny (*c.* 2.16 - 2.0 Ga), mirrored by granitoid plutons and flysh-type deposits, as well as mafic dykes. The Archean geological features in the study area are now reduced to a small exposure (*c.* 10 000 km^2) within the very inner part of the former Archean platform (Fig. 6), due to the Paleo to Mesoproterozoic Espinhaço thrust-rift episodes, extensive Neoproterozoic sedimentation (Bambuí Group), as well as the Brasiliano collage against the SFC.

Detailed geological mapping is mainly available for the Au-bearing Rio das Velhas Archean greenstone belt and the Lake Superior-type banded-iron formation units of the Paleoproterozoic Minas Supergroup. These rock assemblages crop out in the Quadrilátero Ferrífero - QF that makes up a folded and fault structure of *c.* 7000 km^2 (Dorr, 1969; Herz, 1970).

The geological knowledge of the areas (Fig. 6) in which gneissic and granitoid rocks crop out is still fragmentary, due to the paucity of systematic mapping and a thick regolith cover. This limitation together with geochemical, petrographical and geochronological data has led to definition of several metamorphic complexes: Belo Horizonte, Bonfim, Campo Belo, Passa-Tempo, Bação, Caeté, and Florestal. Although some of these complexes may represent distinct crustal fragments welded during the Neoarchean, as it will be discussed later on the basis of U/Pb and Sm/Nd data and geochemistry, their boundaries cannot be associated with any prominent geological feature because the polyphase deformation and metamorphic overprints. As a consequence, we summarize the composition, structure and isotope data of the Archean crust as a whole, in terms the lithological associations.

Archean Lithological Associations

Gneiss, migmatite and granitoid

Amphibolite facies TTG gneiss and migmatite make up the largest part of the Archean crust (Fig. 6) intruded by tonalitic to granitic plutons. The best studied gneissic terranes situated in the QF, where they occur in domes (Bonfim, Belo Horizonte, Pará de Minas) surrounded by troughs containing the Rio das Velhas greenstone belt and the Minas Supergroup (Alkmim and Marshak, 1998). Supracrustal relicts are also present to the W and SW of the QF (Campo Belo-Itapecerica-Cláudio).

The Bonfim Dome comprises a granite-greenstone terrane (2.78 - 2.7 Ga) that originated during the Rio das Velhas Orogeny (Machado and Carneiro, 1992; Carneiro *et al.* 1998b). The geological framework includes two different gneiss (Alberto Flores and Souza Noschese), amphibolite units (Paraopeba and Candeias), and the Samambaia Tonalite and Brumadinho Granite (Fig. 7) (Machado and Carneiro, 1992; Carneiro *et al.*, 1998b). The oldest lithostratigraphic unit is the Alberto Flores Gneiss that exhibits a well-developed N-S mylonitic foliation. This

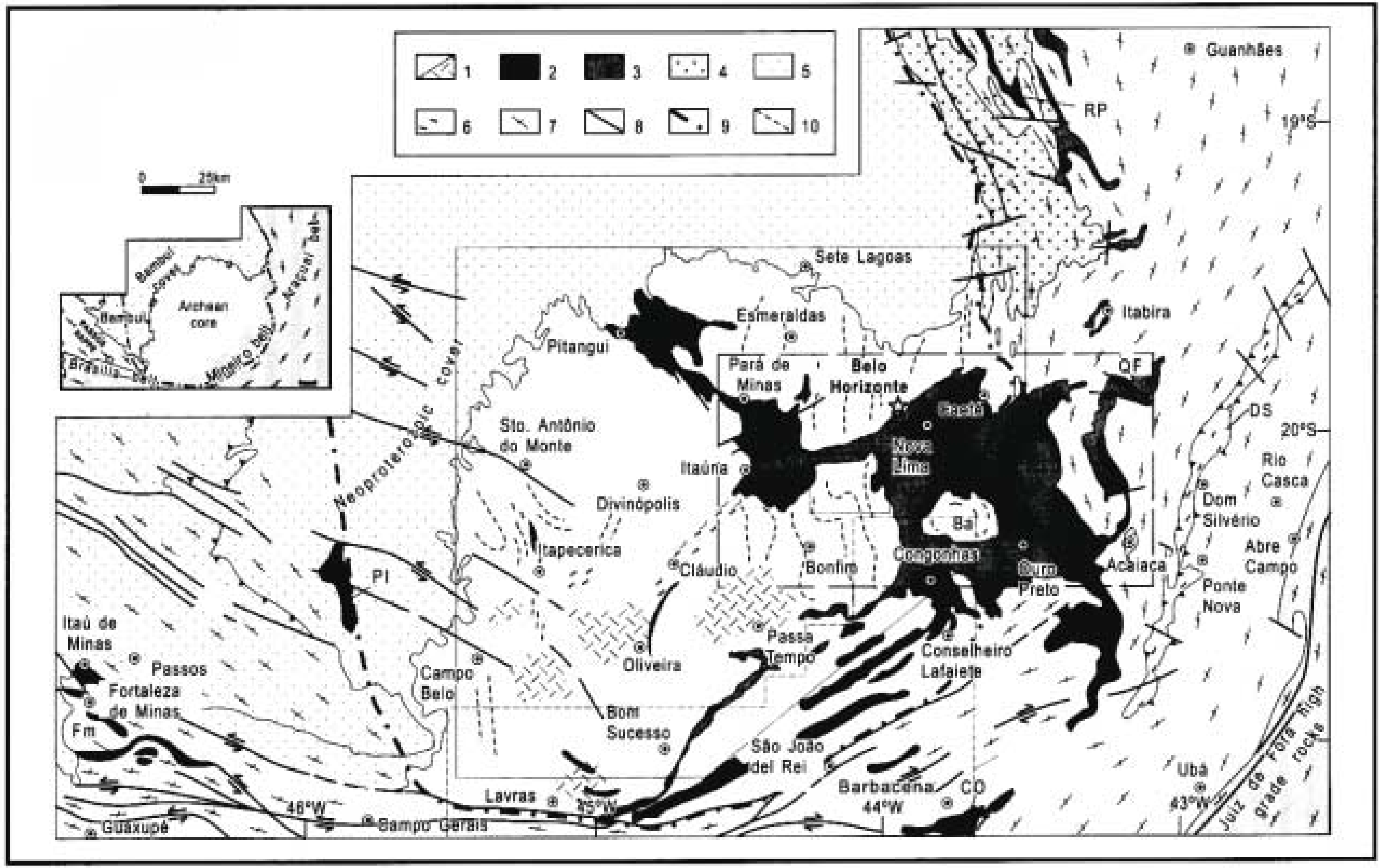


FIGURE 6 - Geological outline of the southern part of the SFC, showing the Archean and Paleoproterozoic terranes, as well as the Neoproterozoic marginal mobile belts (adapted from Pedrosa Soares et al., 1994). Archean: 1= granitoid-gneiss-migmatite rocks and granulite (chaotic symbol), partially reworked in the Paleoproterozoic; 2= greenstone belts; 3= Minas Supergroup (Paleoproterozoic); 4= Espinhaço rift system (Paleo to Mesoproterozoic);5= Bambuí cover (Neoproterozoic); 6= shear zone/shear belt; 7= Brasiliano structures of the marginal belts (Araçuaí; southern Brasília and Passos nappe), partly overprinting crystalline basement rocks (> 1.8 Ga) - see inset; 8= major faults; 9= Limit of the craton. 10= Gneiss foliation. Keys: PI, RP, FM, CO= Piumhí, Rio Paraúna, Fortaleza de Minas and Congonhas greenstone belts; DS= Dom Silvério Group (Neoproterozoic).

gneiss underwent remobilization at 2.772 ± 0.006 Ga (U/Pb age on a single zircon overgrowth). The zircon core gives a $^{207}Pb/^{206}Pb$ date of 2.92 Ga, interpreted to be a minimum age for the protolith, in coherence with the Sm/Nd T_{DM} model age of 3.15 Ga reported for the same rock (Teixeira *et al.*, 1996). The weakly foliated Souza Noschese granitic gneiss, which is intrusive into the Alberto Flores Gneiss, displays a preliminary U/Pb discordia line of 2.775 ± 0.077 Ga (F. Chemale Jr., written communication, 1998). The Samambaia calc-alkaline intrusive tonalite yields a U/Pb zircon age of 2.78 + 0.003 / - 0.002 Ga (zircon abraded) interpreted as the magmatic crystallization age. Two T_{DM} ages (2.94 and 2.8 Ga) indicate a short crust-residence for the protoliths. Titanite ages on the same rock at 2.774 ± 0.006 Ga (Carneiro *et al.*, 1998a) constrain the time of the amphibolite grade metamorphism in Bonfim Dome. The emplacement of this tonalite was accompanied by mafic magmatism (now amphibolite) that yield slightly positive (+0.6; +0.1) $\varepsilon_{Nd(2.78 Ga)}$ values (Carneiro *et al.*, 1998b).

The Bonfim felsic gneiss has a geochemical composition comparable with TTG suites, and a REE pattern compatible with the high Al_2O_3, REE envelope pattern of most Neoarchean tonalite/trondhjemite suites world-wide (Fig. 8). The Souza Noschese Gneiss in particular shows a geochemical signature (Fig. 7) suggestive of derivation by the partial melting of trondhjemitic crust (Carneiro *et al.*, 1998b). Limited chemical data for the Paraopeba and Candeias amphibolite units demonstrate their tholeiitic affinity, and the REE patterns are similar and have the same $La_N / Yb_N = 2$. In addition, the average composition of these mafic suites is comparable with mid-ocean ridge basalts, as suggested by typical enrichement in the K, Rb, Ba and Nb content (Fig. 9). As a whole, the geological features of the Bonfim felsic and mafic rocks, together with their geochemical and isotope signatures are compatible with a convergent margin tectonic setting for the Neoarchean evolution (Carneiro *et al.*, 1998a).

The youngest Archean magmatic activity in the Bonfim dome is represented by intrusive plutons, two of which show U/Pb crystallization ages between 2.72 and 2.7 Ga. The Mamona Granitoid (Carneiro *et al.*, 1998b) is part of a composite pluton, showing a granodioritic to granitic composition and well-preserved igneous texture (Lacerda *et al.*, 1996). This pluton shows a U/Pb age of 2.721 ± 0.003 Ga (Machado *et al.*, 1992). The Brumadinho Granite (U/Pb zircon age of 2.703 + 0.024/ - 0.02 Ga) comprises weakly foliated and folded dykes that truncate the regional, steeply dipping N-S foliation of the Bonfim TTG-gneiss (Figs. 8 and 10). This is a clear time constraint that this N-S flow plane developed during the 2.78 - 2.7 Ga crustal evolution of the Bonfim rocks.

In the Belo Horizonte Dome, N and W of the QF, the dominant rock-type is a migmatitic-banded gneiss with schlieren and stromatic structures and mafic enclaves. The regional foliation varies from N-S to NNE-SSW, dips 40 - 60° to the W and WNW, being broadly coeval with migmatization, as indicated by leucosomes both parallel and secant to the regional foliation (Noce *et al.*, 1998). The migmatization event was dated at 2.86 + 0.014/ - 0.017 Ga (U/Pb zircon age; Noce *et al.*, 1998) which compares well with the Rb/Sr whole-rock isochron ages available in different places of the Belo Horizonte dome (Teixeira, 1993). The Belo Horizonte rocks show therefore an older tectonic event than that reported at the Bonfim Dome (see above). The T_{DM} ages fall in the range 3.4 - 3.1 Ga, indicating a protracted evolution for the Belo Horizonte Dome. Chemical analyses of the least migmatized gneiss reveal a predominant trondhjemitic composition. The REE mean patterns are comparable with that shown by the Alberto Flores gneiss (see above), as well as by Archean high Al_2O_3 tonalite/ trondhjemite wordwide (Fig. 11) (Noce *et al.*, 1997).

Evidence for two main Archean granitoid pulses occurs N of the QF (Fig. 12). The older is manifest by the Caeté granodiorite pluton and the Mateus Leme Granite that show U/Pb emplacement ages at 2.776 + 0.007/ - 0.006 Ga (Machado *et al.*, 1992a, b) and 2.755 + 0.014/ - 0.013 Ga (Romano, 1989), respectively. A T_{DM} age of 2.78 Ga for the latter (Romano, 1989) compares well with the Sm/Nd dates reported for the Samambaia tonalite occurring to the S (Bonfim Dome; see above). The second episode, which included calc-alkaline and peraluminous granite to tonalite, was dated at 2.712 + 0.005/ - 0.004 Ga (Santa Luzia Granite), 2.714 ± 0.002 Ga (Capelinha Trondhjemite, Oliveira, 1999) and 2.698 ± 0.018 Ga (Ibirité Granodiorite, F. Chemale Jr., written communication. 1993). These intrusions display a well-developed N to NW foliation, which contrasts with the relatively undeformed *c.* 2.7 Ga Brumadinho and Mamona plutons of the Bonfim Dome.

The QF was also affected by a third Neoarchean granitic pulse (2.612 - 2.555 Ga). The emplacement of these plutons was probably associated with reactivation of earlier discontinuities as they occur at the margins of the Rio das Velhas greenstone belt (Endo and Machado, 1998). The examples are the Caio Martins Granodiorite (2.593 + 0.018/ - 0.019 Ga; Romano 1989), Salto do Paraopeba Granite (2.612 ± 0.005 Ga; Noce *et al.*, 1997, 1998), as well as the Itabirito Granite (2.567 ± 0.008 Ga) and aplite (2.555 ± 0.024 Ga; Fig. 12).

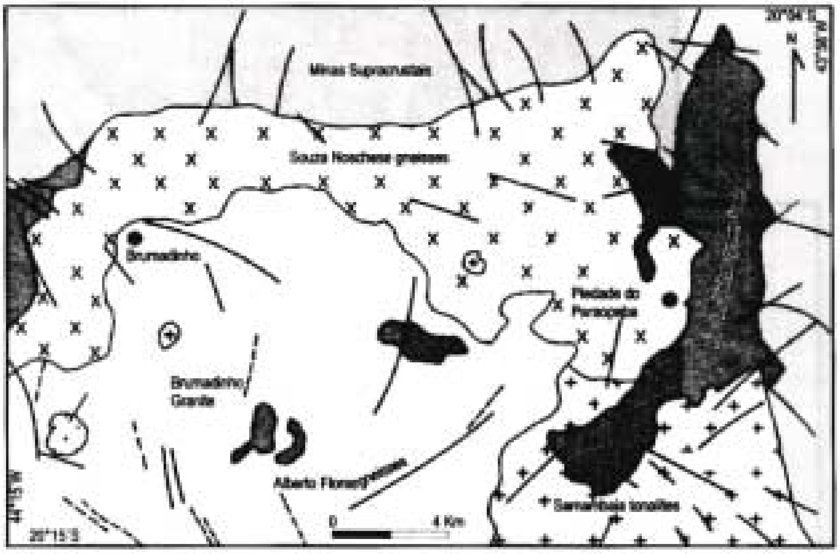
Tectonically, the oldest granitoid intrusions (2.78 Ga) are coeval with the Rio das Velhas greenstone belt volcanism, and their chemical and isotope signatures indicate both mantle and crustal contributions to the magma (Carneiro *et al.*, 1998b). The granitoid plutons post-dating the 2.78 Ga event are mainly crust-derived, as suggested by variation of the T_{DM} ages between 2.77 Ga (Caio Martins) and 3.1 - 2.94 Ga (Brumadinho), and the isotope evidence is also consistent with derivation from different crustal components. In particular, the youngest generation of granites (2.612 - 2.555 Ga) provides a chronological marker for the succeeding Paleoproterozoic evolution in the southern part of the SFC. Limited geochemical data on these granitoid plutons indicate the following: (i) a part of the plutons are K-rich calc-alkaline granites showing Sr, Nb and Ti depletion (Santa Luzia, General Carneiro; Noce *et al.*, 1997; Carneiro *et al.*, 1998b); (ii) the Florestal (Caio Martins) Granitoid is a sub-alkaline rock, enriched in Th, Ce and Sm compared to the Nb and Zr content, but the MgO / TiO_2 ratio is typical of calc-alkaline granitic suites as well (Romano *et al.*, 1995); (iii) the Capelinha Pluton is a peraluminous calc-alkaline intrusion of trondhjemitic composition, displaying a fractionated REE pattern (Oliveira *et al.*, 1998).

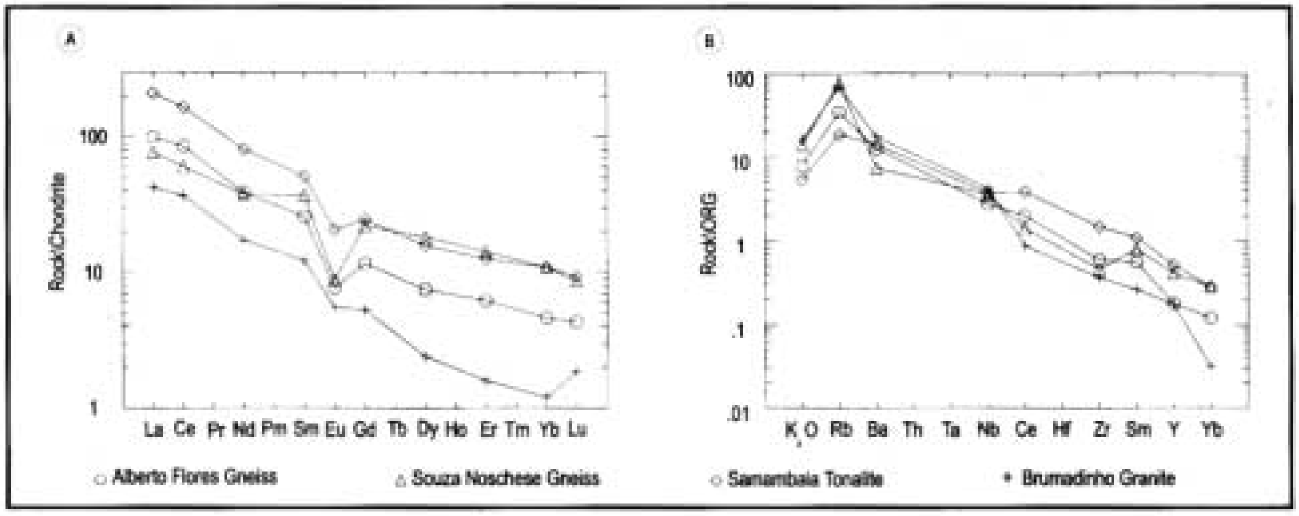
At the western part of the exposed Archean crust, the main rock-units are partly migmatized TTG suites intruded

FIGURE 7 - Sketch map of the Bonfim Dome showing the main geological units, and the Paraopeba amphibolite (dashed lines). Symbols as in Fig. 6. See text for details.

FIGURE 8 - a) Chondrite-normalized REE patterns (after Massuda et al., 1973) for the gneissic units of Bonfim area. b) ORG-normalized spidergram for the felsic units (after Pearce et al., 1984).

FIGURE 9 - a) Chondrite-normalized REE patterns values (after Massuda et al., 1973) for the mafic suites of Bonfim area; b) MORB-normalized spidergram of the mafic suites (after Pearce, 1982, 1983).





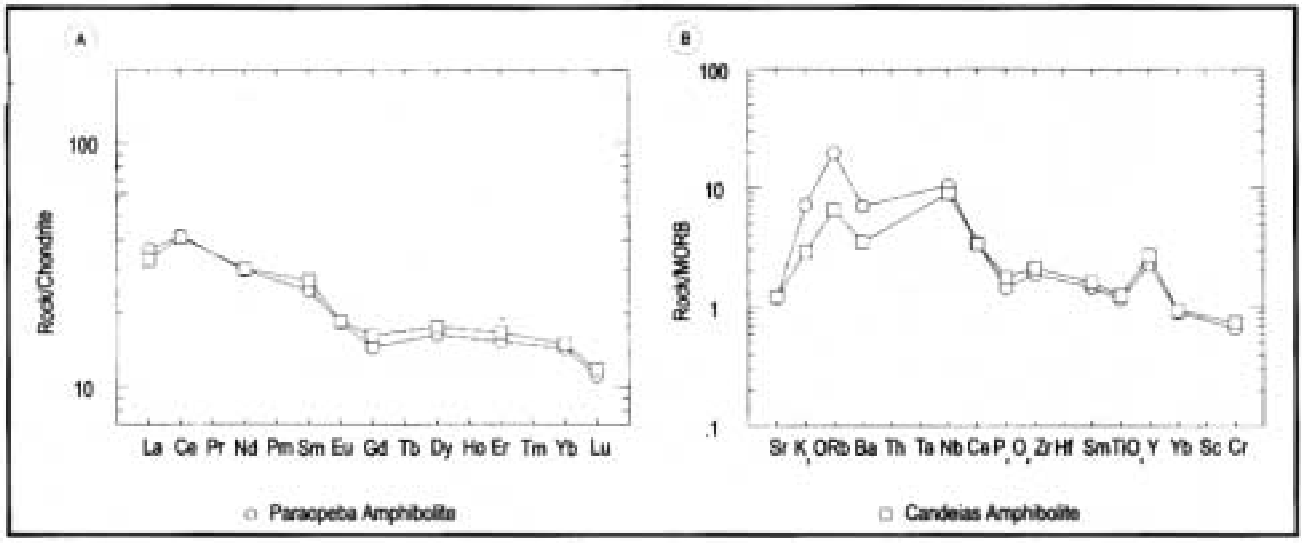


FIGURE 10 - Cross-cutting relationships between the Alberto Flores trondhjemitic banded gneiss and Brumadinho late-Archean granite (grey) - Bonfim Dome.

by granitoid plutons and mafic dykes, as well as amphibolite and supracrustal relicts some of which bearing graphite deposits. Regional metamorphism is amphibolite grade, but near Campo Belo, Lavras, Oliveira and Passa Tempo granulite facies gneiss predominate (Fig. 6). Near Claudio-Itapecerica-Oliveira, the migmatitic gneiss bearing amphibolitic boudins and minor ultramafic relicts are affected by NE-SW and NW-SE shear zones, along which supracrustal rocks (garnet-sillimanite quartzite and schist, and bif) are subvertical and exhibit mylonitic textures. The gneiss foliation is a second-order structure, considered to be related with the main deformation event in the area during which migmatites were also formed. Moreover, earlier and latter phases of deformation attest the polyciclic tectonic framework (Oliveira, 1999a, b).

In Campo Belo a migmatite, dated by SHRIMP U/Pb geochronology, defined the presence of three generations of melts (Teixeira *et al.* 1998a). The oldest zircon crystals yielded 3.205 ± 0.017 Ga which is interpreted to be a relict age component in the migmatite. This age compares well with the oldest Sm/Nd T_{DM} ages available for the Campo Belo orthogneiss up to 3.2 Ga. A second group of zircon crystals showed a 3.047 ± 0.025 Ga age, and may be related to the main magmatic event that is represented by less-migmatized orthogneiss exposed nearby. In the vicinity of Campo Belo, these orthogneiss have $\varepsilon_{Nd(t)}$ parameters close to zero and Pb mantle-like single stage signature ($^{238}U/^{204}Pb$ = 8.18; Teixeira *et al.*, 1996). Except for a few older T_{DM} ages, the orthogneiss show broadly comparable T_{DM} ages between 3.07 Ga and 2.90 Ga, indicating the main period of the accretion. Additional Rb/Sr and Pb/Pb whole-rock isochrons on granulitic gneiss of Lavras-Oliveira-Passa Tempo suggest the high-grade metamorphism took place between 2.97 - 2.85 Ga. These rocks show slightly positive (+ 0.9 to + 0.2) and negative (- 0.6 to - 1.2) $\varepsilon_{Nd(2.97\,Ga)}$ values, in agreement with the interpretation of the SHRIMP data cited above.

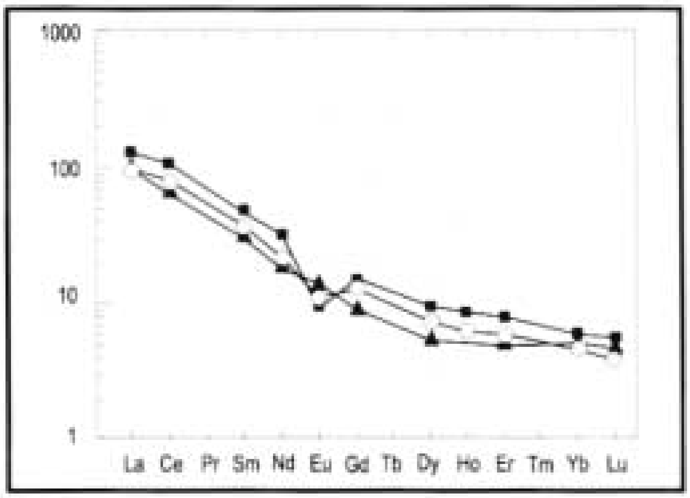
The younger zircon population in the Campo Belo migmatite yielded a SHRIMP U/Pb age of 2.839 ± 0.017 Ga considered to be the crystallization age of the neosome. This material has a $\varepsilon_{Nd(2.84\,Ga)}$ = - 0.8 indicating its derivation from a mantle source close to this age. The 2.84 Ga age compares well with that of the Belo Horizonte migmatitic gneiss (2.860 + 0.014 - 0.017 Ga), implying therefore a regional scale for the migmatization event in the southern part of the SFC. Plutonic activity post-dating the 2.84 Ga event is represented by less foliated granitoid plutons (Candeias, Santo Antônio do Monte, Itapecerica, see Fig. 12) that give Rb/Sr isochron ages between 2.75 and 2.70 Ga (Teixeira, 1993). These plutons are interpreted as being tectonically related to the Neoarchean granite intrusives of the QF (Teixeira *et al.*, 1998a).

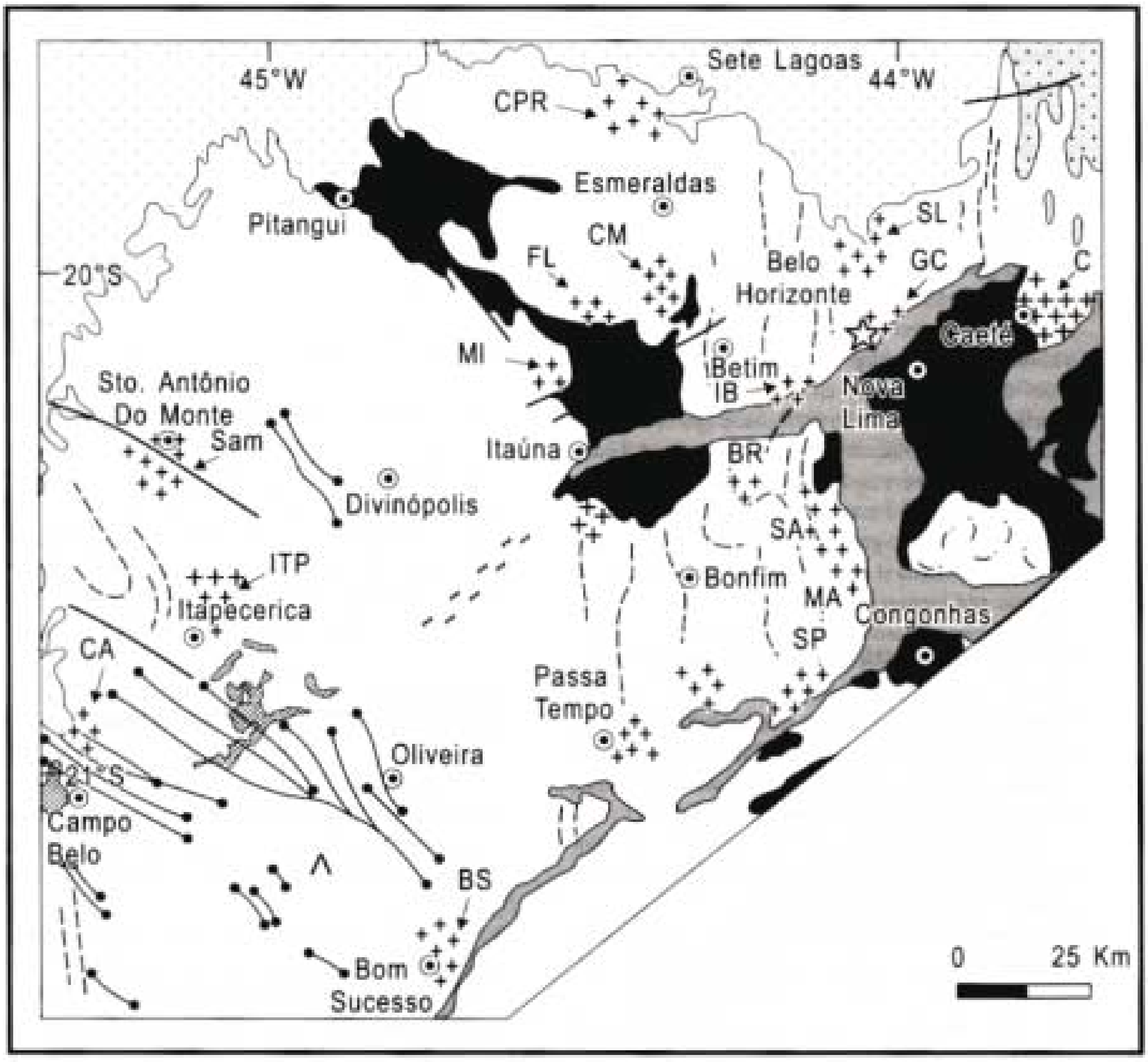
Greenstone Belts

Scattered relicts of low-grade supracrustal rocks (metapelite, schist, quartzite, and bif) and continuous greenstone belts occur within the Archean crust of the southern part of the SFC. Most of these occurrences have been correlated with the Rio das Velhas Supergroup of the QF. Some greenstone remnants (Fortaleza de Minas, Piumhi, Congonhas, and Rio Paraúna) are exposed within both the Paleoproterozoic and Neoproterozoic marginal belts (Fig. 6). The distribution of these greenstone relicts indicates the presence of Archean inliers outside the limits of the Neoproterozoic craton and suggests a former larger

Fig. 11 - Chondrite-normalized REE patterns (after Noce *et al.*, 1997; normalized values from Evensen *et al.*, 1978) for the Belo Horizonte gneiss (open circles) and the Alberto Flores gneiss (full squares); and high-Al (full triangles) Archean tonalite-trondhjemite suites (data from Condie, 1981).

FIGURE 12 - Distribution of Archean plutonism, mafic-ultramafic and mafic intrusions in southern part of the SFC. Symbols: cross-hatched pattern (the Ribeirão dos Motas layered sequence); cross (granitoid); dot-heavy lines (Archean mafic dykes). Other symbols as in Fig. 6. Keys: SAM = Santo Antônio do Monte. ITP = Itapecerica. BS = Bom Sucesso. CA = Candeias. CPR = Cachoeira da Prata. SL= Santa Luzia. GC = General Carneiro. C = Caeté. IB = Ibirité. CM= Caio Martins. FL = Florestal. ML = Mateus Leme. BR= Brumadinho. SA = Samambaia. MA = Alto Maranhão. SP = Santana do Paraopeba. See text for details.





extension of the Archean paleocontinent, further reworked during the Proterozoic evolution.

The Rio das Velhas Supergroup comprises the basal Nova Lima Group and the Maquiné Group (Door *et al.*, 1957). The Nova Lima Group comprises a volcano-sedimentary sequence hosting the main gold deposits of QF. The volcanic rocks include komatiite with spinifex texture (Schorscher *et al.*, 1982; Noce *et al.*, 1992) and tholeiitic basalt, intermediate to felsic volcaniclastic and volcanic rocks. Clastic and chemical metasedimentary rocks include pelitic and carbonaceous-pelitic rocks. Algoma-type iron formation units, greywacke, sandstone and conglomerate (Dorr, 1969; Ladeira, 1980). A significant part of the metasedimentary assemblage has been interpreted as a turbidite sequence (Schrank and Silva, 1993).

Two felsic volcanic rocks of the Nova Lima Group were dated by U/Pb in zircon at 2.772 ± 0.006 Ga and 2.776 + 0.023/ - 0.01 Ga (Machado *et al.*, 1992a, b). The felsic volcanism corresponds to the final stages of the predominantly ultramafic to mafic lower volcanic sequence of the Nova Lima Group, and the more precise age of 2.772 ± 0.006 Ga is the best constrained age for the greenstone belt. Detrital zircon dating for Nova Lima and Maquiné sedimentary rocks showed $^{207}Pb/^{206}Pb$ ages ranging from 2.877 ± 3.5 Ga (Machado *et al.*, 1996), suggesting that older continental crust was the main source area for the greenstone belt sediments. However, the Nova Lima volcanism was also a source for some of the overlying sediments, as indicated by the presence of detrital zircon dated at 2.777 - 2.771 Ga. Greenstone relicts (Piumhí, Fortaleza de Minas, Rio Paraúna) older than the Rio das Velhas Supergroup occur to the W of the QF (Fig. 6). The Piumhí greenstone belt (Schrank and Brousse, 1979) consists of a mainly volcanic sequence up to 3000 m thick. U/Pb dating on intrusive rocks into the lower sequence yielded: *a*) upper intersept discordia age of 3.116 + 0.01 / - 0.007 Ga (three abraded zircon crystals) on a gabbro anorthosite sill and, *b*) $^{207}Pb/^{206}Pb$ zircon ages between 3.0 and 2.965 Ga on a riodacite dome (Machado and Schrank, 1989). From the above the Piumhí greenstone belt probably originated during the main period of crust growth in the Campo Belo granite-gneissic terrane (3.1 - 2.9 Ga), as proposed by Teixeira *et al.* (1998a). The Fortaleza de Minas vulcano-sedimentary sequence, that crops out as discontinuous remnants within the Proterozoic marginal belt (Fig. 6) is probably contemporary with the Piumhí greenstone belt. Migmatite in thermal equilibrium with the surrounding Fortaleza de Minas rocks have a Rb/Sr isochron age of 2.918 ± 0.105 Ga (Schrank and Silva, 1993). Both the migmatite and the greenstone units were deformed, providing therefore indirect evidence for the age of the latter.

The Rio Paraúna greenstone belt occurs in the form of small tectonic sheets positioned within gneissic and granitoid rocks of the basement of the Espinhaço Supergroup (Fig. 6). It is mainly composed of mafic schist, ultramafic schist, sericite-quartz schist (in part corresponding to acid volcanics) and BIF units (Fogaça *et al.*, 1984). U/Pb zircon analysis on one acid volcanic rock yielded a crystallization age of 2.971 Ga (Machado *et al.*, 1989a), suggesting a tentative correlation with the Piumhí and Fortaleza de Minas greenstone belts.

Finally, to the S of the QF (near Barbacena, Fig. 6) a greenstone remnant (Baars, 1997) gave a U/Pb zircon upper intercept age of 3.13 ± 0.008 Ga (Söllner *et al.*, 1991). In addition, a Sm/Nd errorchron from the metakomatiitic rocks yielded *c.* 2.7 Ga, suggesting the time of metamorphism, whereas the Nd isotope inferences support their derivation from a depleted mantle source, as indicated by their slightly LREE-depleted pattern (Seixas and Baars, 1991).

Mafic-ultramafic and mafic intrusives

The Ribeirão dos Motas mafic-ultramafic layered sequence (RMS) is intrusive into the medium to high-grade gneiss in the southwestern part of the area (Fig. 12). It comprises mainly alternate igneous layers of peridotite and pyroxenite with euhedral crystals and cumulus textures. Subordinate gabbronorite and amphibolite are also present. The ultramafic rocks show MgO content ranging from 23.8 to 34.9 wt% (MgO = 30.1 average wt%); TiO_2 = 0.182 (average wt%); $CaO:Al_2O_3$ average ratio of 0.72, and $Al_2O_3:TiO_2$ average ratio of 33. In addition, they have Pt and Pd averages of 28 and 10.6 ppb, respectively (Carneiro *et al.*, 1999). Although the above geochemistry is broadly comparable with that of the Rio das Velhas komatiite (Schorscher, 1992), the RMS shows relatively higher MgO, Cr, Ni average contents and lower average contents of SiO_2, TiO_2, K_2O and REE. As a whole, the major and minor elements suggest that the RMS derived from komatiite magma similar to siliceous high magnesian basalt.

The RMS yields a Sm/Nd isochron age of 2.755 ± 0.062 Ga (whole-rock and minerals) with $\varepsilon_{Nd(t)}$ = + 1.1 (Carneiro *et al.*, 1997). The positive ε_{Nd} value implies that the RMS was emplaced in sialic crust soon after its mantle-differentiation at *c.* 2.75 Ga.

A NW-SE swarm of basic noritic-gabbroic dykes intrudes both the RMS and the country rocks. The dykes are in places overlain by small ferruginous quartzite remnants (M. A. Carneiro, written communication, 1999). The longest dykes are up to 30 km long and up to 100 m wide. Petrographically, the dykes are characterized by subophitic to intergranular textures, and the geochemistry reveals that they are predominantly andesi-basalt and subordinate tholeiitic basalt. A Sm/Nd (whole-rock and minerals) isochron age of 2.658 ± 0.044 Ga constrains the emplacement of the swarm (Pinese *et al.*, 1995).

Archean Crustal Evolution

U/Pb and Sm/Nd data older than 3.1 Ga are very uncommon in southern part of the SFC, and further work is therefore necessary to constrain the relevance of Paleoarchean and early Mesoarchean rocks in the crustal growth history (Fig. 13).

In the QF a few detrital zircon grains among 449 grains analyzed from the Archean Rio das Velhas and Paleoproterozoic Minas Supergroups dated by $^{207}Pb/^{206}Pb$ laser ablation-ICPMS yielded individual ages between 3.54 and 3.26 Ga (Machado *et al.*, 1996). The 3.54 Ga date is consistent with an inherited Pb age component (3.4 Ga) in zircon from banded gneiss in Barbacena (Söllner *et al.*, 1991). Moreover, a few Sm/Nd (T_{DM}) ages in the range 3.67 and 3.32 Ga are now available for the country rocks to the W of the QF (W. Teixeira, unpublished data), showing these to be the oldest protoliths within the studied area.

Table 3 summarizes the Archean crustal evolution of the southern part of the SFC, supported by the geochronological and isotope data. Inherited Pb age components in the Campo Belo Migmatite give an age of 3.2 Ga, which compares well with the T_{DM} ages available for the Archean granite-gneiss rocks in the area (Fig. 13). As stated before, the Campo Belo high-grade gneissic rocks formed between 3.1 and 2.9 Ga. This period of crustal growth can be also extrapolated to the QF, as suggested indirectly from the age bias of detrital zircon that increases after 3.0 Ga (Machado *et al.*, 1996), as well as by the bulk of the T_{DM} ages (Fig. 13).

The geochronological history is therefore somewhat compatible with the evolutionary scenario delineated for the Gavião Block (item 1), which assembled between 3.4 and 3.0 Ga. In particular, the oldest Pb inherited age component found in detrital zircon from the supracrustal rocks (3.54 Ga) permits a comparison of the crustal sources with the 3.40 Ga TTG suites in the northern part of the SFC, exposed as tectonic slices into the Contendas-Mirante Belt (Sete Voltas Massif).

Between 2.97 - 2.85 Ga, deformation, migmatization and magmatic events affected the entire pre-existing continental crust, which therefore explain the scattered isotopic evidence for primitive crust older than 3.0 Ga. Archean events younger than 2.85 Ga were otherwise periodic and short-lived. These could be related to progressive accretion of crustal blocks that reached final cratonization at *c.* 2.6 Ga. The best-defined event took place in the QF between 2.78 and 2.7 Ga, the Rio das Velhas Orogeny (Machado and Carneiro, 1992; Carneiro *et al.*, 1998b). This event comprises greenstone belt formation, granitoid intrusion and reworking of the older TTG crust (Fig. 13 and Table 3). The isotope evidence is compatible with either an island arc setting close to an old, evolved continental block or active convergent continental margin for the greenstone belt (Machado *et al.*, 1996; Carneiro *et al.*, 1998b). Older sialic crust clearly participated in the Rio das Velhas Orogeny, as shown by the ^{207}Pb / ^{206}Pb ages of detrital zircon from the greenstone belt sediments. Adjacent TTG suites were the main continental source for the basin (Machado *et al.*, 1996), as supported by the inherited U/Pb components (older than 2.92 Ga) in the Bonfim rocks and the Sm/Nd T_{DM} ages between 3.15 and 2.94 Ga (Carneiro *et al.*, 1998a, b).

The Rio das Velhas Orogeny, which started at 2.78 Ga, provided terrane assembly concomitantly with regional amphibolite facies metamorphism. Late-tectonic crust-derived granitoid intruded the crystalline basement at 2.72 - 2.7 Ga, and were the first rocks common to the coalesced continental crust (Teixeira *et al.*, 1996; Noce *et al.*, 1998; Carneiro *et al.*, 1998b). Some of the late-tectonic plutons were weakly deformed or not deformed at all during the Neoarchean, thus providing a minimum age for the last deformation of the related country rocks, as is the case of the Brumadinho Granite (2.703 + 0.024 / - 0.02 Ga) and the Mamona Granite (2.721 ± 0.003 Ga). These granite plutons show calc-alkaline and sub-alkaline affinities, respectively, and are associated to a transitional magmatic arc or intra-plate tectonic setting (Lacerda *et al.*, 1996; Carneiro *et al.*, 1998b).

Emplacement of the Ribeirão dos Motas mafic-ultramafic sequence and gabbro-noritic dykes took place between 2.75 and 2.66 Ga (Table 3). These Neoarchean activities result from major extension episodes, contemporary with late-tectonic granite emplacement and cratonization (Teixeira *et al.*, 1996; 1998a; Carneiro *et al.*, 1998b). Eventually, a regional, low-grade shearing episode overprinted the continental crust (Campo Belo, Lavras, and Itaúna) at *c.* 2.65 Ga (Teixeira *et al.*, 1996). In the QF the latest Archean episode is represented by intrusion of 2.6 - 2.55 Ga granitic plutons and veins.

Examples of such an intraplate regime operating at the end of the Archean are likewise seen in the northern part of the SFC (2.56 - 2.47 Ga), such as the formation of the Campo Formoso mafic-ultramafic layered complex (Oliveira and Knauer, 1993), the Rio Jacaré Sill, as well as the sub-alkaline and alkaline magmatisms (Pé de Serra). As a whole these igneous activities constrain the time of tectonic stability, crustal thickness and intracontinental rifting, that took place after the Neoarchean continental aggregation in the SFC. Late to post-orogenic mafic-ultramafic activities in Neoarchean times are also seen in many Archean cratons worldwide (Stillwater Complex in the Superior Province of Canada). The Yilgarn Craton of Western Australia and the Kaapvaal and Zimbabwe cratons of Southern Africa show also similar kinds of broadly contemporaneous igneous and metamorphic events at 2.8 - 2.6 / 2.5 Ga, which may reflect the formation of supercontinents, at this time (Myers, 1995).

Finally, different tectonic environments have been proposed for the formation of the Neoarchean greenstone belt and final cratonization. In the Superior Province, greenstone belts such as the Abitibi greenstone belt evolved largely in an oceanic environment (Corfu, 1993) and island-arc accretion played a major role in continental growth (Card, 1990). Conversely, older continental crust in QF and the southern part of the SFC was a major source for the Rio das Velhas greenstone belt sediments, and cratonization was probably accomplished by amalgamation of a number of distinct microcontinents.

The Paleoproterozoic terrane

The Transamazonian Orogeny (2.16 - 2.0 Ga) developed along the margins of the Archean continent in the southern part of the SFC (Teixeira and Figueiredo, 1991; Teixeira *et al.*, 1996). The tectonic scenario is illustrated by the evolution of the Mineiro mobile belt which formed granitoid and alkaline plutons, mafic dykes, as well as by the metamorphism of the Minas Supergroup and parts of the Archean crust (Noce *et al.*, 1997; Teixeira and Figueiredo, 1991; Figueiredo and Teixeira, 1996). The associated Transamazonian structures are NW-verging folds and thrusts affecting both the Rio das Velhas greenstone and the Paleoproterozoic Minas supracrustal rocks (Alkmim and Marshak, 1998). Tectonically associated Paleoproterozoic features include dome-and-keel

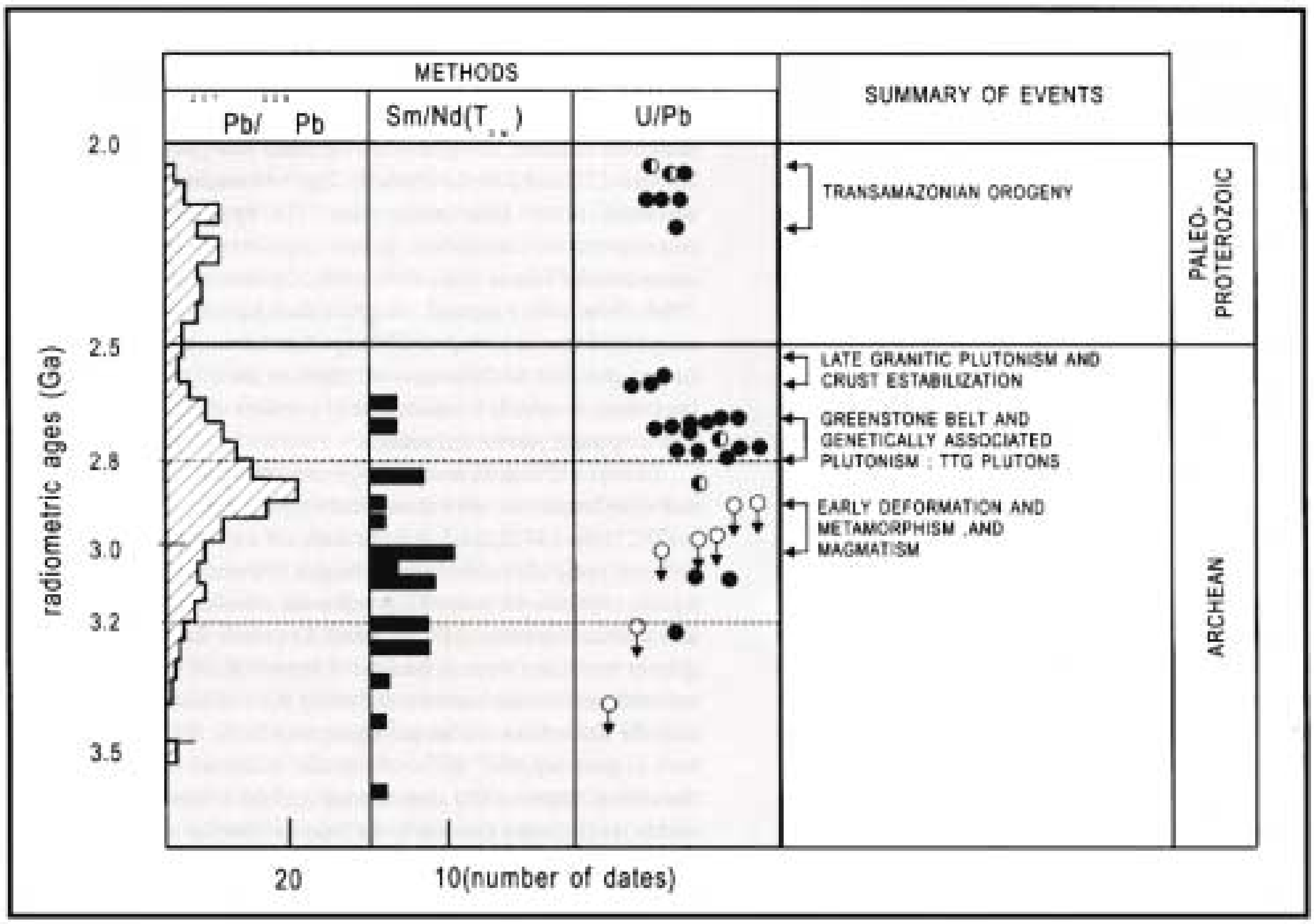


FIGURE 13 - Geochronological compilation and summary of the Archean crustal evolution in the southern part of the SFC. Emplacement ages (U/Pb dates) of Transamazonian intrusive granitoid plutons are also plotted (see text for details). Key symbols: 1) Hatched and full bars refer to the number of $^{207}Pb/^{206}Pb$ (detrital zircon) and SmNd (whole-rock) ages, respectively. 2) U/Pb dates: open dots with arrow are minimum ages of discordant analyses; half-open dots are metamorphic ages (titanite; monazite); full-dots are concordant zircon analyses and upper intercept ages.

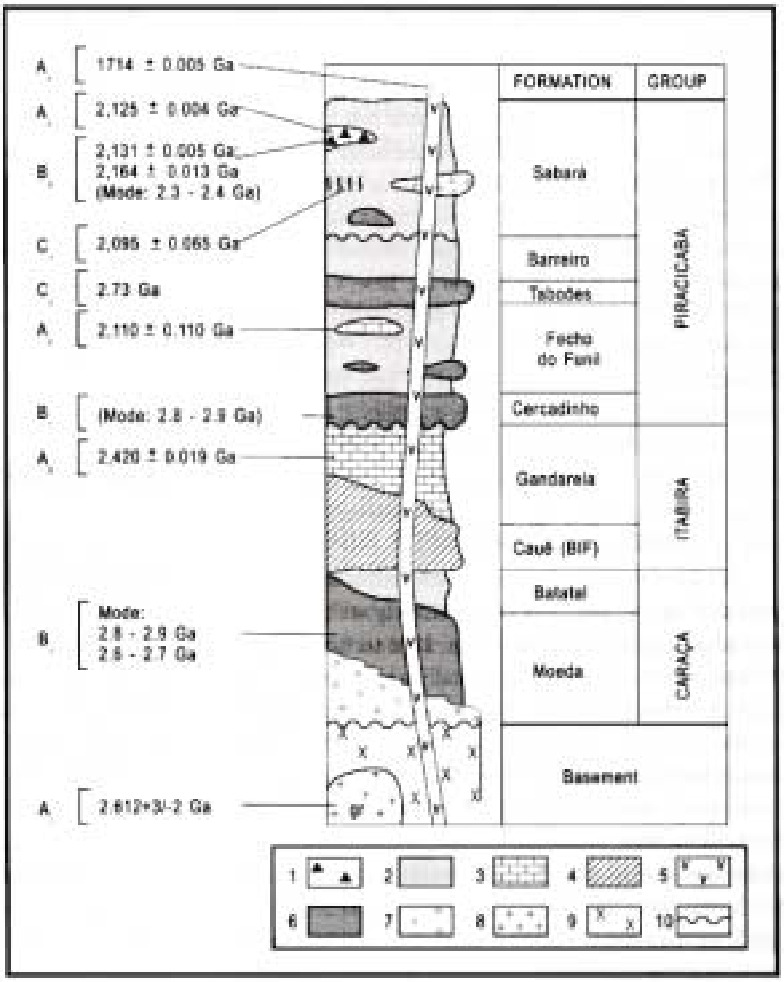


Fig. 14 - Chronostratigraphy of the Minas Supergroup (adapted from Babinski *et al.*, 1995; Alkmin and Marshak, 1998): 1= greywacke; 2= phyllite and schist; 3= carbonate rock; 4= banded iron formation units; 5= quartzite; 6= conglomerate; 7= crystalline basement. Keys: gr= Salto do Paraopeba Granite. (a) Pb/Pb whole-rock isochron age; (b) ^{207}Pb / ^{206}Pb laser ablation ICPMS age (detrital zircon); (c) U/Pb isotopic dilution concordant age (detrital zircon); (d) Sm/Nd whole-rock T_{DM} crustal residence age. See text for details.

structures in the QF, resulting from the orogenic collapse of the Transamazonian mobile belt.

U/Pb age determinations in titanite from the Belo Horizonte Gneiss, and titanite and monazite from amphibolite and pegmatite in the QF constrain the metamorphic peak at 2.065 - 2.035 Ga of the Transamazonian Orogeny (Machado *et al*,. 1992a, b). Comparable K/Ar amphibole ages in the range from 2.1 to 1.9 Ga probably resulted from progressive uplift of the Mineiro Belt, accompanying the tectonic stabilization and isotopic resetting. However, the majority of K/Ar biotite ages in the country rocks are younger and variable. They are interpreted as reflecting tectonothermal overprints, associated with the evolution of the Paleo to Mesoproterozoic Espinhaço Supergroup, as well as the Neoproterozoic marginal belts adjacent to the Craton (Teixeira, 1993).

Supracrustal rocks

The Minas Supergroup, a passive-margin sequence, constitutes the most important Paleoproterozoic sedimentary succession in the southern part of the SFC, mainly exposed in the QF, and discontinuosly as relicts to the SW (Serra de Bom Sucesso, Fig. 6). This Supergroup comprises the Caraça, Itabira and Piracicaba groups (Dorr, 1969). The contact between the first two is gradational, whereas an erosional unconformity marks the contact between the Itabira and the overlying Piracicaba Group (Fig 14). The basal Caraça Group consists mainly of fluvial sandstone and stable-shelf pelite, and its lower unit (Moeda Formation) is characterized by the presence of gold and uranium-bearing alluvial conglomerate. The Cauê Formation, at the base of the Itabira Group, consists mainly Lake Superior-type banded iron formation units, whereas the Gandarela Formation, at the top of the Group, consists of ferruginous dolomite and dolomite units.

Stromatolitic dolomite of the Gandarela Formation showed a Pb/Pb whole-rock isochron of 2.42 ± 0.019 Ga, interpreted to be the depositional age of the lower sequences of the Minas Supergroup (Babinski *et al.*, 1995). The Transvaal Supergroup in the Kaapvaal Craton (South Africa) and the Hamersley Group at the margin of the Pilbara Craton (Australia) are similar to the lower Minas Supergroup. These supracrustal rocks are broadly time equivalent as supported by U/Pb ages at 2.47 - 2.43 Ga (Cheney, 1996; Sumner and Bowring, 1996).

The Piracicaba Group consists of interbedded quartzite and phyllite containing carbonate lenses. Pb/Pb data from greenschist facies carbonate rocks from one of these lenses within the Fecho do Funil Formation indicated the isotopic systematic was reset at 2.1 Ga. This age is consistent with Rb/Sr and Pb/Pb whole-rock geochronology and U/Pb titanite and monazite ages on QF country rocks (2.124 - 2.0 Ga) which are consistent with the time of the Transamazonian metamorphism (Teixeira and Figueiredo, 1991; Machado *et al.*, 1992a, b; Machado *et al.*, 1996; Noce *et al.*, 1997).

The uppermost unit of the Minas Supergroup (Sabará Formation) is composed of chlorite schist, phyllite, greywacke, conglomerate, quartzite and rare iron formation units. Its maximum thickness, *c.* 3000 m, is close to the thickness of all the other units together, which reach 3500 m (Dorr, 1969). Detrital zircon from a Sabará greywacke gave two age modes at 2.9 - 2.8 Ga and 2.4 - 2.3 Ga (Fig. 14), which contrasts with the exclusively Archean $^{207}Pb/^{206}Pb$ age pattern of the underlying Minas formations (Machado *et al.*, 1996). However, the 2.73 Ga T_{DM} crustal residence age (Fig. 14) obtained for the same greywake suggests that the detritus was derived predominantly from the adjacent Neoarchean continent.

Deposition of Sabará flysch-like sediments marks a change both in depositional environment and source of sediments from those represented in the older units of the Minas Supergroup (Dorr, 1969). A contact aureole Sabará rock from a shear contact with the Archean basement displays a Sm/Nd garnet-muscovite-whole-rock age of 2.097 ± 0.065 Ga, constraining the time of regional metamorphism as likewise indicated by a U/Pb titanite age of the country rocks. In addition, the striking coincidence between the 2.125 ± 0.004 Ga age for the youngest zircon from a greywake of the Sabará Formation (Machado *et al.*, 1992a, b), and the 2.124 ± 0.001 Ga age of a tonalitic pluton occurring to the S of the QF (Noce *et al.*, 1998) indicates contemporaneity between flysch sedimentation and igneous activity during the Transamazonian Orogeny. The Itacolomi Group is separated from the Minas Supergroup by an angular unconformity (Guimarães, 1931). The group is exposed in restricted areas to the S and SW of the QF, and is composed predominantly of impure quartzite and conglomerate.

The youngest granite emplacement into the Archean crust (Salto do Paraopeba Granite; 2.612 Ga) constrains the maximum age of the Minas rocks. As both Minas and Itacolomi rocks were affected by the Transamazonian Orogeny developed at the margins of the Archean platform, their lower age limit is indicated by the 2.03 Ga monazite age from an undeformed pegmatite vein cutting basement gneiss (Machado *et al.*, 1992a, b). Subsequently, the Minas Supergroup was intruded by pegmatite veins and mafic dykes (Herz, 1970), one of which gives a U/Pb baddeleyite age of 1.714 ± 0.005 Ga (Silva *et al.*, 1995).

Finally, a Sm/Nd crustal residence of 2.2 Ga was recently reported for a metasedimentary rock the Dom Silvério sequence, a N-S trending tectonized metasedimentary belt (Fig. 6) exposed to the E of the QF (Brueckner *et al.*, 1998). This belt has been chronocorrelated with the Neoproterozoic Araçuaí supracrustal rocks on the basis of geological and tectonic inferences. The Sm/Nd age implies that the material was mostly derived from Transamazonian rather than from Archean crust. A similar paleotectonic scenario is suggested for the supracrustal assemblages of the Andrelândia Group that crop out to the W of Barbacena. A U/Pb age on the detrital zircon from this Group gave a U/Pb age of 1.872 ± 0.011 Ma, indicating the sedimentary provenance from a pre-existing Paleoproterozoic crust (Söllner and Trouw, 1997).

Plutonic activities related to the Transamazonian Orogeny

A string of granitoid plutons that extends nearly 300 km along the southern reworked border of the SFC constitutes a calc-alkaline plutonic arc developed in association with the evolution of the Mineiro Belt (Fig. 15). A number of plutons (Alto Maranhão, Ressaquinha, Tabuões, Cassiterita) are metaluminous to peraluminous tonalite (trondhjemite), diorite and granodiorite, with their chemical compositions indicative of a mafic-ultramafic (mantle) source (Ávila *et al.*, 1998; Noce *et al.*, 1999). They yield Paleoproterozoic T_{DM} ages (2.50 - 2.27 Ga), positive and slightly negative $\varepsilon_{Nd(T)}$ values (+ 1.3 to - 3.8) and low $^{87}Sr/^{86}Sr$ ratios (0.702 - 0.704; Noce *et al.*, 1999). The Cassiterita trondhjemite and Brumado quartz-diorite occurring in the vicinity of São João del Rei have $^{207}Pb/^{206}Pb$ evaporation zircon ages of 2.162 ± 0.01 Ga and 2.128 ± 0.004 Ga, respectively (Ávila *et al.*, 1998), whereas the Alto Maranhão Tonalite yields a U/Pb age of 2.124 ± 0.002 Ga (Noce *et al.* 1998) (Table 4).

A second group of plutons (Ritápolis, Itutinga, Alto Jacarandá, Porto Mendes, Lavras) consist mainly peraluminous granite that displays Archean T_{DM} ages (3.07 - 2.6 Ga), negative $\varepsilon_{Nd(T)}$ (- 4.9 to - 11.0) and high initial $^{87}Sr/^{86}Sr$ ratios (0.7096 - 0.7584). Therefore the isotope signatures indicate derivation from predominant crustal sources (Noce *et al.*,1999). The emplacement of the Ritápolis Granite (Fig. 15), a representative pluton of this second group, is well constrained by $^{207}Pb/^{206}Pb$ evaporation zircon ages of 2.122 ± 0.006 Ga and 2.121 ± 0.007 Ga (Ávila *et al.*, 1998). Northwards from the QF, a small and weakly-deformed granitic stock yielded a $^{207}Pb/^{206}Pb$ age of 2.04 Ga, interpreted to be the approximate age of the intrusion (Machado *et al.*, 1996).

From the above, the period of granitoid activity of the Mineiro Belt probably took place from 2.162 to 2.04 Ga (Fig. 13). Additional Rb/Sr and Pb/Pb whole-rock isochron ages on the Mineiro Belt granitoid plutons are between 2.1 and 1.9 Ga (Teixeira and Figueiredo, 1991; Table 4), but the youngest ages may be related to late isotopic disturbances and/or subordinate alteration episodes.

In summary, the origin of the Transamazonian granitoid plutons in the southern part of the SFC can be related to contrasting source-regions, from mantle to crustal-derived ones, and a number of them are probably derived from mixing of Paleoproterozoic juvenile material and variable proportions of Archean crustal material. The fact that Paleoproterozoic mantle-derived material played an important role in the generation of some of the granitoid plutons implies that a significant crustal accretion episode occurred in connection with the Transamazonian Orogeny. Therefore, processes of oceanic crust subduction and mafic magma underplating are likely to have occurred at the early evolutionary stages of the Mineiro Belt (Noce *et al.*, 1999). In this regard, tholeiitic dykes (*c.* 1.9 Ga) intruded into the Archean continental margin during the late extensional orogenic phase, tectonically conditioned to NW-SE, N-NE extension fractures and E-W anti-Riedel crustal fractures (Endo and Machado, 1998). Field relations indicate that these dykes are contemporaneous with emplacement of some of the granitoid plutons of the Mineiro Belt (Pinese *et al.*, 1995).

Archean-Paleoproterozoic basement of the Brasiliano marginal belt

The Paleoproterozoic domain containing scattered Archean inliers situated to the E of the QF was strongly reworked during of the Brasiliano Orogeny (Brito Neves and Cordani, 1991). A general feature of the country rocks is the increasing deformation and metamorphism combined with development of shear zones, folds and low-angle thrusts that caused tectonic transport of the geological units over the southern part of the SFC (Pedrosa Soares *et al.*, 1992; Figueiredo and Teixeira, 1996; Alkmim and Marshak, 1998).

Polyphase TTG gneiss and migmatite, and relicts of volcano-sedimentary assemblages comprise the main lithological units of the Paleoproterozoic domain, that crop out to the N, E and S of the QF (Guanhães, Gouveia, Ponte Nova, Barbacena, Lavras; figs. 6, 15), (Pedrosa Soares *et al.*, 1992; Figueiredo and Teixeira 1996; Noce *et al.*, 1997; Söllner and Trouw, 997). Elsewhere, most of the K/Ar mineral systems (country rocks and metasediments) were isotopically reset between 640 - 500 Ma, but Proterozoic dates older than 750 Ma have also been obtained (Teixeira, 1993). This heterogenous K/Ar cooling pattern is compatible with the polycyclic evolution and the complexity of the structural framework within the border of the SFC (Heilbron *et al.*, 1989; Figueiredo and Teixeira, 1996).

In Guanhães (Fig. 6) polycyclic medium to high grade gneiss-migmatite rocks and supracrustal relicts constitute the crystalline basement of the Neoproterozoic Araçuaí Belt. Structurally, these rocks display a N-S foliation and an E-W mineral stretching lineation (Marshak and Alkmin, 1989; Uhlein *et al.*, 1998). The U/Pb zircon and T_{DM} ages (3.02 - 2.97 Ga and 2.86 - 2.83 Ga) date the pre-existing Archean crust (Machado *et al.*, 1989b; Sato 1998). Rb/Sr dates indicate that further metamorphism and migmatization took place at 2.8 - 2.6 Ga, and later during the Transamazonian Orogeny (Brito Neves *et al.*, 1980). Eventually migmatization and pegmatite emplacement took place in the region between 600 - 512 Ma, in association with the Araçuaí Belt collision (Machado *et al.*, 1989b; Pedrosa Soares *et al.*, 1992; Dussin, 1994). Such a polycyclic history is likewise seen in the Itabira ore district, where a Paleoproterozoic banded-iron formation is tectonically imbricated with amphibolite. The latter rocks have been tentatively correlated with the mafic rocks of the Rio das Velhas Supergroup, as suggested by a Pb mineral isochron age of 2.66 Ga (Olivo *et al.*, 1996).

Between Ouro Preto and Abre Campo, in the E of the QF (Fig. 6), the banded gneiss are exposed as roughly parallel N-S belts, exhibiting subhorizontal foliation. The area is believed to be part of the Mantiqueira Province, a large arcuate Transamazonian belt containing gneissic rocks and migmatite of amphibolite to granulite facies, remobilized during the Brasiliano Orogeny (Alkmim and Marshak, 1998; Teixeira and Figueiredo, 1991). The Mantiqueira Gneiss yield Rb/Sr isochron dates between 2.3 - 2.0 Ga with high $^{87}Sr/^{86}Sr$ initial ratios (0.710), suggesting the partial reworking of Archean crust. Geochemically, the gneiss comprise two distinct suites ranging from basic to acid in composition, with low-K calc-alkaline and LILE-enriched high-K calc-alkaline affinities, respectively, which are comparable with

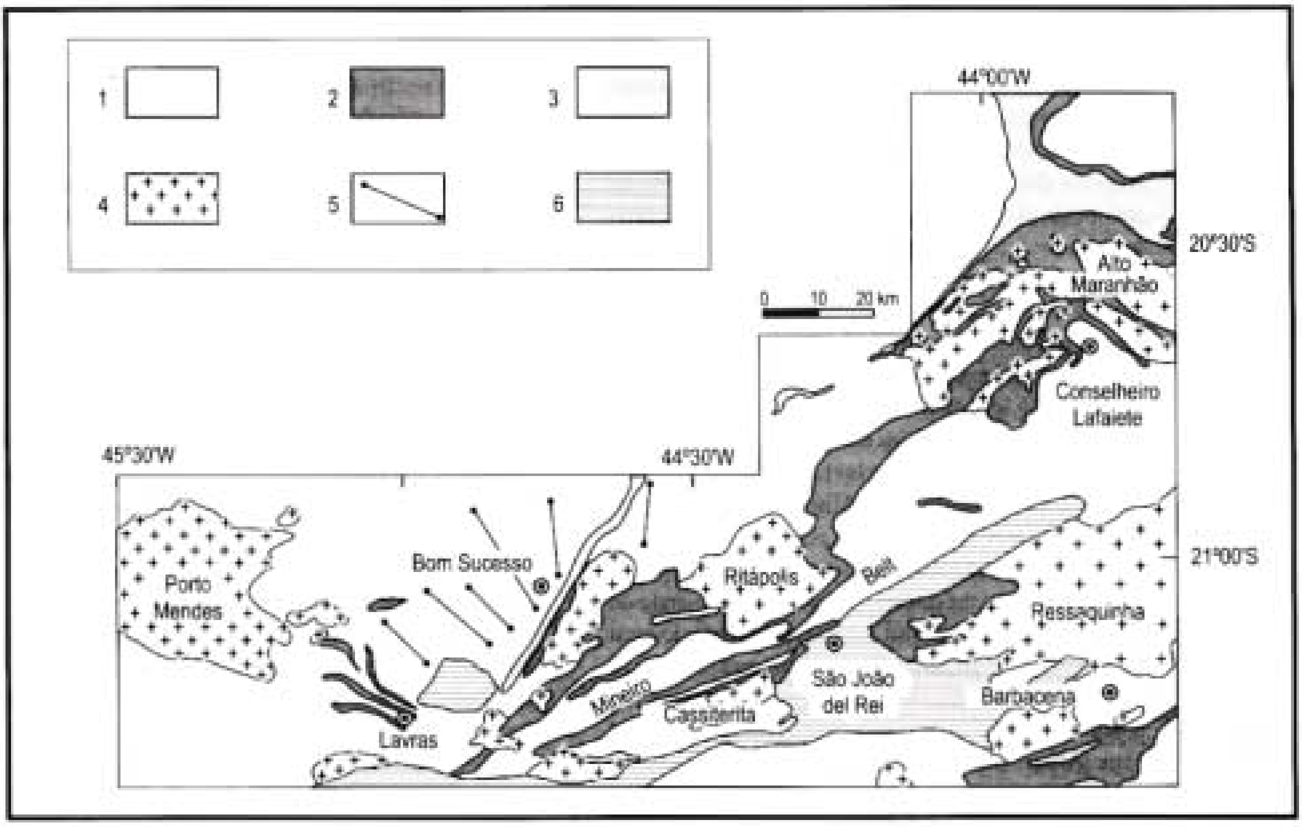


FIGURE 15 - The Mineiro mobile belt and its main plutonic bodies, related to the Transamazonian Orogeny. Archean: 1= TTG crust, partly reworked in the Proterozoic; 2= metasedimentary and metavolcanic rocks; Paleoproterozoic: 3= Minas Supergroup; 4= intrusive granitoid, diorite, gabbro; 5= mafic dykes; 6= Mesoproterozoic cover, reworked during the Brasiliano Orogeny.

subduction-related lithotypes in general (Figueiredo and Teixeira, 1996).

Between Acaiaca and Abre Campo granulite and kinzigite predominate, originating from reworking of the Archean crust (including a sedimentary protolith), as supported by their high $^{87}Sr/^{86}Sr$ initial ratio (0.706) and model μ_1 value of 8.6 (Teixeira *et al.*, 1997), and the large variations in the geochemical data and the field relations. In like manner, the Archean protholith is confirmed by T_{DM} ages (3.06 - 2.78 Ga) reported for a charnockitic gneiss in Ponte Nova situated to the E, whereas a Sm/Nd mineral isochron (2.2 Ga) on the same rock indicates the metamorphic overprint due to the Transamazonian Orogeny (Sato 1998; Buecker *et al.*, 1998). Additional Sm/Nd isotope mapping carried out on the Mantiqueira high-grade rocks between Rio Casca and Abre Campo again indicated the presence of an Archean protoliths in the range 3.26 - 3.02 Ga and 2.64 - 2.53 Ga (Fischel *et al.*, 1998). However a contrasting isotope signature is revealed within the southern extremity of the Mantiqueira Province toward the S of Barbacena and Ubá (Fig. 6), where the gneissic rocks disclose Archean (3.10 Ga) and Paleoproterozoic (2.22 - 2.19 Ga) T_{DM} ages (Sato, 1998).

From the considerations above the Mantiqueira Province represents part of the reworked margin of the Archean continent that had coalesced by the end of the Rio das Velhas Orogeny (item 2.1). However, juvenile accretion during the Paleoproterozoic also took place within the Mantiqueira Province, as suggested by isotopic and chemical data on the plutonic rocks. Such a polycyclic framework for the Transamazonian Belt is similarly seen at the southern end of the Craton where the gneiss of different ages and origins in conjunction with migmatite, supracrustal rocks and greenstone belt remnants form the polyphase, deformed basement of the Proterozoic São João del Rei, Carandaí and Andrelândia Proterozoic basins (Heilbron *et al.*, 1989; Söllner and Trouw, 1997; Ribeiro *et al.*, 1998; Noce *et al.*, 1997) (Fig. 15).

Paleoproterozoic crustal evolution

Cratonizaton attained after 2.6 Ga was succeeded by scarce magmatic-metamorphic activity during the Archean-Paleoproterozoic transition, which is reflected in the relatively few radiometric dates in the range 2.6 / 2.5 - 2.3 Ga (Teixeira and Figueiredo, 1991; Teixeira 1993; Machado et al., 1996). During such a long period of crustal stability erosion of pre-existing continental rocks and transport and deposition of the resulting material led to the deposition of the Minas Supergroup.

The lower Minas sequences (Caraça and Itabira groups) display an upward gradation from alluvial conglomerate and sandstone to shallow-water marine pelite, iron-formation units and carbonate, associated with the evolution of a passive margin basin (Teixeira and Figueiredo, 1991; Renger et al., 1994; Alkmin and Marshak, 1998; Noce et al., 1998, 1999). Carbonate units at the top of the Itabira Group were deposited c. 2.4 Ga ago (Fig. 14). The basal units of the Piracicaba Group which overlies the Itabira Group by an erosional unconformity were probably deposited in a basin that underwent slow subsidence, as indicated by the gradual disappearance of sandstone and increasing sedimentation of pelite (Machado et al., 1996), covered by flysh-like sediments of the Sabará Formation. According to these authors, the Piracicaba Group and the overlying Itacolomi Group may represent the filling of a foreland basin related to the Transamazonian orogenic stages. On the other hand, the Sabará Formation contains detrital zircon derived from Transamazonian magmatism, as young as 2.124 Ga. This fact supports the idea that the Minas Supergroup actually comprises two distinct basins separated by a large time-gap.

Alkmin and Marshak (1998) proposed a two-stage model for the evolution of the Mineiro Belt. First, a fold-thrust belt was created in response to NW-verging contraction, related to the accretion of an island arc and/or exotic terranes on the eastern and southeastern margins of the SFC. This episode must have envolved oceanic-crust consumption and generation of mantle-derived tonalite and trondhjemite dated at 2.162 - 2.124 Ga, followed by intrusion of syn to late collisional crust-derived granite, and foreland basin sedimentation, after *c.* 2.12 Ga (Table 4). However, the origin of the Mantiqueira plutonic gneiss probably involved a significant Archean component, as supported by their Nd and Sr signature (Figueiredo and Teixeira, 1996; Sato 1998).

The second stage of the evolution of the Mineiro Belt represents the orogenic collapse of the fold and thrust belt, resulting in the development of a dome-and-keel structure, which is the dominant structural feature of the QF. The rise of hot basement domes relative to supracrustal rocks (Minas Supergroup) resulted in metamorphic aureoles in the latter. A Sm/Nd garnet-muscovite-whole-rock isochron on the Sabará Schist indicates the metamorphism at 2.097 ± 0.065 Ga. Emplacement of late to post-tectonic granite plutons (Porto Mendes, Tabuões, Alto Jacarandá, Fig. 15) and U/Pb ages on titanite and monazite from Archean gneiss and pegmatite in the QF define the final extensional phase within the Mineiro Belt between 2.06 - 2.03 Ga. During this phase the reactivation of older NW-SE structures within the Archean foreland led to the appearance of tholeiitic dykes. Moreover, the K/Ar ages on biotite from the Archean country rocks (2.0 - 1.8 Ga) revealed the period of continent exhumation that accompanied tectonic stability of the Mineiro Belt (Teixeira *et al.*, 1997).

Such dynamics is again broadly comparable with the tectonic convergence between the Gavião and Jequié continents (northern part of the SFC), during the Paleoproterozoic. This scenario finally led to sedimentation in basins (Jacobina Group, Contendas Mirante upper formations), and the development of mobile belts, metamorphism and plutonic pulses associated with the Transamazonian Orogeny, in like manner to the Paleoproterozoic picture shown by the Minas Supergroup and the Mineiro Belt plutonism within the southern part of the SFC. Nevertheless, attempts at geometric and structural reconstruction between the northern and southern cratonic areas are strongly complicated by the lack of paleomagnetic data, by the further development of the Espinhaço rift-thrust belt along the axis of the Craton, and locally by the Neoproterozoic tectonic overprints within the Paramirim Province.

Finally, from the considerations above, the sedimentary, magmatic and tectonic records in the southern part of the SFC, between 2.5 and 2.0 Ga, seem to comprise the various stages of a Paleoproterozoic Wilson Cycle (Alkmim and Marshak, 1998), the orogenic and post-orogenic ones being represented by development of the Mineiro Belt. Nevertheless only a small remnant of this belt is preserved, contrasting with the larger areas strongly reworked during the Brasiliano Orogeny. Thus, many aspects of the evolution of the Mineiro Belt must be better understood with the on-going mapping, geochemistry and isotope data. In the future, this effort will lend support to more reliable tectonic correlation between the geological units in the SFC.

Table 1 – Main tectonic episodes from 3.4 to 1.9 Ga, in the northern part of the SFC. An asymmetric mantle convective system controls the growth process model, which included in the Paleoproterozoic the formation of a major suture zone (Contendas-Jacobina Lineament).

Age (Ga)	*Western domain*	*Collision domain*	*Eastern domain*
Archean Episodes			
3.7-3.4	Oceanic tholeiite, first sources of TTG suites (Sm/Nd T_{DM} model ages)	Oceanic crust ?	?
3.4-3.1	Formation of the Gavião continent (TTG accretion)	Oceanic crust ?	First sources of granitic rocks (Sm/Nd T_{DM} model ages)
3.1-3.0	Main greenstone belts Lower unit of the Contendas-Mirante belt	Oceanic crust ?	
2.9-2.6	Local magmatism; felsic volcanics (Umburanas greenstone belt); calc-alkaline plutonism	Oceanic crust ?	Formation of the Jequié continent; (Granite accretion leading to charnockitic terrains)
Early Paleoproterozoic Episodes			
Age (Ga)	*Western domain*	*Collision domain*	*Eastern domain*
2.5 ?	Progressive downfall of dense oceanic crust		
2.5-2.4	Eastward transport (Gavião continent); calcalkaline volcanism; Jacaré sill emplacement (extensive regime ?)	Downfall of dense oceanic crust. Beginning of converging dynamics	Westward movement of the Jequié continent. Extensive mechanisms. Pé de Serra alkali granite emplacement
2.3		Beginning of obduction	Transpressive regime Shortening of the Pé de Serra granite; synkinematic emplacement of alkali-granite
Pre-collisional episodes			
2.3-2.1	Extensive regime =	Obduction Development of foreland basins (Jacobina, Contendas) <u>Lower units deposits</u> Mafic magmatism; marine, chemical, clastic and volcaniclastic deposits	Transpressive regime Calc alkaline magmatism of the Salvador-Curaçá and Itabuna-Atlantic coast belts D_1 tectonics Early granitic plutonism (Serrinha block)
Collisional episodes			
2.1-2.0	Alkali magmatism of Guanambi-Urandi batholith. Horizontal tectonics Formation of tectonic slices =	Beginning of collision Evolution of foreland basins (Jacobina, Contendas) Compressive to transpressive D_1 tectonics within the lower units West vergence transports Intercalation of slices into the belt <u>Upper units deposits</u> Continental fluvio-deltaic deposits	Transpressive regime Granulitic metamorphism Uplift of the Jequié block Late granitic intrusions Alkali magmatism of the Salvador-Curaçá belt Post-kinematic granitic plutonism in the Serrinha Block
2.0-1.9	E-W shortening Anatexis and emplacement of late-granite	Emplacement of early peraluminous granite Compressive to transpressive D_2 tectonics within both the lower and upper units / E-W shortening Emplacement of late peraluminous granite	Transcurrent regime Post-kinematic alkali-magmatism of Salvador-Curaçá belt

Table 2 - Summary of the Paleoproterozoic evolution of the West Central African Belt (CGC), as compared with the northern part of the SFC (compiled from Feybesse et al., 1998).

West Central African Belt		**Similarities with the Northern SFC**	
Main steps	*Age (Ga)*	*Age (Ga)*	*Main tectonic steps*
Archean Craton **break up** without ocean spreading or juvenile accretion	?	2.7	Existence of two distinct Early and Neoarchean paleocontinents
Emplacement of **plutonic rocks**	2.52-2.44	2.56-2.47	**extensive regime:** Pé de Serra granite; calc-alkaline volcanics; Jacaré sill; Itabuna-Atlantic coast magmatism
Formation of **sedimentary basins** (Ogooué ?, Nyong, Ikobé-Waka)	?		
shortening event (E1) : nappe tectonism (unknown vergence); basinal deposits metamorphosed under 7-12 kbar P conditions	2.30-2.23	2.30	**shortening regime** (Jequié Block only) Pé de Serra granite; arc-subduction related magmatism in the Itabuna-Atlantic coast belt
crustal thinning and new break up of a nearby Archean Craton, and opening of pre-oceanic extentional basins (Francevillian lower sediments)	2.15	2.30 2.22-2.17	**Development of foreland basins** Jacobina and Contendas-Mirante lower units; beginning of collision tectonics
second tectonic event (E2) Thrusts and metamorphic peak at 9kbar (810°C) in the belt inner zone	2.12	2.13	Syn-kinematic plutonism in the Serrinha Block
Brittle tectonism in the foreland with accelerated subsidence of the Francevillian basins (upper sediments = flysch deposits)	2.10	2.10	Late-kinematic plutonism in the Serrinha Block Plutonism in the Salvador-Curaçá belt
Thrust transport of belt inner zone onto the outer zones (eastward vergence). Metamorphism peak (8 kbar, 600°C) and anatexis		2.08	Jacobina and Contendas-Mirante upper sediments. Salvador-Curaçá belt plutonism. Granulite metamorphism (Jequié, Salvador-Curaçá, Itabuna)
Transport of the pile onto the autochtonous Archean Chaillu Block (eastward general vergence)	2.05-2.04	2.06	Guanambi Batholith emplacement. Late granitic plutonism in the Serrinha Block
Terrigenous molasse deposits in Francevillian basins			
third tectonic event (E3) Generation of folds and strike-slip faults Late orogenic granites	2.04 2.04-1.92	2.00-1.98 1.94-1.90	Last intrusions (Guanambi Batholith); Jacobina Belt metamorphism and peraluminous granite. Successive thrust tectonics (westward vergence)and related granites. Au, Mo, Be concentrations (Jacobina basin)

Table 3 – Summary of the Archean tectonomagmatic events in the southern part of the SFC in the light of U/Pb and Sm/Nd data.

Age (Ma)	Characteristics of the crustal evolution in Southern SFC
3,205 ± 25	[1] - Emergence of early continental crust, compatible with T_{DM} ages on the country rocks, up to 3.25 Ga. Formation of the Piumhi greenstone belt and correlatives.
3,047 ± 25	[1] - Major generation of Campo Belo crust and progressive magmatic accretion. The gneisses have Pb isotopic signature consistent with mantle-like single-stage evolution, yielding comparable T_{DM} ages between 3.00 and 2.90 Ga. Origin of the Belo Horizonte gneiss protholiths [3], as well as part of the Bonfim gneiss protholiths. Development of granulitic facies metamorphism.
2,860 +/- 10 to 2,839 ± 17	[1] - Crystallization of neossome material in migmatites, as well as TTG crust reworking [2,3].
2,778 - 2,698	Formation of the Bonfim TTG suite and the associated Rio das Velhas greenstone belt (including Congonhas and Caeté) - the Rio das Velhas orogeny [2]. Crustal reworking and emplacement of mafic-ultramafic bodies and gabbro-noritic dikes [1]. Granitoid intrusions in the entire terrain [1,2] and regional migmatization [2]. Progressive terrain assembly starting at 2, 780 Ma ago.
2,612 - 2,593	Late-tectonic granitoid intrusions after amalgma and final crust stabilization [1,2].

Data compiled from Carneiro *et al.* 1998a; Teixeira *et al.* 1996, 1998a; Noce *et al.* 1998). See text and Fig. 6 for details.
*Obs.: [1] = Campo Belo gneissic-migmatitic terrain; [2] = Bonfim dome; [3] = Belo Horizonte dome.

*Table 4: Main geochronology and isotope characteristics for Paleoproterozoic granitoid plutons of the Mineiro Belt (see text for details and Fig. 15). *ε_{Nd} values are calculated for the ages indicated in parenthesis.*

Granitoid	Age (Ga); [method]; (MSWD)	$^{87}Sr/^{86}Sr$ initial	ε_{Nd} [T (Ga)*]	T_{DM} (Ga)	Characteristics
Ritápolis ("R")	2121±0.007 [Pb/Pb zircon] 1863±0.044 [Rb/Sr]; (18.3)	0.7584±0.0087	-4.9 (2.1) -5.8 (2.1)	2.62 2.71	Weakly-foliated; highly differen0.0tiated peraluminous pluton of calc-alkaline affinity, bearing xenoliths of "C" and "B"
Cassiterita ("C")	2162±0.010 [Pb/Pb zircon]	——	——	—	Peraluminous trondhjemitic to tonalitic pluton. Calc-alkaline affinity; strongly REE fractionated with weak Eu anomaly
Brumado ("B")	2.128±0.004 Ma	——	——	—	Metaluminous dioritic to tonalitic pluton. Calc-alkaline affinity; moderate REE fractionated with weak Eu anomaly
Itutinga	~ 1.90 [Rb/Sr; inferred from "R"]	~ 0.710	-7.7 (1.90)	2.77	Peraluminous granite. LREE enriched, with significant Eu negative anomaly
Alto Jacarandá	1.900±0.108 [Rb/Sr]; (3.95)	0.7096±0.0018	-7.6 (1.90) -10.8 (1.90)	2.85 2.95	Strong foliated, peraluminous granite
Lavras	1.940±0.100 [Rb/Sr]; (0.70) 1.982±0.134 [Rb/Sr]; (0.61)	0.70417±0.00107 0.7041±0.0017	-7.2 (1.94) -3.8 (1.98)	2.62 2.46	Metaluminous to peraluminous pluton (granite to tonalite)
Porto Mendes	2.061±0.082 [Rb/Sr]; (6.75) 1.855+0.176-0.200 [Pb/Pb]	0.70405±0.00362 μ_1 = 8.103	-4.9 (2.06) -11.0 (2.06) -10.3 (2.06) -9.3 (2.06)	2.62 3.03 3.07* 3.01*	Predominantly undeformed granite, weekly peraluminous
Campolide	1.998±0.017 [Rb/Sr]; (1.91)	0.7157±0.0018	——	—	Porphyritic granodiorite
Alto Maranhão	2.124±0.002 [U/Pb]	0.70739±0.00030 0.70788±0.00012	+1.3 (2.12)	2.27	Well-foliated metaluminous tonalite. REE fractionated with HREE depletion and no Eu anomaly
Ressaquinha	2.010±0.052 [Rb/Sr]; (9.43)	0.7086±0.0006	-1.3 (2.01)	2.30	Metaluminous to peraluminous tonalite, rich in xenoliths of gneiss and granulite
Tabuões	1.962±0.020 [Rb/Sr]; (5.75)	0.70245±0.00005	-2.8 (1.96)	2.43	Metaluminous to peraluminous trondhjemite

Table 5 - Age distribution of plutons

Sete Voltas		Sete Voltas	Jequié	Pé de Serra	Salvador-Curaçá	Itiuba	Paramirim
Early Archean Sources	Boa Vista Old TTG	Aracatu Young TTG	charnockitic plutonism	alkaline Itabuna Atlantic coast shoshonitic	TTG to calcialkaline	syenite Contendas Jacobina peraluminous	Meso-Proterozoic plutonism
3.7Ga	3.4Ga	3.2-3.1Ga	2.9-2.6Ga	2.4Ga	2.1Ga	1.9Ga	1.7Ga

300	200-300	300-500	200-300	300	200	100

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REFERENCES

Alibert, C. and Barbosa, J.S.F. (1992). Âges U-Pb déterminés à la 'SHRIMP' sur des zircons du complexe de Jequié, Craton de São Francisco, Bahia, Brésil. In: *Soc. Géol. France, 14° Réun. Sci. Terre, Toulouse, Abstr.*, 4.

Alkmin, F.F., Brito Neves, B.B. and Castro Alves, J.A. (1993). Arcabouço tectônico do Cráton do São Francisco. Uma revisão. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L., Misi, A., pp. 45-62. Soc. Bras. Geologia, núcleo Bahia/ Sergipe, Salvador.

Alkmin, F.F. and Marshak, S. (1998). Transamazonian Orogeny in the Southern São Francisco Craton Region, Minas Gerais, Brazil: evidence for Paleoproterozoic collision and collapse in the Quadrilátero Ferrífero. *Precambrian Research*, 90, 29-58.

Almeida, F.F.M. (1977). O Cráton São Francisco. *Revista Brasileira Geociências*, 7, 349-364.

Ávila, C.A., Valença, J.G., Moura, C., Ribeiro, A. and Paciullo, F. (1998). Idades ^{207}Pb-^{206}Pb em zircões de corpos metaplutônicos da região de São João Del Rei, borda Sul do Cráton do São Francisco. Minas Gerais. In: *Brazilian Geol. Congr. 35×, Belo Horizonte, Anais, SBG*, 34.

Baars, F.J. (1997). The São Francisco Craton. In: *Greenstone Belts* (De Witt, M and Ashwal, L. D.). *Oxford Monographs on Geology and Geophysics*, 35, 529 - 557.

Babinski, M., Chemale, F. and Van Schmus, W.R. (1995). The Pb/Pb age of the Minas Supergroup carbonate rocks, Quadrilátero Ferrífero, Brazil. *Precambrian Research*, 72, 235-245.

Barbosa, J.S.F. (1986). *Constitution lithologique et métamorphique de la région granulitique du sud de Bahia, Brésil.* Thèse doct. Univ.Paris VI, 86-34, 300 p.

Barbosa, J.S.F. (1990). The granulites of the Jequié complex and Atlantic mobile belt, southern Bahia, Brazil - An expression of Archean-Proterozoic plate convergence. In: *Granulites and crustal evolution*, eds. D. Vielzeuf and Ph. Vidal Kluwer Academic Publishers, 195-221.

Barbosa, J.S.F., Alves da Silva, F., Sapucaia, N.S. and Iyer, S.S. (1998). Petrology and geochemical characteristics of the granulitic terrain of Brejões, Bahia, Brazil. *Gondwana Research*, 1(3/4), 343-356.

Barbosa, J.S.F., Marinho, M.M.M. and Sabaté, P. (1994). Geology of the Jequié-Itabuna granulitic belt. In: *Petrologic and Geochronologic evolution of the oldest segments of the São Francisco craton, Brazil*, eds. M.C.H. Figueiredo and A.J. Pedreira. *Bol. IG-USP*, 17, 9-36.

Barbosa, J.S.F and Dominguez, J.M.L., coord. (1996). *Texto Explicativo para o Mapa Geológico ao Milionésimo da Bahia.* Publ. Especial, SICM/SGM, Salvador-Ba, 400 p.

Barbosa, J.S.F and Fonteilles, M. (1989). Caracterização dos protólitos da região granulítica do sul da Bahia. *Revista Brasileira de Geociências*, 19, 3-16.

Barreto dos Santos, E., Rosa, M.L.S., Conceição, H., Scheller, T., Macambira, M.J.B., Meier, M., Martin, H. Paim, M.M., Pereira Santos, C.G. and Leahy, G.A.S. (1999). Geochronology Rb/Sr, Sm/Nd, $^{207}Pb/^{206}Pb$ and U/Pb in the syenites and monzonites of Estreito massif (SW of Bahia state, NE Brazil). In: *II South Amer. Symp. on Isotope Geology, Córdoba, Argentina*, Actas, 11-14.

Bastos Leal, L.R.B., Teixeira, W, Macambira, M.J.B., Cordani, U. and Cunha, J.C. (1996). Evolução crustal dos terrenos TTGs arqueanos do bloco do Gavião, cráton de São Francisco, Geocronologia U-Pb (Shrimp) e Pb-Pb em zircões. In: *Brazilian Geol. Congr.* 39° Salvador, Anais, SBG., 6, 539-541.

Bastos Leal, L.R.B. Teixeira, W, Cunha, J.C. and Macambira, M.J.B. (1997). Crustal evolution of the Gavião block of the São Francisco craton: A geochronological study with U-Pb, Pb-Pb, Sm-Nd, Rb-Sr and K-Ar. In: *South Amer. Symp. Isotope Geology, Campos do Jordão, Brazil, Extended Abstr.*, 161-162.

Bordini, R.M., Cruz, M.M.J. and Sabaté, P. (1999). O Maciço de Potiraguá no lineamento de intrusões gabro-anortosíticas do cinturão granulítico do Sul da Bahia (Brasil), Especificidades Geológicas e Estruturais. In: *Anais VII Simp. Nac. Estudos Tectônicos, Lençóis, Brazil*, 48-51.

Brito, R.S.C. (1984). Geologia do sill estratificado do Rio Jacaré. *33° Congr. Bras. Geol., Rio de Janeiro, Anais, Soc. Bras. Geociências*, 9, 4316-4334.

Brito Neves, B.B., Cordani, U.G. and Torquato, J.R. (1980). Evolução geocronológica do Pre-cambriano do Estado da Bahia. *Geologia e Recursos Minerais do Estado da Bahia, Textos Básicos*, SGM/CPM, 1, 101 p.

Brito Neves, B.B. and Cordani, U.G. (1991). Tectonic evolution of South America during the Late Proterozoic. *Precambrian Research*, 53, 23-40.

Bueckner, H., Cunningham, W.D., Alkmin, F.F. and Marshak, S. (1998). Implications of new Precambrian Sm-Nd dates from the Quadrilátero Ferrífero and adjacent northern Mantiqueira province. In: *Inter. Conf. Basement Tectonics, 14× Ouro Preto, MG. Extended Abstr.*, 24-25.

Brunet, D. and Machetel, P. (1998). Large scale tectonic features induced by mantle avalanches with phase, temperature and pressure lateral variations of viscosity. *Journal of Geophysical Research*, 103, 4929-4945.

Card, K.D. (1990). A review of the Superior Province of the Canadian Shield, a product of Archean accretion. *Precambrian Research*, 48, 99-156.

Carneiro, M.A., Teixeira, W., Carvalho Jr., I.M., Oliveira, A.H. and Fernandes, R.A. (1997). Archean Sm/Nd isochron age from the Ribeirão dos Motas Layered rocks sequence, southern São Francisco Craton, Brazil. In: *1st South-American Symp. Isotope Geology, Extended Abstr., Campos do Jordão, Brazil*, 63-64.

Carneiro, M.A., Carvalho Junior, I.M. and Teixeira, W. (1998a). Petrologia, geoquímica e geocronologia dos diques máficos do Complexo Bonfim Setentrional (Quadrilátero Ferrífero) e suas implicações na evolução crustal do Cráton do São Francisco. *Revista Brasileira de Geociências*, 28, 95-102.

Carneiro, M.A., Teixeira, W., Carvalho Junior, I.M. and Fernandes, R.A. (1998b). Ensialic tectonic setting of the Archean Rio das

Velhas Greenstone belt: Nd and Pb isotopic evidence from the Bonfim Metamorphic Complex, Quadrilátero Ferrífero, Brazil. *Revista Brasileira de Geociências*, **28** (2), 189-200.

Carneiro, M.A., Grossi Sad, J.H., Carvalho Júnior, I.M., Nalini Jr., H.A., Lima, J.F., Camporez, J.A.P. and Teixeira, W. (1999). Possible Komatiitic Affinity and Potencial PGE Mineralization of the Archean Ribeirão dos Motas Layered Sequence, Brazil. In: *Third International Workshop on Orogenic Lherzolites and Mantle Processes. Pavia, Italy, Abstr.*, **24**, 80-81.

Chauvet, A., Alves da Silva, F.C., Faure, M. and Guerrot, C. (1997). Structural evolution of the Paleoproterozoic Rio Itapicuru granite-greestone belt (Bahia, Brazil): the role of synkinematic plutons in regional tectonics. *Precambrian Research*, **84**, 139-162.

Cheilletz, A., Ferraud, G. Giuliani, G. and Ruffet, G. (1993). Emerald dating through ^{40}Ar/^{39}Ar step-heating and laser spot analysis of syngenetic phlogopite. *Earth and Planetary Science Letters*, **120**, 473-485.

Cheney, E.S. (1996). Sequence stratigraphy and plate tectonics significance of the Transvaal succession of southern Africa and its equivalent in Western Australia. *Precambrian Research*, **79**, 3-24.

Choukroune, P., Bouhallier, H. and Arnt, N.T. (1995). Soft lithosphere during periods of Archaean crustal growth or crustal reworking. In: Early Precambrian Processes, eds. Coward, M.P. and Ries, A.C. *Geological Society Special Publ.*, **95**, 67-86.

Conceição, H., Sabaté, P., Alonso, M.D. and Bonin, B. (1989). Mise en évidence d'un contrôle de la mise en place du massif syénitique d'âge Protérozoïque inférieur d'Itiuba (Bahia, Brésil). *Comptes Rendus de l'Académie des Sciences, Paris*,, **309**, 403-408.

Conceição, R.V. (1998). Petrologia dos sienitos potássicos do maciço de Santanápolis e alguns aspectos do seu embasamento granulítico. In: Sienitos alcalino-potássicos e ultrapotássicos paleoproterozóicos do estado da Bahia, (eds) Conceição, H. and M.J.M., Cruz. *SBG, Núcleo Ba-Se, Publ. Especial*, **4**, 40-81.

Conceição, R.V., Rosa, M.L.S., Nardi, L.V., Conceição, H., Lafon, J.M., Soliani, E., Oberli, F. Meier, M. and Martin, H. (1999). Geochronology and isotopic signature of the Paleoproterozoic Santanápolis syenite (Bahia, Brazil). In: *II South Amer. Symp. on Isotope Geology. Córdoba, Argentina, Actas*, 178-171.

Condie, K.C. (1981). *Archean greenstone belts*. Elsevier, 434 p.

Condie, K.C. (1993). Chemical composition and evolution of the upper continental crust: contrasting results from surface samples and shales. *Chemical Geology*, **104**, 1-37.

Cordani, U.G., Kawashita, K., Sato, K., Iyer, S.S. and Taylor, P.N. (1992). Pb-Pb, Rb-Sr and K-Ar systematics of the Lagoa Real Uranium Province (South Central Bahia, Brazil) and the Espinhaço cycle (ca. 1.5-1.0 Ga). *Journal of South American Earth Sciences*, **5**(1), 33-36.

Cordani U.G., Sato, K. and Marinho, M.M. (1985). The geological evolution of the ancient granite-greenstone terrain of central-southern Bahia, Brazil. *Precambrian Research*, **27**, 187-213.

Cordani, U.G., Sato, K. and Nutman, A. (1999). Single zircon SHRIMP determination from Archean tonalitic rocks near Uauá, Brazil. In: *II South American Symp. on Isotope Geology. Córdoba, Argentina, Actas*, 27-30.

Corfu, F. (1993). The evolution of the southern Abitibi greenstone belt in light of precise U-Pb geochronology. *Economic Geology*, **88**, 1323-1340.

Cox, D.P. (1967). Regional environment of the Jacobina auriferous conglomerate, Brazil. *Economic Geology*, **62**, 773-780.

Cruz Filho, Rios, D.C., Conceição, H., Scheller, T., Macambira, M.J.B., Oliveira, A.E.L. and Burgos, C.M.G. (1999). Petrography and ^{207}Pb/^{206}Pb age of the Nordestina Batholith, Northeastern of the state of Bahia. In: *II South Amer. Symp. on Isotope Geology, Córdoba, Argentina, Actas*, 36-39.

Cuney, M., Sabaté, P., Vidal, Ph., Marinho, M.M. and Conceição, H. (1990). The 2 Ga peraluminous magmatism of Jacobina-Contendas Mirante Belt (Bahia) Brazil : Major and trace element geochemistry and metallogenetic potential. *Journal of Volcanology and Geothermal Research*, **44**, 123-141.

Cunha, J.C. and Fróes, R.J.B. (1994). Komatiitos com textura spinifex do greenstone belt de Umburanas, Bahia. *Série Arquivos Abertos, CBPM, Salvador*, 29 p.

Cunha, J.C., Lopes, G.A.C. and Sabaté, P. (1994). Estrutura do bloco do Gavião no cráton do São Francisco (Bahia, Brasil), exemplo de tectogênese diacrônica Proterozóico Inferior a Superior de um segmento continental arqueano. In *Brazilian Geol. Congr., 38° Camboriú - SC, Anais, SBG*, **5**, 381-382.

Cunha, J.C., Bastos Leal, L.R., Fróes, R.J.B., Teixeira, W. and Macambira, M.J.B. (1996). Idades dos greenstone belts e dos terrenos TTG's associados da região de Brumado, centro-oeste do cráton São Francisco (Bahia, Brasil). In: *Brazilian Geol. Congr. 39°, Salvador, Anais, SBG*, **1**, 67-70.

Davison, I., Teixeira, J.B.G., Silva, M.G., Rocha Neto, M.B. and Matos, F.M.V., (1988). The Itapicuru greenstone belt, Bahia, Brazil: Structure and stratigraphical outline. *Precambrian Research*, **42**, 1-17.

Delhal, J. and Demaiffe, D. (1985). U-Pb Archaean geochronology of the São Francisco Craton (Eastern Brazil). *Revista Brasileira de Geociências*, 15, 55-60.

Dorr II, J.V.N., Gair, J.E., Pomerene, J.G. and Rynearson, G.A. (1957). Revisão da estratigrafia pré-cambriana do Quadrilátero Ferrífero. *DNPM/DFPM*, **81**, 1-31.

Dorr II, J.V.N. (1969). Physiographic, stratigraphic and structural development of the Quadrilátero Ferrífero, Minas Gerais, Brazil. *USGS Prof. Paper*, **641**-A, 110 p.

Duarte, P.M. and Fontes, G.F. (1986). Minas de cromita das Fazendas Limoeira e Pedrinhas, Município de Campo Formoso, Bahia. In C. Schobbenhaus and D.E.S. Coelho (eds.) *Principais Depósitos Minerais do Brasil*, DNPM/CVRD, Brasília. Brazil, **2**, 207-214.

Dussin, I.A. (1994). *Evolution structurale de la partie méridionale de l'"Espinhaço sur la bordure orientale du Craton São Francisco, Minas Gerais - Brésil: un exemple de tectonique protérozoique superposé.* Dr. Sci. thesis, University of Orleans, France, 200 p.

Endo, I. and Machado, R. (1998). The geologic architeture of the Quadrilátero Ferrífero (Minas Gerais, Brazil) after multiple transpressional and transtensional tectonic events. In: *Int. Conf. Basement Tectonics, Ouro Preto, Minas Gerais, 14×, Extended Abstr.*, 126-127.

Evensen, N.M., Hamilton, P.J. and O'Oniona, R.K. (1978). Rare-earth abundances in chondritic meteorites. *Geoch. Cosmoch. Acta*, **42**, 1199-1212.

Feybesse, J.L., Johan, V., Triboulet, C., Guerrot, C., Mayaga Mikolo, F., Bouchot, V. and Eko N'dong, J. (1998). The West Central African Belt, a model of 2.5-2.0 Ga accretion and two-phase orogenic evolution. *Precambrian Research*, **87**(3-4), 161-216.

Figueiredo, M.C.H. (1981). Geoquímica das rochas metamórficas de alto grau no nordeste da Bahia, Brasil. *Geologia e Recursos Minerais do Estado da Bahia, Textos Básicos*, SGM/CPM, **4**, 1-17.

Figueiredo, M.C.H. (1989). Geochemical evolution of eastern Bahia, Brazil: A probably Early-Proterozoic subduction-related magmatic arc. *Journal of South American Earth Sciences*, **2**(2), 131-145.

Figueiredo, M.C.H. and Barbosa, J.S.F. (1993). Terrenos metamórficos de alto grau do cráton do São Francisco. In: O Cráton de São Francisco, eds. Dominguez, J.M.L. and Misi, A. *II Simp. Cráton São Francisco Publ. Esp., Salvador, SGB/SGM/CNPq*, **213**, 99 p.

Figueiredo, M.C.H. and Teixeira, W. (1996). The Mantiqueira metamorphic complex, Eastern Minas Gerais State: preliminary geochronological and geochemical results. *Anais da Academia Brasileira de Ciências*, **68**(2), 223-246.

Fischel, D.P., Pimentel, M.M., Fuck, R.A., Costa, A.G. and Rosière, C.A. (1998). Geology and Sm-Nd isotopic data for the Mantiqueira and Juiz de Fora complexes (Ribeira belt) in the Abre Campo - Manhuaçu region, Minas Gerais. In: *Int. Conf. Basement tectonics, 14×. Ouro Preto, MG, Brazil. Abstr.*, 21-23.

Fogaça, A.C.C., Almeida-Abreu, P.A. and Schorscher, H.D. (1984). Estratigrafia da sequência supracrustal arqueana da porção mediana central da Serra do Espinhaço – MG. In: *Brazilian Geol. Congr., 33°, Rio de Janeiro, SBG., Anais.*, **2**, 2652-2667.

Fornari, A. (1992). *Petrologia, geoquímica e metamorfismo das rochas enderbíticas-charnockíticas da região de Laje e Mutuípe, Bahia.* Dissert. Mestrado, Inst. Geociências, Univ. Fed. Bahia, 143 p.

Gaál, G., Teixeira, J.B.G., D'el Rey Silva, L.J.H. and Silva, M.G. (1987). Early Proterozoic Crustal Evolution and Metallogenesis, Northwestern Bahia, Brazil. *Intern.Symp. Granites Assoc. Mineralizations* - ISGAM, Salvador.

Guimarães, D. (1931). Contribuição à geologia do Estado de Minas Gerais. *Bol. Serviço Geológico e Mineralógico*, **55**, 1-36.

Gomes, R.A.A.D., Gomes, P.J.P. and Silveira Filho, N.C. (1994). *Mapa Bouguer do Estado da Bahia*. MME/CPRM - Serv. Geol. Bras., Publ. Espec., 1 mapa.

Gomes, R.A.A.D., Arcanjo, J.B.A. and Reginaldo, A.S. (1991). Colisão de blocos com subducção na costa sul da Bahia. In: *2nd Int. Congr. Brazilian Geophys. Soc., Salvador, Ext. Abstr.*, 1, 154-159.

Heilbron, M., Gonçalves, M.L., Teixeira, W., Trouw, R.A.J., Padilha, A.V. and Kawashita, K. (1989). Geocronologia da Região entre Lavras, São João del Rei, Lima Duarte e Caxambu (MG). *Anais da Academia Brasileira de Ciências*, **61** (2): 177-199.

Herz, N. (1970). Gneissic and igneous rocks of the Quadrilátero Ferrífero, Minas Gerais, Brazil. *USGS Prof. Paper*, **641**-B, 1-58.

Inda, H.A.V. and Barbosa, J.S.F. (1978). *Texto Explicativo para o Mapa Geológico do Estado da Bahia - 1: 1,000,000 scale*. SME/CPM, Salvador, 137 p.

Johnson, A.D. and Wyllie, P.J. (1988). Constraints on the origin of Archaean trondhjemites based on phase relationships of Nûk gneiss with H_2O at 15 kbar. *Contributions to Mineralogy and Petrology*, **86**, 398-408.

Kishida, A. and Riccio, L. (1980). Chemostratigraphy of lava sequences from Rio Itapicuru greenstone belt, Bahia State, Brazil. *Precambrian Research*, **11**, 161-178.

Kröner, A. (1984). Evolution, growth and stabilization of the Precambrian lithosphere. *Physics and Chemistry of the Earth*, **15**, 69-106.

Lacerda, M.P.C., Bilal, E. and Correia-Neves, J.M. (1996). Archean and Late Archean crustal evolution within the Moeda complex in the southern São Francisco craton (Minas Gerais). In: *Brazilian Geol. Congr., Anais. 30×, Salvador, SBG*, **6**, 396-399.

Ladeira, E.A. (1980). *Metallogenesis of gold at the Morro Velho Mine and in Nova Lima District, Quadrilátero Ferrífero, Minas Gerais, Brazil*. London (Ontario), PhD Thesis, University of Western Ontario, 272 p.

Leahy, G.A.S., Conceição, H., Rosa, M.L.S., Macambira, M.J.B., Martin H., Paim, M.M., Santos, E.B. and Bastos Leal, L.R. (1998). Maciço Sienítico de Ceraima (Sudoeste da Bahia): Idade, petrografia e geoquímica do magmatismo pós-orogênico alcalino-potássico com afinidade lamprofírica. In: Contribuição ao Estudo dos Granitos e Rochas Correlatas, ed. H. Conceição. *SBG - Núcleo Ba-Se, Publ. Esp.* **5**, 61-77.

Leahy, G.A.S., Rosa, M.L.S., Conceição, H., Scheller, T., Macambira, M.J.B., Oberli F., Meier, M., Martin, H., Barreto dos Santos, E. and Paim, M.M. (1999). Geochronology and isotopic signature of the syenitic intrusion of Ceraima, Southwest Bahia, Brazil. In: *II South Amer. Symp. on Isotope Geology, Córdoba, Argentina, Actas*, 234-237.

Ledru, P., Cocherie, A., Barbosa, J.S.F., Johan, V. and Onstott, T. (1994). Âge du métamorphisme granulitique dans le craton du São Francisco (Brésil): implications sur la nature de l'orogène transamazonien. *Comptes Rendus de l'Académie des Sciences, Paris*, **318**, 251-257.

Ledru, P., Milési, J.P., Johan, V., Sabaté, P. and Maluski, H. (1997). Foreland basins and gold-bearing conglomerates, a new model for the Jacobina basin (São Francisco province, Brazil). *Precambrian Research*, **86**, 155-176.

Machado, N. and Carneiro, M.A. (1992). A major Archean tectonothermal event in the São Francisco shield, Brazil: U/Pb evidence from the Quadrilátero Ferrífero, Minas Gerais. *Canadian Journal of Earth Sciences*, **29**(11), 2341-2346.

Machado , N. and Schrank, A. (1989). Geocronologia U/Pb no Maciço de Piumhí- resultados preliminares. In: *Simp. Geol. Núcleo Minas Gerais, 5, Simp. Geol. Núcleo Brasília, 1., Belo Horizonte, 1989. Anais, Belo Horizonte, SBG*, 45-49.

Machado, N., Noce, C.M., Belo de Oliveira, O. and Ladeira, E.A. (1989a). Evolução Geológica do Quadrilátero Ferrífero no Arqueano e Proterozóico Inferior com base em Geocronologia U-Pb. In: *Simp. Geol. Núcleo Minas Gerais, Belo Horizonte, SBG, Bol.* **10**, 1-5.

Machado, N., Schrank, A., Abreu, F.R., Knauer, L.G. and Almeida-Abreu, P.A. (1989b). Resultados preliminares da geocronologia U-Pb na Serra do Espinhaço Meridional, In: *Simp. Geol. Núcleo Minas Gerais, SBG, Belo Horizonte, Bol.* **10**, 171-174.

Machado, N., Noce, C.M. and Carneiro, M.A. (1992a). U-Pb age constraints on a major Late Archean event in the Southern São Francisco Craton, Brazilian Shield. In: *GAC/AGC-MAC/AMC Joint Ann. Meet., Abstr., Nova Scotia*, **17**, A70.

Machado, N., Noce, C.M., Ladeira, E.A. and Belo de Oliveira, O. (1992b). U-Pb geochronology of Archean magmatism and Proterozoic metamorphism in the Quadrilátero Ferrífero, southern São Francisco Craton, Brazil. *Geological Society of America Bulletin*, **104**, 1221-1227.

Machado, N., Schrank, A., Noce, C.M. and Gauthier, G. (1996). Ages of detrital zircon from Archean-Proterozoic sequences: implications for Greenstone Belt setting and evolution of a Transamazonian foreland basin in Quadrilátero Ferrífero, southwest Brazil. *Earth and Planetary Science Letters*, **141**, 259-276.

Marinho, M.M.M. (1991). *La séquence volcano-sédimentaire de Contendas-Mirante et la bordure occidentale du bloc Jequié (Craton du São Francisco, Brésil), Un exemple de transition Archéen-Protérozoïque*. Thèse doct., Univ. Blaise Pascal, Clermont-Ferrand, France, 388 p.

Marinho, M.M.M. and Sabaté, P. (1982). The Contendas-Mirante volcano-sedimentary sequence and its granitic migmatitic basement. In: *Internat. Symp. Archean and Early Proterozoic Geologic Evolution and Metallogenesis (ISAP), Secr. Minas Energia /CPM, Salvador, Excursion guide*, 139-184.

Marinho, M.M.M., Silva, E.F.A., Andrade, A.P., Senna Filho, A.A., Morais, A.M.V., Silva, B.C.E. Soares, J.V., Cunha, J.C., Silva, J.C.E, Santos, J.G., Cruz, M.J.M., Oliveira, N.S., Costa, P.H., Mota, P.R.R., Rangel, P. A., Fróes, R.J.B., Souza, S.L., Lago, S.E.S, Silveira, W.P., Souza, W.S.T., Torquato, J.R.F., Coutinho, J.M.V., Lewis, R.W. and Monteiro, T.P.R. (1979). *Projeto Contendas-Mirante*, CBPM/SME, Bahia, Salvador, 24 v.

Marinho, M.M.M., Sabaté, P. and Barbosa, J.S.F. (1994a). The Contendas-Mirante volcano-sedimentary belt. In: Petrologic and Geochronologic evolution of the oldest segments of the São Francisco Craton, Brazil*Bol. IG-USP*, **17**, 37-72.

Marinho, M.M.M, Vidal, Ph. Alibert, C., Barbosa, J.S.F. and Sabaté, P. (1994b). Geochronology of the Jequié-Itabuna granulitic belt and of the Contendas-Mirante volcano-sedimentary belt. In: Petrologic and Geochronologic evolution of the oldest segments of the São Francisco Craton, Brazil, eds. Figueiredo, M.C.H. and Pedreira, A.J. *Bol. IG-USP*, 17, 73-96.

Marshak, S. and Alkmim, F.F. (1989). Proterozoic extension/contraction tectonics of the southern São Francisco craton and adjacent regions, Minas Gerais, Brazil: a kinematic model relating Quadrilátero Ferrífero, São Francisco basin and Cordilheira do Espinhaço. *Tectonics*, 8, 555-571.

Martin, H. (1987). Petrogenesis of Archaean trondhjemites, tonalites and graniodiorites from eastern Finland: Major and trace element geochemistry. *Journal of Petrology*, 28, 921-953.

Martin H. (1994). The Archaean grey gneisses and the genesis of continental crust. In: *The Archean crustal evolution*, ed. Condie K.C., pp. 205-259. Elsevier.

Martin H., Sabaté, P., Peucat, J.J. and Cunha, J.C. (1991). Un segment de la croûte continentale d'âge archéean ancien (3.4 millards d'années): le massif de Sete Voltas (Bahia, Brésil). *Comptes Rendus de l'Académie des Sciences, Paris*, 313, 31-538.

Martin, H. Peucat, J-J, Sabaté, P. and Cunha, J.C. (1997). Crustal Evolution in Early Archaean of South America: Example of the Sete Voltas Massif, Bahia state, Brazil. *Precambrian Research*, 82, 35-62.

Mascarenhas, J.F. and Garcia, T.M. (1989). *Mapa geocronólogico do estado da Bahia. Texto Explicativo*. Secr. Minas Energia Estado Bahia, Salvador, 130 p.

Mascarenhas, J.F. and Silva, E.F.A. (1994). *Greenstone belt de Mundo Novo, caracterização e implicações metalogenéticas e geotectônicas no Cráton do São Francisco*. CBPM, Salvador, Arquivos Abertos, 5.

Massuda, A., Nacamura, N. and Tanaka, T. (1973). Fine structures of mutially normalized rare-earth pattern of chondrites. *Geochimica et Cosmochimica Acta*, 36, 239-348.

Müller, G., Hohndorf, A., Lauenstein, H.J. and Lenz, H. (1986). Petrological and geochemical data on a high metamorphic Archean BIF-bearing rock sequence near Guanhães, Minas Gerais, Brazil. *Geologische Jahrbuch*, Hannover, 79, 3-20.

Matos, F.M.V. and Davidson, I. (1988). Basement or intrusion? the Ambrósio dome, Rio Itapicuru greenstone belt, Bahia, Brazil. *Revista Brasileira de Geociências*, 17, 442-449.

Mello, E.F., Lacerda, C.M.M., Oliveira, E.P. and McNaughton, N.J. (1999a). SHRIMP U-Pb geochronology on xenotime and zircon from the Ambrósio dome, Rio Itapicuru greenstone belt, Brazil: A major syntectonic granodiorite intrusion. In: *II South Amer. Symp. on Isotope Geology. Córdoba, Argentina, Actas*, 331-334.

Mello, E.F., Oliveira, E.P. and McNaughton, N.J. (1999b). SHRIMP U-Pb geochronology of Early Precambrian quartzite and its basement (Caldeirão belt), N.E. São.Francisco craton, Bahia. In: *II South Amer. Symp. on Isotope Geology. Córdoba, Argentina, Actas*, 78-81.

Melo, R.C. (1991). *Programa de Levantamentos Geológicos Básicos do Brasil (PLGB), Folha* SC.24-Y-D-V: *Pintadas, Estado da Bahia*. DNPM, Brasília, Convênio DNPM/CPRM, 192 p. (2 mapas).

Molinari, L. (1983). Mineralizações auríferas em Jacobina, Bahia. In: *Simp. Mineral. Aurif. Estado da Bahia, Salvador, Anais Soc. Bras. Geol., CBPM/SGB/SME*, 26-31.

Motta, A.C., Gomes, R.A.A.D, Delgado, I.M., Siqueira, L.P. and Pedreira, A.J. (1981). Feições gravimétricas e magnéticas do cráton de São Francisco. In *Simpósio do Cráton de São Francisco e Suas Faixas Marginais, Salvador, 1979. Anais Soc. Bras. Geol. - Núcleo Bahia, Salvador*, 1, 17-33.

Mougeot, R. (1996). *Étude de la limite Archéen Archéen-Protérozoique et des minéralisations Au, U associées: Exemples de la région de Jacobina (Etat de Bahia, Brésil) et de Carajás (Etat de Pará, Brésil)*. Thèse doct. Univ. Montpellier, 301 p.

Mougeot, R., Respaut, J.P., Ledru, P. and Milési, J.P. (1995). U-Pb geochronological constraint for the evolution of the Palaeoproterozoic Jacobina auriferous basin (São Francisco province, Bahia, Brazil). EUG Strasbourg. *Terra Nova, Abstracts*, 7, 354.

Myers, J.S. (1995). The generation and asembly of an Archean supercontinent: evidence from the Yilgarn craton, Western Australia. In: Early Precambrian Processes, eds. Coward, M.P. and Ries, A.C. *Geological Society Special Publ.* 95, 143-154.

Nascimento, H.S. (1996). *Evolução petrogenética da intrusão ácido-básica de Cansanção*. Dissert. Mestrado. Inst. Geociências, Univ. Federal Bahia, 205 p.

Noce, C.M., Teixeira, W. and Machado, N. (1997). Geoquímica dos gnaisses TTGs e granitóides neoarqueanos do Complexo Belo Horizonte, Quadrilátero Ferrífero, Minas Gerais. *Revista Brasileira de Geociências*, 27(1), 25-32.

Noce, C.M., Pinheiro, S.O., Ladeira, E.A., Franca, C.R., Kattach, S., 1992. A sequência vulcanossedimentar do Grupo Nova Lima na região de Piedade do Paraopeba, oeste do Quadrilátero Ferrífero. *Revista Brasileira de Geociências*, 22(2), 175-183.

Noce, C.M., Machado, N. and Teixeira, W. (1998). U-Pb Geochronology of gneisses and granitoids in the Quadrilátero Ferrífero (Southern São Francisco Craton): age constraints for Archean and Paleoproterozoic magmatism and metamorphism. *Revista Brasileira de Geociências*, 28(1), 95-102.

Noce, C.M., Teixeira, W., Quéméneur, J.J.G., Martins, V.T.S. and Bolzachini, E. (1999). Isotopic signatures of Paleoproterozoic granitoids from the southern São Francisco Craton and implications for the evolution of the Transamazonian Orogeny. *Journal of South American Earth Sciences* (submitted).

Nutman, A.P. and Cordani, U.G. (1993). SHRIMP U-Pb zircon geochronology of Archaean granitoids from the Contendas-Mirante area of the São Francisco craton, Bahia, Brazil. *Precambrian Research*, 63, 179-188.

Nutman, A.P., Cordani, U.G. and Sabaté, P. (1994). SHRIMP U-Pb ages of detrital zircons from the Early Proterozoic Contendas-Mirante supracrustal belt, São Francisco craton, Bahia, Brazil. *Journal of South American Earth Sciences*, 7, 109-114.

Oliveira, A.H. (1999a). *Evolução tectônica de um segmento crustal arqueano do Craton do São Francisco*. Dissert. Mestrado. Escola de Minas, Univ. Federal de Ouro Preto, Ouro Preto, 104 p.

Oliveira, E.A. (1999b). *Geologia, Petrografia e geoquímica do Maciço granitoide de Cachoeira da Prata*. Dissert. Mestrado. Univ. Federal de Minas Gerais, Belo Horizonte, 80 p.

Oliveira, E.P. and Knauer, L.G. (1993). Corpos máficos e ultramáficos do Craton do São Francisco. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and Misi, A., pp. 119-136. SBG-Sup. Geol. Rec. Min., Bahia - CNPq. Salvador.

Oliveira, E. A., Correia Neves, J. M., Fetter, A. and Bilal, E. (1998). Geocronologia U-Pb do maciço granitóide de Cachoeira da Prata - MG. In: *Brazilian Geol. Congr.*, 35. *Belo Horizonte, 1998, Anais...*, BH, SBG, 508 p.

Oliveira, L.L., Rios, D.C., Burgos, C.M.G., Conceição, H., Macambira, M.J.B., Santos, C.G.P. and Scheller, T. (1999). Geochronology of Cansanção massif by $^{206}Pb/^{207}Pb$ evaporation method, Northeastern of Bahia state, Brazil. In: *II South American Symp. on Isotope Geology. Córdoba, Argentina, Actas*, 90-93.

Olivo, G.R., Gauthier, M., Gariépy, C. and Carignan, J. (1996). Transamazonian tectonism and Au-Pd mineralization at the Cauê Mine, Itabira District, Brazil: Pb isotopic evidence. *Journal of South American Earth Sciences*, 9, 273-279.

Padilha, A.V. and Melo, R.C. (1991). Estruturas e Tectônica. In: Programa de Levantamentos Geológicos Básicos do Brasil (PLGB), Folha SC.24-Y-D-V, Pintadas, Estado da Bahia. *DNPM, Brasília, Convênio DNPM/CPRM*, 3, 49-54.

Padilha, A.V., Santos, R.A., Martins, A.A.M., Arcanjo, J.B.A., Oliveira, J.E. and Gomes, R.A.A.D. (1990). O ciclo Jequié no Sudeste da Bahia. Uma colisão arco de ilhas-continente no Arqueozóico Superior. In: *Brazilian Geol. Congr.* 36°, *Natal, SBG*, 345.

Paim, M.M. (1998). Petrologia da intrusão sienítica potássica de Cara Suja (Sudoeste da Bahia). In: Sienitos alcalino-potássicos e ultrapotássicos paleoproterozóicos do Estado da Bahia, eds. Conceição, H. and Cruz, M.J.M. *SBG - Núcleo Ba-Se, Publ. Especial*, 4, 112-145.

Pearce, J.A. (1982). Trace element characteristics of lavas from destructive plate boundaries. In: *Andesites, Orogenic andesites and related rocks*, ed. Thorpe, R.S., pp. 525-547. Chichester, J. Wiley and Sons.

Pearce, J.A. (1983). Role of the Sub-continental lithosphere in Magma Genesis at Active Continental Margins. In: *Continental Basalts and Mantle xenoliths*, eds. Hawkesworth, C.J. and Norry, M.J., pp. 230-349. Nantwich, Shiva.

Pearce, J.A., Harris, N.B.W. and Tindle, A.G. (1984). Trace Element Discrimitation Diagrams for the tectonic interpretation of granitic rocks. *Journal of Geology*, 25 (4), 956-983.

Pedrosa Soares, A.C., Noce, C.M., Vidal, Ph., Monteiro, R.L.B.P. and Leonardos, O.H. (1992). Toward a new tectonic model for the late Proterozoic Araçuaí (SE Brazil) - West Congolian (SW Africa) belt. *Journal of South American Earth Sciences*, 6 (1/2), 33-47.

Pedrosa Soares, A.C., Dardenne, M.A., Hasui, H., Castro, F.D.C., Carvalho, M.V.A. and Reis, A.C. (1994). *Mapa Geológico de Minas Gerais, 1: 1,000,000 scale*. COMIG, Governo de Minas Gerais - Paulo Abib Engenharia.

Pereira, L.H.M. (1992). *Programa de Levantamentos Geológicos Básicos do Brasil (PLGB), Folha SC.24-Y-D-VI, Serrinha*, Estado da Bahia. DNPM, Brasília, Brazil, Convênio DNPM/CPRM, 132 p.

Pinese, J.P.P., Teixeira, W., Piccirillo, E.M., Quéménéur, J.J.G. and Bellieni, G. (1995). The Precambrian Lavras Mafic dykes, southern São Francisco Craton, Brazil: preliminary geochemical and geochronological results. In: *Physics and Chemistry of Dykes*, eds. Baer, G. and Heimann, A., pp. 205-218. Balkema, Rotterdam,.

Reinhardt, M.C. and Davison, I. (1990). Structural and lithologic control on gold deposition in the shear zone hosted Fazenda Brasileiro mine, Bahia state, northern Brazil. *Economic Geology*, 85, 952-967.

Renger, F.E., Noce, C.M., Romano, A.W. and Machado, N. (1994). Evolução sedimentar do Supergrupo Minas (500 Ma de registro geológico), no Quadrilátero Ferrífero, Minas Gerais, Brazil. *Geonomos*, 2, 1-11.

Ribeiro, A., Paciullo, F.V.P., Valença, J.G., Ávila, C.A., Andreis, R.R., and Trouw, R.A.J. (1998). Domínios do embasamento na região de São João Del Rei, borda sul do Cráton do São Francisco, Sul de Minas Gerais. in: *Brazilian Geol. Congr., 35×, Belo Horizonte, Anais..., SBG*, 21 p.

Romano, A.W. (1989). *Evolution tectonique de la région NW du Quadrilatère Ferrifère - Minas Gerais, Brésil*. Thèse Docteur, Université de Nancy I, 259 p.

Romano, A.W., Bilal, E., Correa Neves, J.M., Giret, A. and Moutte, J. (1995). O Complexo granitóide de Florestal - parte meridional do Cráton do São Francisco - primeiros dados petrográficos e geoquímicos. In: *Simp. Geol. Minas Gerais, Diamantina, MG, SBG, Núcleo Minas Gerais, Bol.* 13, 86-88.

Rios, D.C. (1998). Petrologia do magmatismo potássico-ultrapotássico e lamprofírico de Morro do Afonso - Bahia. In: Sienitos alcalino-potássicos e ultrapotássicos paleoproterozóicos do Estado da Bahia, eds. Conceição, H. and Cruz, M.J.M. *Soc. Bras. Geol. - Núcleo Ba/Se, Publ. Esp.* 4, 168-215.

Rios, D.C., Conceição, H., Macambira, M.J.B., Burgos, C.M.G., Peixoto, A.A., Cruz Filho, B.E., Oliveira, L.L. and Lisboa, M.P. (1998). Granitogênese da parte meridional-oriental do núcleo Serrinha: Idade, petrografia e geoquímica. In: *Contribuição ao Estudo dos Granitos e Rochas Correlatas*, ed. Conceição, H. *SBG,-Núcleo Ba-Se, Publ. Especial*, 5, 91-113.

Rosa, M.L.S., Conceição, H., Paim, M.M., Santos, E.B., Alves da Silva, F.C., Leahy, G.A.S. and Bastos Leal, L.R. (1996). Magmatismo potássico ultrapotássico pós- a tardi- orogênico no oeste da Bahia: Batólito monzo-sienitico de Guanambi-Urandi e sienitos de Correntina. *Geochimica Brasiliensis*, 10, 27-42.

Rosa, M.L.S., Conceição, H., Conceição, R.V.., Rios, D.C., Nardi, L.V.S. Martin, H., Oberli, F., Meier, M., Scheller, T., Macambira, M.J.B., McReath, I., Pereira Santos, C.G, Barreto Santos, E., Paim, M.M. and Leahy, G.A.S. (1999). Isotopic of the Paleoproterozoic potassic-ultrapotassic syenites from Bahia state, Brazil. In: *: II South Amer. Symp. on Isotope Geology. Córdoba, Argentina, Actas*, 272-274.

Rudowski, L., Giuliani, G. and Sabaté, P. (1987). Les phlogopitites à émeraudes au visinage des granites de Campo-Formoso et de Carnaiba (Bahia, Brésil): un exemple de minéralisations protérozoïques à Be, Mo et W dans des ultrabasites métasomatisées. *Comptes Rendus de l'Académie des Sciences, Paris*, 307, 935-940.

Sabaté, P. (1991). Evolution transamazonienne et structures de collison dans le Craton du São Francisco, Brésil. In: *Evolution crustale du protérozoique inférieur, Afrique de l'ouest et Amérique du Sud.* CNRS-Rennes, 10.

Sabaté, P. (1993). Granites and tectonic evolution of the Contendas-Jacobina lineament. In: Magmatismo Granítico e Mineralizações Associadas. *Workshop Academia Brasileira de Ciências, Rio de Janeiro. Extended Abstract.*, 40-43.

Sabaté, P. (1996). Estruturas e tectônica do embasamento Arqueano e Proterozóico Inferior do Estado da Bahia. In: *Texto Explicativo para o Mapa Geológico da Bahia ao Milionésimo*, coord. Barbosa, J.S.F and Dominguez, J.M.L., pp. 201-226. Publ. Especial SICM/SGM, Salvador-Ba.

Sabaté, P. and Marinho, M.M.M. (1982). Chemical affinities of low-grade metamorphic formations of the Contendas-Mirante Complex. *Revista Brasileira de Geociências*, 12, 302-402.

Sabaté, P. and Cruz, M.J.M. (1998). Configuração estrutural e contexto dinâmico das intrusões gabro-anortosíticas no cinturão granulítico da Costa Atlântica (cráton do São Francisco, Bahia, Brasil). In: *Brazilian Geol. Congr.*, 40°, *Belo Horizonte. Anais SBG*, 3, 140.

Sabaté, P., Machado, G.V. and Souza, M.Z.A. (1980). Données structurales des formations précambriennes épimétamorphiques du complexe Contendas-Mirante (Bahia, Brésil). *Cahiers Orstom, Géologie*, 11, 1-13.

Sabaté, P., Correa Gomes, L.C. and Anjos, J.A.S.A. (1988). *Mapa temático "Granitogênese da Bahia" - Folha Vitória da Conquista 1/250.000.* Secr. Minas Energia/SGM-Bahia, Salvador - Aerofoto, Rio de Janeiro, 1 mapa).

Sabaté, P., Barbosa, J.S.F. and Marinho, M.M.M. (1994). The São Francisco craton: a short outline. In: Petrologic and Geochronologic evolution of the oldest segments of the São Francisco craton, Brazil, eds. Figueiredo, M.C.H. and Pedreira, A.J. *Bol. IG-USP*, 17, 1-8.

Sabaté, P., Marinho, M.M.M, Vidal, P. and Caen-Vachette, M. (1990). The 2 Ga peraluminous magmatism of Jacobina - Contendas-Mirante belts (Bahia, Brazil), Geologic and isotopic constraints on sources. *Chemical Geology*, 83, 325-338.

Santos Pinto, M.A, Peucat, J.J, Martin H. and Sabaté, P. (1998). Recycling of the Archaean continental crust: the case study of the Gavião, state of Bahia, NE Brazil *Journal of South American Earth Sciences*, 11, 487-498.

Santos, R.A. and Souza, J.D. (1983). *Projeto Mapa Metalogenético e de Previsão de Recursos Minerais - Escala 1/250.000*, Folha SC.24-Y-D, Serrinha, CPRM, Salvador, Convênio CPRM/DNPM, 2 vol.

Sato, K. (1998). *Evolução crustal da plataforma Sul Americana, com base na geoquímica isotópica Sm-Nd*. Tese Dout., Universidade de São Paulo, Brazil, 297p.

Scarpelli, W. (1991). *Aspects of gold mineralization in the iron* quadrangle, Brazil. In: *Brazil Gold 91*, ed. Ladeira, E.A., pp. 11-30. Balkema, Rotterdam.

Seixas, L.A.R. and Baars, F.J. (1991). Características geoquímicas das rochas ultramáficas do Lineamento Congonhas, Minas Gerais, Brasil. In: *Congr. Brasileiro de Geoquímica*, 3×, *Anais*, 1, 185-190.

Schorscher, H.D., Santana, F.C., Polonia, J.C. and Moreira, J.P.M. (1982). Quadrilatero Ferrífero-Minas Gerais State: Rio das Velhas greenstone belt and Proterozoic rocks. In: Excursions Annex, *Intern. Symp. Archean and Early Proterozoic Geologic Evolution and Metallogenesis (ISAP)*. Coord. Prod. Mineral, Secr. Minas e Energia do Estado da Bahia, Salvador, Bahia, 44 p.

Schorscher, H.D. (1992). *Arcabouço petrográfico e evolução crustal de terrenos precambrianos do sudeste de Minas Gerais: Quadrilátero Ferrífero, Espinhaço meridional e domínios granito-gnáissico adjacentes*. Tese de Livre Docência. Instituto de Geociências, University of São Paulo, São Paulo, 2 vol., 349 p.

Schrank, A. and Brousse, R. (1979). Découverte de komatiites en Amérique du Sud (Bresil). *Comptes Rendus de l'Académie des Sciences, Paris*, **290**, 151-154.

Schrank, A. and Silva, M.G. (1993). Greenstone belt do cráton de São Francisco, Brasil. In: *O Cráton de São Francisco*, eds. Dominguez, J.M.L. and Misi, A., pp. 85-118. Soc. Bras. Geol., SBG/SGM/CNPq, Salvador.

Silva, A.M., Chemale Junior, F., Kuyumjian, R.M. and Heaman, L. (1995). Mafic dike swarms of Quadrilátero Ferrífero and Southern Espinhaço, Minas Gerais, Brazil. *Revista Brasileira de Geociências*, **25**(2), 124-137.

Silva, M.G. (1987). *Geochimie, petrologie und geotketnonische etwicklung eins Proterozoischen grensteignertlels, Rio Itapicuru, Bahia, Brazil*. Doct. Thesis, Univ. Freiburg, Germany, 125 p.

Silva, M.G. (1992). Evidências isotópicas e geocronológicas de um fenômeno de acrescimento crustal transamazônico no cráton São Francisco, Estado da Bahia. In: *Brazilian Geol. Congr. 37, São Paulo, Anais. São Paulo, SBG*, 2, 181-182.

Silva, M.G. and Vidal, Ph. (1992). Evidences of an arc-continent collision in the NE portion of the São Francisco Craton. *14° Réun. Sci. Terre., Toulouse, Soc. Géol., France*, 69.

Söllner, F., Lammerer, B. and Weber-Diefenbach, K. (1991). Die Krustenentwicklung in der Küstenregion nördkich von Rio de Janeiro: Altersbestimmungen (U-Pb an Zirkonen und Rb-Sr an Gesteinen) an hochdruck- und temperaturfaziellen Gesteinen des Ribeira mobile Belt und des São Francisco Kraton (Espírito Santo/ Minas Gerais). *Münchner Geologische Hefte*, **4**, 101 p.

Söllner, F. and Trouw, R.A.J. (1997). The Andrelândia depositional cycle (Minas Gerais/ Brazil), a post-Transamazonian sequence south of the São Francisco Craton: evidence from U-Pb dating on zircons of a metasediment. *Journal of South American Earth Sciences*, **10** (1): 21-28.

Sumner, D.Y. and Bowring, S.A. (1996). U-Pb geochronologic constraints on deposition of the Cambellrand Subgroup, Transvaal Supergroup, South Africa. *Precambrian Research*, **79**, 25-35.

Teixeira, L.R. (1991). Geoquímica. In *Programa de Levantamentos Geológicos Básicos do Brasil (PLGB), Folha SC.24-Y-D-V, Pintadas, Estado da Bahia*. DNPM, Brasília, Convênio DNPM/ CPRM, Cap. 5.

Teixeira, L.R. (1997). *O Complexo Caraíba e a Suíte São José do Jacuípe no Cinturão Salvador-Curaçá (Bahia-Brasil), Petrologia, Geoquímica e Potencial Metalogenético*. Tese Dout., Univ. Federal Bahia, Salvador, 202 p.

Teixeira, L.R. and Melo, R.C. (1990). Geoquímica dos ortognaisses granulíticos do Complexo Carnaíba, região de Riachão de Jacuipe, Bahia. In: *Brazilian Geol. Congress, 36° Natal. Anais, Soc. Bras. Geol.*, 6, 1849-1860.

Teixeira, J.B.G., Kishida, A., Marimon, M.P., Xavier, R.P. and McReath, I. (1990). The Fazenda Brasileiro gold deposit, Bahia: hydrothermal alteration and fluid inclusion studies. *Economic Geology*, **85**, 990-1009.

Teixeira, W. (1993). Avaliação do acervo de dados geocronológicos e isotópicos do Cráton do São Francisco - implicações tectônicas. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and Misi, A., pp. 11-33. SBG – Sup. Geol. Rec. Minerais, Bahia - CNPq. Salvador.

Teixeira, W. and Figueiredo, M.C.H. (1991). An outline of Early Proterozoic crustal evolution in the São Francisco craton, Brazil: a review. *Precambrian Research*, **53**(1/2), 1-22.

Teixeira, W., Jordt-Evangelista, H., Kawashita, K. and Taylor, P.N. (1987). Complexo granulítico de Acaiaca, MG: idade, petrogênese e implicações tectônicas. In: *Simpósio de Geologia de Minas Gerais, Núcleo - MG, Belo Horizonte. 4×, Anais. SBG*, 58-71.

Teixeira, W., Carneiro, M.A., Noce, C.M., Machado, N., Sato, K. and Taylor, P.N. (1996). Pb, Sr and Nd isotope constraints on the *Archean evolution of gneissic-granitoid complexes in the* southern São Francisco Craton, Brazil. *Precambrian Research*, 78(1/3), 151-164.

Teixeira, W., Onstott, T.C., Makenya, M. and Szabó, G.A.J. (1997). Proterozoic thermochronologic implications from $^{40}Ar/^{39}Ar$ and K-Ar dating of the Campo Belo metamorphic Complex, southern São Francisco Craton, Brazil. *Anais da Academia Brasileira de Ciências*, **69** (1): 559-575.

Teixeira, W., Cordani, U.G., Nutman, A.P. and Sato, K. (1998a). Polyphase Archean evolution in the Campo Belo metamorphic complex, Southern São Francisco Craton, Brazil: SHRIMP U-Pb zircon evidence *Journal of South American Earth Sciences*, 11 (3), 279-289.

Torquato, J.R., Tanner de Oliveira, M.A.F. and Bartels, R.L. (1978). Idade radiométrica do granito de Campo Formoso, Bahia. Uma idade mínima do Grupo Jacobina. *Revista Brasileira de Geociências*, **8**, 171-179.

Uhlein, A., Trompette, R.R. and Egydio-Silva, M. (1998). Proterozoic rifting and closure, SE border of the São Francisco Craton, Brazil. *Journal of South American Earth Sciences*, **11** (2): 191-204.

Ussami, N. (1992). Estudos geofísicos no Cráton do São Francisco, estágio atual e perspectivas. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and Misi, A., pp. 35-43. Soc. Bras. Geol., SBG/SGM/CNPq, Salvador.

Wilson, N. (1987). *Combined Sm-Nd, Pb-Pb and Rb/Sr geochronology and isotope geochemistry in polymetamorphic Precambrian terrains, examples from Bahia, Brazil and Channel Island* - U.K. Master Thesis, Univ. Oxford, Great Britain.

Windley, B.F. (1995). *The Evolving Continents*. 3d. edition. Wiley, UK, 526 p.

Buenos Aires Cratonic Region

Carlos Cingolani and Luis Dalla Salda

The Buenos Aires Craton was part of southwestern Gondwana's cratonic core (Fig. 1). In this review it is defined as consisting of three litho-structural units situated between the Rio de la Plata to the N and the North Patagonian Massif to the S. These litho-structural units are: the Rio de la Plata Craton, consisting of Precambrian basement rocks exposed on the Sierra Tandilia, and on the Martin Garcia Island; the Paleozoic fold and thrust belt of the Sierra de la Ventana; and three sedimentary basins known as the Salado, the Claromecó and the Colorado that hold Mesozoic and Cenozoic sedimentary rocks.

Sierra Tandilia Region and the Martin Garcia Island

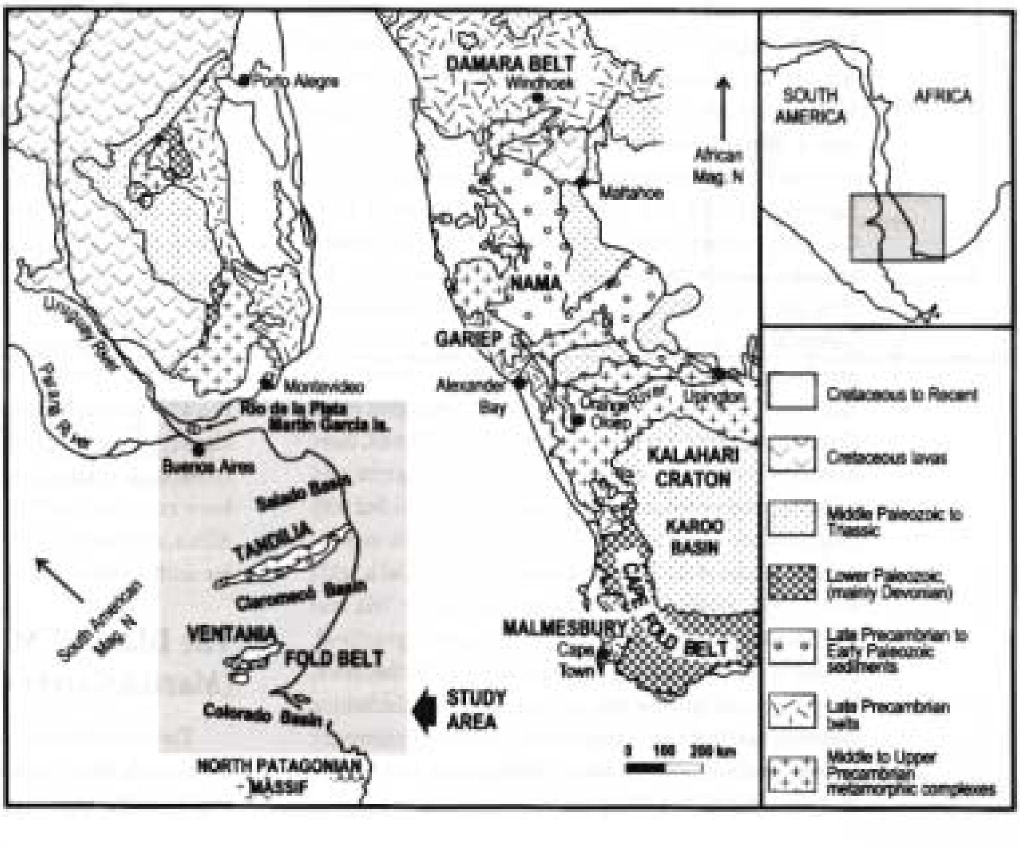
Rio de la Plata Craton (Buenos Aires Complex and Martin Garcia Complex)

The Rio de la Plata Craton underlies parts of western Uruguay, the Martin Garcia Island at the beginnning of the Rio de la Plata, and the Tandilia ranges (Fig. 2). The craton evolved during the Precambrian, and underwent both Transamazonian and Brasiliano orogenies, with more intensity in the former. Teruggi and Kilmurray (1980) and Dalla Salda *et al.* (1987, 1988), gave dates of 2.2 to 1.8 Ga. These rocks are the oldest records of southwestern Gondwana's basement in South America. Detailed studies have demonstrated the affinity between the rocks exposed on the Sierra Tandilia (Buenos Aires Complex) with those in the Martin Garcia Island (Martin Garcia Complex), and the Precambrian rocks of Uruguay.

Petrology of the Buenos Aires Complex

The rocks of the Buenos Aires Complex consist mainly of gneiss, granitic to tonalitic in composition; migmatite; amphibolite and granite plutons (Marchese and Di Paola, 1975). Subordinate rock-types include schist (rare), marble, and dykes of acid and basic composition. Conspicuous features are wide belts of mylonite. Metavolcanic units within the Buenos Aires Complex have been described by Lema and Cucchi (1985). According to Teruggi *et al.* (1988),

FIGURE 1 - Southwestern Gondwana showing the main geological units, and the studied area in South America.



the low-grade metamorphites of the El Cortijo Formation consisting of metachert, metagreywacke and metabasite, represent a slice of oceanic crust.

To the W and SW, the complex is partially covered by two platform sequences. The oldest of these is the Serras Bayas Group, also known as La Tinta Formation of Upper Precambrian age, and the youngest is the Balcarce Formation, assigned to the Ordovician (Dalla Salda and Iñiguez, 1979). However, it is interesting to note that a borehole near the town of Mar del Plata (Fig. 3) revealed a low-grade metamorphite in the form of a highly deformed metapelite beneath the Balcarce Formation (Marchese and Di Paola, 1975). According to Cingolani and Bonhomme (1982) this metapelite was dated at around 600 Ma, and could be correlated with similar rocks in eastern Uruguay.

The consolidation of the Buenos Aires Craton is given as 1.7 Ga by Teruggi *et al.* (1974) and by Dalla Salda *et al.* (1988), based on datings obtained from diabase dykes that occur in swarms, and which post-date the late to post tectonic granite intrusions, at a time when the crust was still hot and ductile. Cataclastic rocks were derived mainly from granitoid (Gonzalez Bonorino *et al.*, 1956; Teruggi and Kilmurray, 1980). During the main episode of cataclasis, long wide belts of these rocks were formed in the Tandil and Azul areas (Fig. 3). Metamorphic rocks are conspicuous in the southern and southeastern third of the Balcarce area. These are gneissic rocks related to granulite with orthopyroxene and hornblende; schist (El Quebracho Hill); olivine marble (Punta Tota, Bachicha); pyroxene-rich ultramafic rocks (Cinco Cerros and Punta Tota) and migmatite. Detailed studies carried out in the El Cristo Hills near the town of Balcarce, showed that the main metamorphic event reached the garnet grade of the amphibolite facies, that locally underwent retrograde metamorphism to the chlorite grade.

Although non-granitized rocks are rare in the Tandil area, there occur acid metavolcanic rocks, locally with porphyroblasts and comparable to rocks observed at the Vela Hill, and at La Ribulia Hill to the S of the town of Tandil. A wollastonitic scarn has been described in the San Miguel area. Amphibolite is observed frequently in the Tandil area, especially in the southern and central regions. Granulite has been found in the Azul area to the NW. The Buenos Aires Complex has been intruded by isolated granite plutons related to the mylonitic belts mentioned above. These are typically grey granitoid (the K-feldspar is amazonite), except in the northernmost part of the complex where the granitoid rocks are red as can be observed in the Sierra Chica Hills, near the town of Olivarria. In the central region of the Complex, this is to say to the S of the town of Tandil, there occurs a wide belt consisting of tonalite, granite and mylonite, interpreted as representing plate collision, but with magmatic arc affinities, possibly representing syn to post-tectonic phases of the Transamazonian Orogeny (Dalla Salda *et al.*, 1987, 1988). These rocks are the Alta de Vela and Montecristo Leucogranite, simple collisional granitoid, related to the final phase of the main orogeny. Furthermore, the presence of epidote-rich granite suggests a thickening and/or a crust that was in rapid vertical motion. Preliminary tectonic analysis revealed some S-type granite bodies. The analysis of the texture, feldspars, micas and epidote suggests a protracted thermal evolution that included an increasing temperature-fluid condition along the granite-mylonite belt.

In the Alta de Vela and Montecristo hills, the rocks are leucogranite. Granodiorite, tonalite and granite occur to the N in such areas as the Dos Leones and Movediza quarries of the Tandil area (Dalla Salda *et al.*, 1987). Studies of the major trace elements and Rb/Sr isotope analyses on rocks collected from the El Cortijo Group have given rise to the view that some of the granitoid plutons in the Tandil area are of the continent-continent collision type (Dalla Salda *et al.*, 1987; Ramos, 1988; Varela *et al.*, 1988; Teruggi *et al.*, 1988). The collision favoured thrusting and transcurrent faulting which constituted a favourable condition for the initial development of anatexis. Furthermore, it may be suggested that the emplacement of the granite plutons in the thick gneissic sequence in the Tandil area was coeval with the regional high-temperature metamorphism, mylonitization and anatexis. The Vela Leucogranite is heterogeneous, highly radiogenic and typically post-collisional.

Tectonic evolution of the Buenos Aires Complex

The Buenos Aires Complex resulted from more than one deformational phase (Teruggi *et al.*, 1973; Dalla Salda *et al.*, 1988). The first event (F_1) was manifest by synmetamorphic E-W recumbent isoclinal folding. The second, or principal phase of deformation (F_2) that deformed the first event, is related to the 1.85 Ga granitic syntectonic emplacement and is defined by NE to SW trends (folds, mineral lineations). The third event (F_3) displays diverse styles of NW to SE trending folds (nearly vertical axial planes) that affected the trend of the earlier F_1 and F_2 structures. The interpretation of a wrenching tectonic regime allows an explanation for the polydeformational style of the Tandilia basement, as recorded by the F_2 event. The Transamazonian Orogeny is seen as the product of a continent-continent collision (Cortijo Suture), as suggested by the presence of the leucogranite, with belts of mylonite, the presence of rocks derived from an ocean floor protolith and piling up of the crust (Dalla Salda *et al.*, 1987; Teruggi *et al.*, 1988).

In summary, in the Tandil area of the Rio de la Plata Craton, there are two major belts of Precambrian rocks: the Transamazonian rocks (2.2 - 1.8 Ga), mainly of igneous and metamorphic character; and the Brasiliano rocks (900 - 500 Ma) mainly sedimentary (Punta Mogotes Metapelites and the Sierras Bayas Group (La Tinta Group). These Precambrian sequences are partially covered by an extensive blanket of quartzose Ordovician sediments with trace fossils including *Cruziana*. The Precambrian marine deposits have been correlated with the Nama Group of southwestern Africa, and taken together these may represent an assembly for southwestern Gondwana for late Precambrian times.

The Island of Martin Garcia (Martin Garcia Complex)

The island of Martin Garcia is situated in the beginnning of the Rio de la Plata, some 46 km N of the city of Buenos Aires (Fig. 2). In that region metamorphic rocks are exposed along

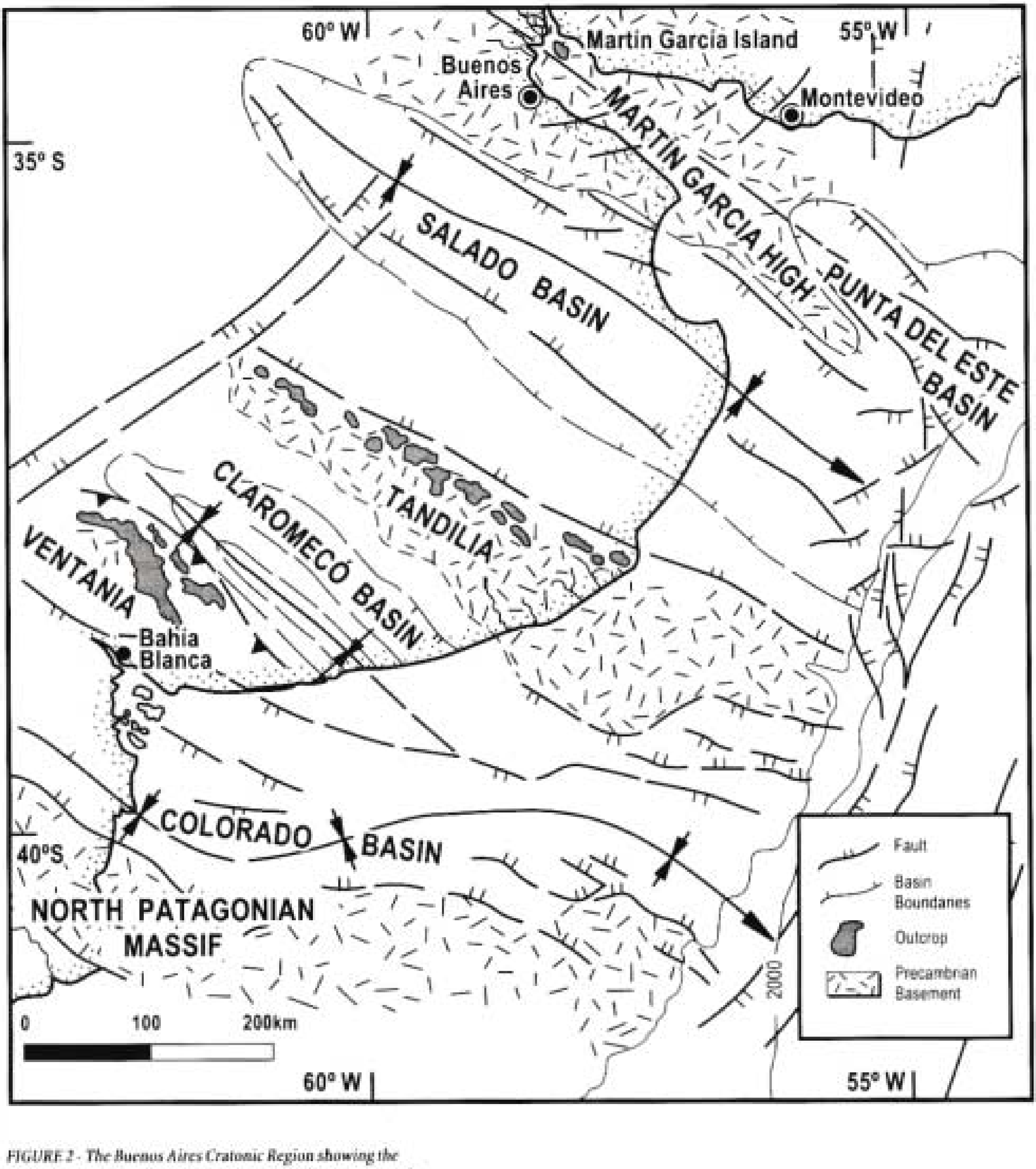


FIGURE 2 - The Buenos Aires Cratonic Region showing the Tandilia Ranges, Martin García Island, and the Ventania Belt, the rift - drift sedimentary basins, and main faults systems (modified after Ramos, 1996).

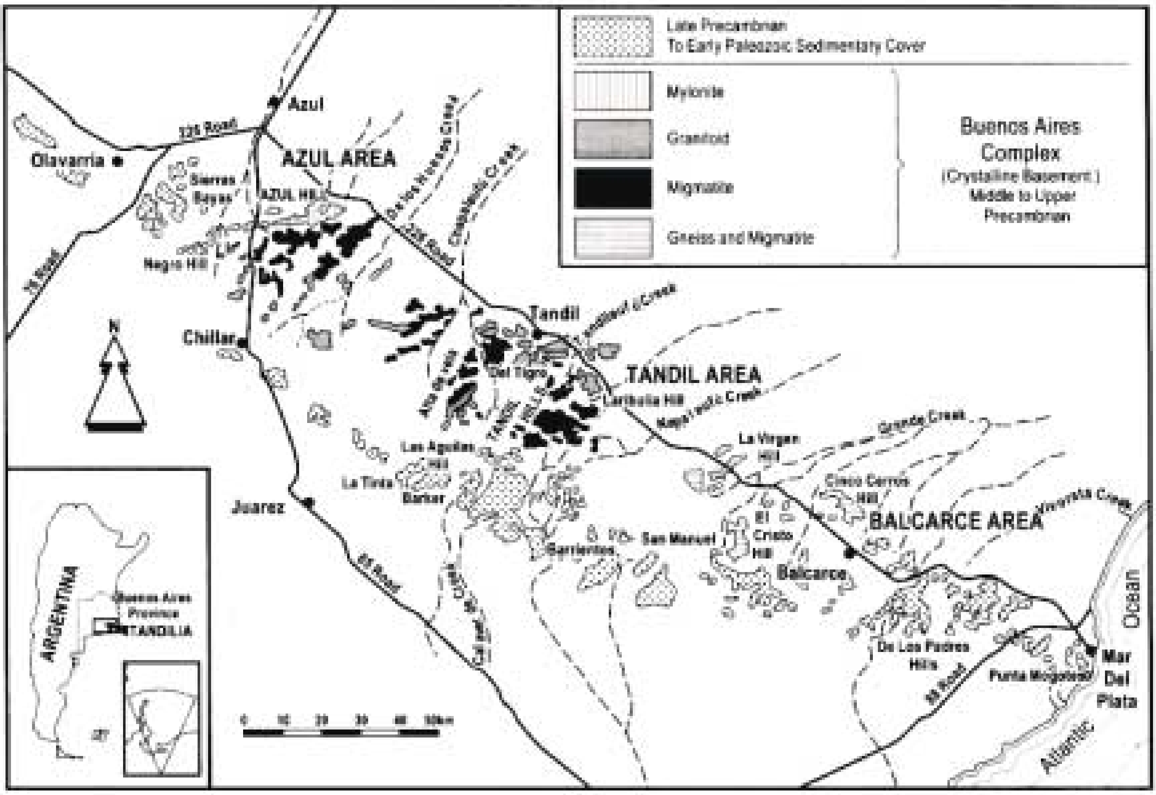


FIGURE 3 - Geology of Sierra de Tandilia showing lithological units of the Buenos Aires Complex, and the late Precambrian to Early Paleozoic sedimentary cover.

the southern, southwestern and southeastern beaches. Near the airport there is an outcrop of pyroxenite. Dalla Salda *et al.* (1988) considered these rocks as belonging to the Rio de la Plata Craton and named them the Martin Garcia Complex. The complex consists of amphibolite, gneiss and minor schist, cut by thin granitic dykes. The rocks often show a small degree of migmatization, and are lentiform due to polydeformation. Dalla Salda (1981a) assigned the Martin Garcia Complex to the Transamazonian Orogeny.

The rocks are interpreted to have been derived from a volcano-sedimentary protolith of basic composition, metamorphosed in the amphibolite facies. Staurolite-almandine and kyanite-almandine are the common metamorphic minerals. The protolith was first deformed into NW to SE trending folds dipping to the SE that may be related to the first episode of regional metamorphism, the grade of which was no lower than the greenschist facies. Isotope data suggest an age of 2.085 to 2.05 Ga for this first metamorphic event. Recent studies suggest that the rocks underwent a second phase of ductile deformation in which there developed an axial plane schistosity trending SW-NE. It has also been suggested that this second phase of deformation may have been related to the second metamorphic event (amphibolite grade, andesine) as the K-feldspar, plagioclase and quartz of the gneiss are synkinematic. K/Ar ages grouped around 1.87 to 1.6 Ga. Following the ductile deformation and two phases of regional metamorphism, the uplifted Martin Garcia Complex developed local strain-slip cleavage together with retrograde metamorphism. This late event was coeval with the minor emplacement of dykes, granitic to tonalitic in composition.

The Precambrian and Lower Paleozoic sedimentary rocks of the Sierra Tandilia

The Buenos Aires Complex of the Sierra Tandilia was partially covered by a thin platform sequence during the latest Precambrian times, constituting the Sierras Bayas Group. The Sierras Bayas Group is also known as La Tinta Formation (Amos *et al.*, 1972), or the La Tinta Group (Dalla Salda and Iñiguez, 1979).

Late Precambrian Sedimentary Cycle

According to Iñigues *et al.* (1989), the Sierras Bayas Group represents a late Precambrian sedimentary cycle, integrated with the Villa Mónica, Cerro Largo and Loma Negra formations. The lowermost unit of the group, the Villa Mónica Formation of about 50 m thick, was defined as the Lower Quartzite or Dolostone (González Bonorino, 1954). Detailed sedimentological studies in the Olavarria-Sierras Bayas region (Fig. 3) suggest two rock units. The first is a unit consisting of quartzite and arkose, 36 m thick, and the second is a dolomite unit, locally with intercalated shale, also 36 m thick. The dolomite unit is biogenic and contains stromatolites. Paleocurrent analyses show that reefs developed according to the E-W and NE-SW orientations. In the Barker area (Fig. 3), local facies changes have given rise to the recognition of a unit known as the La Juanita Formation, defined by a quartzose cross-bedded sandstone and a dolomite bed with stromatolites some 10 m thick.

Paleontological studies by Poiré (1993) suggest that the stromatolites are of Riphean age. Rb/Sr geochronological dating of pellitic rocks within the Sierras Bayas Dolomite gave ages of 793 ± 32 Ma (Cingolani and Bonhomme, 1982). Based on these data, together with assessory palaeontological, paleogeographical (Dalla Salda *et al.*, 1988) and paleomagnetic data (Valencio *et al.*, 1980) the La Juanita Formation can be assigned to the upper Precambrian.

The Cerro Largo Formation is 75 m thick. It begins with a shale-sandstone unit some 15 m thick (Poiré, 1993) that unconformably overlies the Villa Mónica Formation. This shale-sandstone unit is followed by 22 m of quartzose sandstone, previously named the Upper Quartzites that contains trace fossils with a simple structure. The quartzose sandstone unit grades into some 38 m of quartz-rich shale and siltstone. Towards the Barker and San Manuel areas, the beds become sandier with zones of conglomerate and bioturbated beds. Sedimentary interpretation of the Cerro Largo Formation suggests a sea level rise before regression. Geochronology and palaeontology favour an upper Precambrian age for these sediments. Rb/Sr dating revealed 769 ± 32 Ma (Bonhomme and Cingolani, 1980).

The Loma Negra Formation is a limestone unit about 45 m thick (Borrello, 1966). It was mined for a century as a source of carbonate. It is exposed in the Sierras Bayas and Barker areas as dark grey, blue and reddish mudstone, locally with terrigenous fragments at the base and organic matter at the top. In the Barker area, where karst topography has developed locally, collapse breccia-like deposits can be observed (Villa Cacique and El Infierno quarries). These rocks show a final marine regression with karst development at the end. Based on its position in the sequence and the fossil algae content, this formation is assigned to the upper Precambrian.

In the Olavarria-Sierras Bayas region Iñiguez and Zalba (1974) defined the Cerro Negro Formation as green and reddish shale sequence with illite and chlorite some 170 m thick. This formation developed unconformably on the Loma Negra Limestone. Fine-grained sandstone and siltstone beds and pyroclastic units are intercalated with the shale partings.

In the Barker area, olivine green shale overlies an intraformational limestone breccia. At the base of this unit, Leanza and Hugo (1987) described a phosphatic layer (Phosphate Member) that may suggest the infilling of a flat basin on a paleosurface of the Loma Negra Formation during the marine regression. Radiometric data (Rb/Sr fine fraction, 723 ± 30 Ma) suggests an upper Precambrian age which is supported by the presence of achritarch fossils. Possible equivalents of the Cerro Negro Formation have been found in the Cuchilla de las Aguilas Hills (Zalba *et al.*, 1988), and as diamictite units of the Volcán Hill of the Balcarce area described by Spalletti and del Valle (1984).

Lower Paleozoic Sedimentary Cycle

The Balcarce Formation, representing the last sedimentary cycle in the region, is exposed along the southern margin of the Sierra Tandilia from the towns of Olavarria in the NW to Mar del Plata in the SE (Fig.3). However, the best exposures occur between the towns of Balcarce and Mar del Plata in the southeastern region. The Balcarce Formation lies unconformably on the Precambrian basement complex, and it has an average thickness of 75 to 90 m. A maximum thickness of 450 m was obtained in a borehole near Mar del Plata. The Balcarce Formation consists of quartzite, fine-grained quartz-rich conglomerate, and thin zones with kaolinite. Individual beds are 0.3 to 1.5 m thick. They are convolute, display cross-bedding, ripple marks and normal graded stratification.

Paleoenvironmental studies suggest that these beds were deposited under shoreline, near shore and tidal conditions with periodic sea-level fluctuations (Iñiguez *et al.*, 1989). Paleocurrent analysis by Del Valle (1990) shows simple modal and polymodal diagrams with current directions from the NW, SW and W. Some pyroclastic units at the base of the formation and exposed near Cerro del Coral Hills (Fig. 4) were affected by hydrothermal solutions (Dristas and Frisicale, 1987). The Balcarce Formation has trace fossils, mainly *Cruziana*, *Arthophycus* and *Rusophycus*. These beds, mainly quartzite, were designated the *Cruziana* Facies by Aceñolaza (1979), and assigned an Ordovician age by Borrello (1966) or Cambrian-Ordovician age by Del Valle (1987a, 1987b). The sedimentation predated a diabase sill dated at 490 to 396 Ma intruding kaolinitic shale in the Los Barrientos area to the E of the town of Balcarce. Recent studies on trace fossils by Seilacher (in preparation) collected from the Los Pinios Quarry near the town of Balcarce and from Cabo Corrientes near the town of Mar del Plata show new species of *Cruziana*, suggesting an Upper Ordovician to Lower Silurian age for this unit when compared to other localities in Gondwana.

Ventania Fold and Thrust Belt

In the southwestern part of the Buenos Aires Province (Fig. 2) the Sierra de la Ventana, also known as the Sierras Australes, forms a curved fold and thrust belt that strikes approximately NW-SE. The belt is about 180 km long and 60 km wide. The Ventania Fold and Thrust Belt forms part of the Samfrau Geosyncline (Du Toit, 1937), and holds a close relationship with the rocks of the Malvinas Archipelago, the Cape Fold Belt of the Republic of South Africa and the Ellsworth Mountains of Antarctica (von Gosen *et al.*, 1990). The structure and stratigraphy of the Ventania Belt was described by Schiller (1930), Harrington (1947, 1970), Suero (1972), Llambias and Prozzi (1975), and more recently by Selles Martinez (1989), von Gosen and Buggish (1989), von Goshen *et al.* (1991), Cobbold *et al.* (1986, 1991), and Rossello *et al.* (1997).

Igneous rocks

The rocks of the Ventania Fold and Thrust Belt represent a Paleozoic metasedimentary sequence. In the W this overlies the poorly exposed Precambrian basement consisting of granitoid intrusions and rhyolite flows. The Rb/Sr geochronological data for the oldest basement rocks from the Aguas Blancas, Las Lomitas and Pan de Azucar areas show the following ages: 678 ± 30 Ma, 613 ± 30 Ma, and 594 ± 10 Ma (Cingolani and Varela, 1975; Varela *et al.*, 1985). However, some granitoid plutons and meta-rhyolite

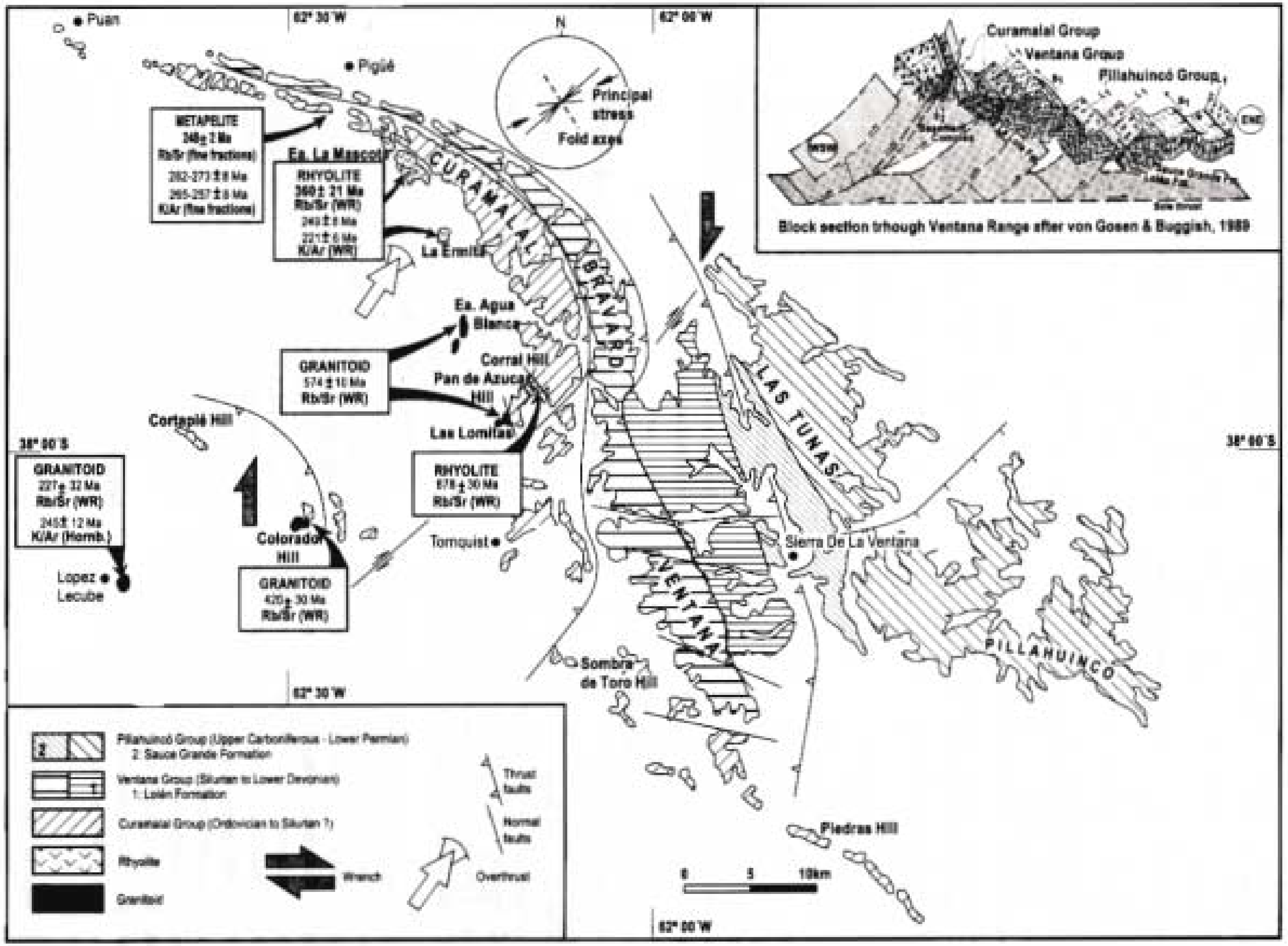


FIGURE 4 - The Ventania Fold and Thrust Belt. regional geology and structure. (tectonic interpretation after Cobbold et al., 1996; Rossello et al., 1997)

flows show Paleozoic ages from 420 ± 30 Ma at Colorado Hill to 360 ± 21 Ma at La Mascota - La Ermita. The post-tectonic Lopez Lecube syenitic granitoid showed a Rb/Sr age of 227 ± 32 Ma, and a hornblende, K/Ar age of 245 ± 12 Ma (Fig. 4).

The Paleozoic siliclastic units

Three groups of siliclastic rocks overlie the Precambrian basement. The oldest is the Curamalal Group, the base of which consists of quartzitic conglomerate of probable Ordovician and Silurian age. This is followed by the Ventania Group that displays coarse-grained quartzite units at the base, passing upwards into fine-grained quartzite with Silurian trace fossils *Daedalus* and *Arthrophycus*. At the top there lies a unit with feldspar-rich micaceous sandstone and black shale containing a Lower Devonian Malvinokaffric fauna known as the Lolen Formation.

At the top of the siliciclastic sequence there occurs the Pillahuincó Group that unconformably overlies the Ventania Group. At the base of the Pillahuincó Group there is sedimentological evidence for the Permo-Carboniferous glaciation of Southern Gondwana, represented by diamictite beds of the Sauce Grande Formation. The Sauce Grande Formation is overlain by typical Lower Permian Gondwana *Glossopteris* flora and *Eurydesma* fauna. The Piedra Azul, Bonete and Tunas formations are sandstone-rich units of Early Permain age that mark the end of the Paleozoic sedimentary succession, showing provenance from the belt, situated to the SW.

Tectonic evolution

The tectonic evolution of the Ventania Fold and Thrust Belt may have occurred as a single event during the Permian to Triassic. Alternatively, it developed during two events: the first from the Devonian to the Carboniferous, and the second during the Lower Permian. Schiller (1930) maintained that the deformation of the Ventania Belt occurred as the result of different thrusts. Later, Harrington (1947, 1970) suggested that deformation occurred by folding as the result of a single tectonic event. Ramos (1984), proposed a Patagonia-Gondwana collisional model and described the mountain range as a fold and thrust belt, having a sigmoidal form with evidence for progressive deformational diminishing towards the NE. Harrington (1970), demonstrated a decrease in deformational intensity and pointed to the very tight folds that become more concentric to the NE. Von Gosen and Buggish (1989) and von Gosen *et al.* (1990, 1991) proposed a fold and thrust model with two deformational events based on studies of the structure and metamorphic history. Cobbold *et al.* (1986, 1991) gave weight to ductile shearing and oblique convergence that included three different structural domains (Fig. 4). Japas (1989) proposed a model that involved sinistral shearing linked to a transpressive E-W sinistral system.

Selles Martinez (1989) suggested that the ranges show that the Patagonia-South American Craton Suture developed during sinistral shearing as the result of transpressive forces in the vicinity of vertical transcurrent faulting. Rossello *et al.* (1997) considered a transpressive oblique convergence model in which a set of dextral transcurrent faults resulted from overthrusting to the NE. Recently, Tomezzoli and Cristallini (1998) presented new evidence that may affect the interpretation of the tectonic evolution of the Ventania Belt.

During Permian times, Lower to Upper Paleozoic metasediments of the Ventania Belt, as well as the Precambrian basement were folded. According to some authors, including von Gosen and Buggish (1989), the basement rocks of the southwestern area were thrusted northeastwards in a series of imbricate structures during a first tectonic event to be followed by the development of shear and thrust zones and subsequent strike-slip faulting in response to sinistral transpression (Fig. 4).

Towards the E, the rocks of the Ventania Belt dip gradually under the Cenozoic sediments of the Claromecó Basin. Aeromagnetometric and gravimetric data (Kostadinoff, 1993) suggest that the basin deepens towards the E. However, some 80 km to the E, the Pillahuincó Group crops out as part of the Claromecó Foreland Basin.

The development of Permian metamorphism in the westernmost part of the Ventania Belt (Varela *et al.*, 1985; Buggish, 1987) can be correlated with a deformational phase of similar age. Illite crystallinity in the Ventania Belt increases to the W along with dynamo-metamorphic transformations. Metamorphic illite developed along cleavage planes have given K/Ar ages of 273 ± 8 Ma and 265 ± 32 Ma (Buggisch, 1987; von Gosen *et al.*, 1991); and between 282 ± 3 and 260 ± 3 Ma (Varela *et al.*, 1985; Buggish, 1987). The syntectonic Lopes Lecube Granitoid has been assigned K/Ar hornblende dates of 245 ± 12 Ma and Rb/Sr whole rock dates of 227 ± 32 Ma (Cingolani and Varela, 1975).

According to Rossello *et al.* (1997), the main deformational phase would have been contemporaneous with the deposition of the upper part of the Tunas Formation (Kungurian-Kazanian age). Tuff beds, interpreted as ash-fall, observed in this section (Iñiguez *et al.*, 1989) can be correlated along the active margin of Gondwana. The elongate shape suggests the influence of dextral transpressive tectonics developed on the distal side of the active margin (Cobbold *et al.*, 1986, 1991). The evidence points to an increase in tectonic activity during the deposition of the Pillahuincó Group, with growth folds (Cobbold *et al.*, 1991; Rossello *et al.*, 1993) and the syntectonic sedimentation of the Tunas Formation (Lopez Gamundí *et al.*, 1995).

REFERENCES

Aceñolaza, F.G. (1979). El Paleozoico inferior de Argentina según sus trazas fósiles. *Ameghiniana*, 15(1-2), 15-64.

Amos, A., Quartino, B., and Zardini, R. (1972). El Grupo La Tinta (Prov. Bs. Aires, Argentina), Paleozoico o Precámbrico? *Anais do XXV Congresso Brasileiro de Geologia*, São Paulo, Brasil, 3, 211-221.

Borrello, A.V. (1966). Trazas, restos tubiformes y cuerpos fósiles problemáticos de la Formación La Tinta, Sierras Septentrionales de Buenos Aires. *Paleontografía Bonaerense*, V, 1-42.

Bonhomme, M. and Cingolani, C. (1980). Mineralogía y geocronología Rb-Sr y K-Ar de fracciones finas de la Formación La Tinta, Provincia de Buenos Aires. *Revista de la Asociación Geológica Argentina*, 35(4), 519-530.

Buggish, W. (1987). Stratigraphy and very low grade metamorphism of the Sierras Australes de Provincia de Buenos Aires (Argentina) and implications in Gondwana correlation. *Zentralblatt für Mineralogie, Geologie und Paläontologie*, 1, 819-837.

Cingolani, C.A. and Bonhomme, M. (1982). Geochronology of La Tinta Upper Proterozoic sedimentary rocks, Argentina. *Precambrian Research*, **18**(1-2), 119-132.

Cingolani, C.A. and Varela, R. (1975). Examen geocronológico por el método Rb-Sr de las rocas ígneas de las Sierras Australes Bonaerenses. *Actas de lo V Congreso Geológico Argentino*, Córdoba, Argentina, 1, 349-371,.

Cobbold, P.R., Massabie, A., and Rossello, E. (1986). Hercynian wrenching and thrusting in the Sierras Australes Foldbelt, Argentina. *Hercynica*, 2(2), 135-148.

Cobbold, P.R., Gapais, D. and Rossello, E.A. (1991). Partitioning of transpressive motions within a sigmoidal foldbelt: the Variscan Sierras Australes, Argentina. *Journal of Structural Geology*, **13**(7), 743-758.

Dalla Salda, L.H. (1980). Some relationship between the cratonic areas of the Rio de la Plata, South America and the Western Kalahari, Africa. *17th Annual Report of the Chamber of Mines, Precambrian Research Unit, Cape Town, South Africa*, 74-77.

Dalla Salda, L.H. (1981a). El basamento de la Isla Martín García, Río de la Plata. *Revista de la Asociación Geológica Argentina*, **26**(1), 29-43.

Dalla Salda, L.H. (1981b). Tandilia, un ejemplo de tectónica de transcurrencia de basamento. *Revista de la Asociación Geológica Argentina*, **36**(1), 204-207.

Dalla Salda, L.H. (1982). Nama-La Tinta y el inicio del Gondwana. *Acta Geológica Lilloana*, **16**(1), 23-38.

Dalla Salda, L.H. and Iñiguez, A.M. (1979). La Tinta, Precámbrico y Paleozoico de Buenos Aires. *Actas de lo VII Congreso Geológico Argentino, Neuquén, Argentina*, **1**, 539-550.

Dalla Salda, L., Franzese, J., and Posadas, V.B. (1987). The 1800 Ma mylonite-anatectic granitoid association in Tandilia, Argentina. *Abstracts Volume of the 7th International Conference on Basement Tectonics*, Kingston, Jamaica, p. 28.

Dalla Salda, L.H., Bossi, J., and Cingolani, C.A. (1988). The Rio de La Plata cratonic region of Southwestern Gondwanaland. *Episodes*, **11**(4), 263-269.

Del Valle, A. (1987a). Nuevas trazas fósiles en la Formación Balcarce, Paleozoico inferior de las Sierras Septentrionales. Su significado cronológico y ambiental. *Revista Museo La Plata*, **9**, 19-41.

Del Valle, A. (1987b). *Sedimentología de la Formación Balcarce en el sector oriental de Tandilia*. Tesis Facultad de Ciencias Naturales y Museo. Univ. Nac. La Plata (unpublished).

Del Valle, A. (1990). Paleogeografía de la cuenca de depositación de la Formación Balcarce (Ordovícico inferior), Tandilia, Argentina. *Actas de la III Reunión Argentina de Sedimentología, San Juan, Argentina*, 102-108.

Di Paola, E. y Marchese, M. (1974). Relaciones entre la tectosedimentación, litología y mineralogía de arcillas del Complejo Buenos Aires y la Formación La Tinta. *Revista de la Asociación Argentina Mineralogía, Petrología y Sedimentología*, **5**, 3-4.

Dristas, J. and Frisicale, M.C. (1987). Rocas piroclásticas en el sector suroeste de las Sierras Septentrionales de la Provincia de Buenos Aires. *Revista de la Asociación Argentina Mineralogía, Petrología y Sedimentología*, **18**(1-4), 33-45.

Du Toit, A. (1937). *Our Wandering Continents, an Hypothesis of Continental Drift*. Oliver and Boyd, Edinburgh, Scotland, 366 p.

Gonzalez Bonorino, F. (1954). Geología de las Sierras Bayas, Partido de Olavarría, Provincia de Bs. Aires. *Revista MOP. LEMIT, La Plata, Argentina*, **55**, Serie II, 5-37.

González Bonorino, F., Zardini, R., Figueroa, M., and Limousin, T. (1956). Estudio geológico de las Sierras de Olavarría y Azul (Provincia de Buenos Aires). *Revista MOP LEMIT*, La Plata, Argentina, **63**, Serie II, 5-22.

Harrington, H.J. (1947). Explicación de las Hojas Geológicas 33 m (Sierra de Curamalal) y 34 m (Sierra de la Ventana), provincia de Buenos Aires. *Dirección de Minería y Geología, Buenos Aires, Argentina, Boletín*, **61**, 56 p.

Harrington, H.J. (1970). Las Sierras Australes de Buenos Aires, República Argentina: cadena aulacogénica. *Revista de la Asociación Geológica Argentina*, 25(2), 151-181.

Iñiguez, A.M., Del Valle, A., Poiré, D., Spalletti, L., and Zalba, P. (1989). Cuenca Precámbrica-Paleozoica inferior de Tandilia, Provincia de Buenos Aires. In *Cuencas sedimentarias argentinas*, eds Chebli, G. and Spalletti, L.A., pp. 245-263. Universidad Nacional de Tucumán, Serie Correlación Geológica 6, Tucumán, Argentina.

Iñiguez, A.M. and Zalba, P. (1974). Nuevo nivel de arcilitas en la zona de Cerro Negro, Partido de Olavarría, Prov. Buenos Aires. *Revista MOP LEMIT*, La Plata, Argentina, **264**, Serie II, 95-100.

Japas, S. (1989). La deformación de la cadena plegada de las Sierras Australes de Buenos Aires. *Academia Nacional de Ciencias Exactas, Físicas y Naturales (Buenos Aires), Anales*, **40**, 193-215.

Kostadinoff, J. (1993). Geophysical evidence of a Paleozoic basin in the interhilly area of Buenos Aires province, Argentina. *Comptes Rendus 12 ICC-P (Buenos Aires)*, I, 397-404.

Leanza, H., and Hugo, C. (1987). Descubrimiento de fosforitas sedimentarias en el Proterozoico superior de Tandilia, Buenos Aires, Argentina. *Revista Asociación Geológica Argentina*, Buenos Aires, **42**(3-4), 417-428.

Lema, H., and Cucchi, R. (1985). Geología de la Sierra Alta de Vela. *Primeras Jornadas Geológicas Bonaerenses, Actas*, Tandil, 867-879.

Lopez Gamundi, O.R., Conaghan, P., Rossello, E.A. and Cobbold, P.R. (1995). The Tunas Formation (Permian) in the Sierras Australes Foldbelt, East-Central Argentina: evidence of syntectonic sedimentation in a Variscan foreland basin. *Journal of South American Earth Sciences*, **8**(2), 129-142.

Llambías, E.J. and Prozzi, C. (1975). Ventania. *VI Congreso Geológico Argentino (Bahia Blanca), Relatorio*, 78-101.

Marchese, H.G. and Di Paola, E. (1975). Reinterpretación estratigráfica de la Perforación de Punta Mogotes I, Provincia de Buenos Aires. *Asociación Geológica Argentina, Revista*, **XXX**(1), 44-52.

Poiré, D. (1993). Estratigrafía del Precámbrico sedimentario de Olavarría, Sierras Bayas, Provincia de Buenos Aires. Argentina. *XII Congreso Geológico Argentino, y II Congreso Exploración Hidrocarburos, Mendoza, Actas*, **2**, 1-11.

Ramos, V.A. (1984). Patagonia: Un continente a la deriva? *IX Congreso Geológico Argentino (Bariloche), Actas*, **2**, 311-325.

Ramos V. (1988). Late Proterozoic-Early Paleozoic of South America - a Collisional History. *Episodes*, **11**(3), 168-174.

Rossello, E.A., Cobbold, P.R. and Lopez Gamundi, O.R. (1993). Pliegues de crecimiento en la Formación Tunas (Zoilo Peralta, Sierras Australes de Buenos Aires, Argentina). *Proyecto Pangea y Glopals (San Juan). Comunicaciones*, 52-55.

Rossello, E.A., Massabie, A.C., Lopez Gamundi, O.R., Cobbold, P.R. and Gapais, D. (1997). Late Paleozoic transpression in Buenos Aires and northeast Patagonia ranges, Argentina. *Journal of South American Earth Sciences*, **10**(5/6), 389-402.

Schiller, W. (1930). Investigaciones geológicas en las montañas del Sudoeste de la Provincia de Buenos Aires. Anales Museo de La Plata. *Sección Mineralogía y Geología. Segunda Serie*, IV, La Plata, 101 p.

Selles Martinez, J. (1989). The structure of the Sierras Australes (Buenos Aires province, Argentina): an example of folding in a transpressive environment. *Journal of South American Earth Sciences*, **2**(4), 317-329.

Spalletti, L. and Del Valle, A. (1984). Las diamictitas del sector oriental de

Tandilia: caracteres sedimentológicos y origen. Asociación Geológica Argentina, Revista, **XXXIX**(3-4), 188-206.

Suero, T. (1972). Compilación geológica de las Sierras Australes de la provincia de Buenos Aires. *LEMIT (La Plata), Anales*, **3**, 135-147.

Teruggi, M., Kilmurray, J. and Dalla Salda, L. (1973). Los dominios tectónicos de la región de Tandil. *Anales Sociedad Científica Argentina*, CCXCV(1/2), 81-94.

Teruggi, M., Kilmurray, J., and Dalla Salda, L. (1974). Los dominios tectónicos de la región de Balcarce. *Asociación Geológica Argentina, Revista*, **XXIX**(3), 265-276. Buenos Aires.

Teruggi, M., Kilmurray, J., Rapela, C. and Dalla Salda, L. (1974). *Diques básicos en las Sierras de Tandil. Asociación Geológica Argentina, Revista*, **XXIX**(1), 41-60.

Teruggi, M., Leguizamón, M. and Ramos, V. (1988). Metamorfitas de bajo grado con afinidades oceánicas en el basamento de Tandil: su implicancia geotectónica, Provincia de Buenos Aires. *Asociación Geológica Argentina, Revista*, **XLIII**(3), 366-374.

Teruggi, M. E. and Kilmurray, J. O. (1980). Sierras Septentrionales de la Provincia de Buenos Aires. In: *Geología Regional Argentina*. Academia Nacional de Ciencias, Córdoba, **II**, 919-956.

Tomezzoli, R.N. and Cristallini, E.O. (1998). Nuevas evidencias sobre la importancia del fallamiento en al estructura de las Sierras Australes de la Provincia de Buenos Aires. *Asociación Geológica Argentina, Revista*, 53(1), 117-129.

Valencio, D.A., Sinito, A.M., and Vilas, J.F. (1980). Paleomagnetism of upper Precambrian rocks of the La Tinta Formation, Argentina. *Geophysical Journal R. Astrophysics, Society*, **62**, 563-575.

Varela, R., Cingolani, C. and Dalla Salda, L. (1988). Geocronología Rb-Sr en granitoides del basamento de Tandil, Prov. De Buenos Aires, Argentina. *Segundas Jornadas Geológicas Bonaerenses, Actas*, Bahia Blanca, 291-305.

Varela, R., Dalla Salda, L.H. and Cingolani, C.A. (1985). Estructura y composición geológica de las sierras Colorada, Chasicó y Cortapié, Sierras Australes de Buenos Aires. *Asociación Geológica Argentina, Revista* (Buenos Aires), **40**(3/4), 254-261.

Von Gosen, W. and Buggish, W. (1989). Tectonic evolution of the *Sierras Australes fold and thrust belt (Buenos Aires province*, Argentina)-an outline. *Zentlb. Miner. Geol. Paläont.*, 1, 947-958.

Von Gosen, W., Buggisch, W. and Dimieri, L. (1990). Structural and metamorphic evolution of Sierras Australes (Buenos Aires Province, Argentina). *Geologische Rundschau*, **79**(3), 797-821.

Von Gosen, W., Buggish, W. and Krumm, S. (1991). Metamorphism and deformation mechanisms in the Sierras Australes fold and thrust belt (Buenos Aires province, Argentina). *Tectonophysics*, **185**, 335-356.

Zalba, P.E., Andreis, R. R. and Iñiguez, A.M. (1988). Formación Las Aguilas, Barker, Sierras Septentrionales de Buenos Aires, nueva propuesta estratigráfica. *Asociación Geológica Argentina, Revista*, **XLIII**(2), 198-209.

THE BRASILIANO OROGENIC CYCLE

"Since the beginning of the Neoproterozoic break-up and drift controlled by taphrogeny has driven plate kinematics toward orogenic interactions. These processes continued for at least 500 Ma until the Early Ordovician, following all the steps of the classical Wilson Cycle. An interesting characteristic of these orogenies is the diachronism of events, many of which were coeval with taphrogenic processes elsewhere. Many orogenies were controlled by different kinds of plate interaction occurring diachronically in different places.
This plurality of processes that converged to the closure of a wide oceanic space populated by small continental fragments (terranes or microplates) may be described in terms of orogenic systems or branching systems of orogens, rather than the general and geometry-related mobile belt model."

(Campos Neto, this volume)

Tectonic History of the Borborema Province, Northeastern Brazil

B.B. de Brito Neves, E. J. dos Santos, and W. R. Van Schmus

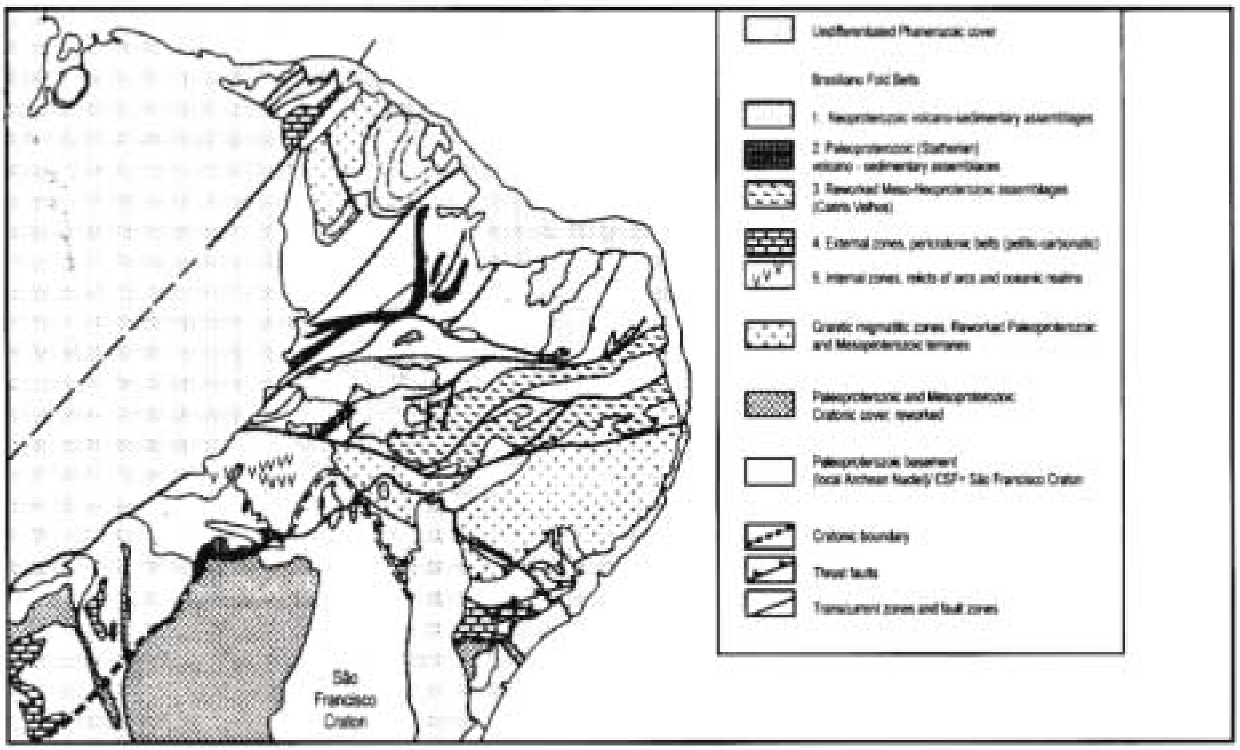
The concept of a Borborema Province was introduced by Almeida *et al.* (1977) and applied to the eastern part of the northeastern region of the South American Platform (Fig. 1). They defined it as a "complex mosaic-like folded region" where there were effective and important tectonic, thermal, and magmatic events of Neoproterozoic age assigned to the Brasiliano Cycle. Since the last century this structural province has been the subject of much geological research, and many papers, especially those concerning mineral resources (sheelite and pegmatite minerals) have been written. The area covered by this province exceeds 450 000 km^2. It consists of successive Cenozoic pediplanes, developed at progressively higher elevations from the coastal regions inland, reaching elevations of 1100 m. Taken as a whole, these pediplanes are referred to as the Serra da Borborema.

The structures of this tectonic region extend beyond the formal limits originally defined by Almeida *et al.* (1977, 1981). For additional reasons, they exceed the geographical limits of the Serra da Borborema itself. To the W and SW the province includes part of the basement of the Parnaíba Basin of Phanerozoic age. To the N and NE, structures associated with this province have been detected in the basement of the Coastal Province and the Atlantic continental margin. Furthermore, in the African counterpart of the Borborema Province (in the area between Togo to the N, and Gabon to the S) there are litho-structural and tectonic continuations of the Borborema Province, including domains of the Trans-Saharan (Pharusian, Dahomeyan), Nigerian, and Oubanguides-Central Africa mobile belts.

This Borborema Province and its African extensions are situated to the N of the São Francisco-Congo Craton, which behaved as a foreland domain during their development. There are several different degrees of structural and tectonic transition between the main Neoproterozoic orogenic system of the Borborema Province and the cratonic foreland domain to the S. These features, in which transitional tectonic domains predominate, have not always been explicitly described as formal tectonic components of the province in published syntheses. It is also possible to observe from both fieldwork and regional maps that some litho-structural components of the basement in the São Francisco Craton continue northwards as reworked basement of the Borborema Province. The same may be expected for relationships with the bounding forelands to the N, which are usually presumed to be the São Luís-West Africa Craton even though the transitional region is mostly buried beneath sediments of the Parnaíba and other coastal basins.

The lithostructural and tectonic framework of the province is well defined. The structures and rock-types

FIGURE 1 - Sketch map of the Borborema Province, north of the São Francisco Craton.



developed mainly during the evolution of two different tectonic cycles: the late Mesoproterozoic to early Neoproterozoic Cariris Velhos Orogeny, and the late Neoproterozoic Brasiliano Orogeny, bearing in mind that the Brasiliano Orogeny inherited structural trends and reworked crust formed during the earlier Cariris Velhos Orogeny. Additionally, in the basement of the province there are other tectonic features in outcrops of basement blocks that can be attributed to important stages of continental amalgamation during the middle Paleoproterozoic Transamazonian orogeneses (collage), and also during the Archean.

The concept of a Brasiliano structural province, based on the importance of this Neoproterozoic-Cambrian Cycle that was part of the process of amalgamation of Western Gondwana is a valid one. This is due to the supremacy of the Brasiliano structures (and their tectonic overprint) as well as on account of the remarkable granitic plutonism associated with these structures. This is particularly relevant in the general distribution of the main structural trends of the province, which display a fan-like pattern for the whole northeastern part of the South American Continent, under the influence of the major tectonic lineaments and their displacements in the Cambrian (fini-Brasiliano) extensional phase. Thus all structural trends extend diagonally to the coast. Last but not least, the bulk of geochronological data supports the importance of the Brasiliano/Neoproterozoic evolution.

For several years during which geological research was carried out in this province, the importance and general features of Brasiliano events as a whole obscured the features formed during the preceding Cariris Velhos Orogeny and Paleoproterozoic (Transamazonian) collage. Only during this decade, with the introduction of U/Pb and Sm/Nd methods of isotope analysis, has it been possible to define the Cariris Velhos Cycle and to increase the knowledge on the nature and extent of Paleoproterozoic (Transamazonian) rock assemblages in the basement rocks of the Borborema Province. It is now possible to see that, collectively, these two older tectonic cycles were much more important in terms of crustal formation throughout the province than the Brasiliano Orogeny.

For an understanding of the province (general shape and geographical features) it is necessary to draw attention firstly for the Phanerozoic cover; secondarily to the phase of transition (Cambrian escape tectonics, important strike-slip displacements); thirdly to the phase of stabilization (full cratonic conditions, from the Ordovician to the Middle Jurassic, including the development of syneclises); and finally to the so-called phase of activation (post-Jurassic to Late Cretaceous), when interior rifts and the Atlantic coastal province were formed under extensional conditions. These events were coeval with the phases of breakup of Pangea, the formation of the South Atlantic Ocean, and the final individualization of the South American Plate. Parallel to and subsequent to the breakup of Pangea and the formation of the South Atlantic Ocean, the continental interior underwent erosion and uplift (phase of re-stabilization, post-Late Cretaceous). Many of the sedimentary deposits formed were related to the structural trends of the Brasiliano Cycle. From that time onwards, tectonism and magmatism had been strongly attenuated. The Phanerozoic sedimentary cover of the Borborema Province includes several basins (syneclises, parts of syneclises, rifts), which are testimony of and representatives of six different and successive cratonic sedimentary sequences formed during the Phanerozoic evolution of the South American Platform. The locations, shapes and other general structural features of all these basins were strongly influenced by the structures of their underlying basement and, particularly by those trends developed during the fini-Brasiliano tectonism.

Previous Studies

A complete inventory of the papers that have covered the evolution of knowledge of the Borborema Province over the last 170 years is a hard task and beyond the scope of this review. To some extent this has been done by Santos and Brito Neves (1984) and Santos *et al.* (1984), and more recently by Jardim de Sá (1994). In the interests of synthesis and for pragmatic reasons, papers published in this decade will preferentially be referred to in this review, considering that they bring the essence of current knowledge, and they cite many of the works published previously.

Special mention is made of the papers by Santos and Brito Neves (1993); Campos Neto *et al.* (1994); Van Schmus *et al.* (1995a, 1995b); Brito Neves *et al.* (1995b, where the designation of Cariris Velho Orogeny was formalized); Kozuch *et al.* (1997); Dantas *et al.* (1998); and Santos and Medeiros (1998). These papers are products and by-products of a series of M.Sc. and Ph.D. dissertations, and research projects by students and professors attached to the Universidade de São Paulo (USP), the Universidade Estadual Paulista (UNESP), the Universidade Federal de Pernambuco (UFPE), the Universidade Rural de Pernambuco (UFRPE), the Universidade Federal do Ceará (UFCE), and the University of Kansas (KU) at Lawrence, Kansas, U.S.A. These programs had the partial or substantial financial support of CNPq, FAPESP, and the U.S. National Science Foundation, and they resulted from agreements and collaboration that started in 1989. These research programmes were carried out in consultation with many other geologists working in the region, and their assistance and comments are gratefully acknowledged. In addition, the important help of the Geological Survey of Brazil - CPRM, and the 1:100 000 and 1:250 000 scale sheets and maps of the P.L.G.B. (Basic Program of Geological Mapping of Brazil), should also be mentioned.

Naturally, data from all these theses and dissertations, as well as from other sources of information will be included in this review. Finally, this synthesis is based, in part, on a large amount of still unpublished data (including on-going theses and research). Much of these data are presented or discussed here for the first time. Nevertheless, they still remain the property of the aforementioned research programmes.

Major structural domains

By building upon previous studies and using the results of geological and isotope research in progress, it is possible to recognize five major contiguous tectonic domains in the

Borborema Province. There is naturally a significant degree of subjective reasoning for our designations, and these subdivisions should be seen as an exercise that must be reviewed using the results from additional geological, geochemical, and geophysical research. There is good geological evidence that favours the view that these major crustal segments or domains were arranged in their present configuration before the end of the Cambrian, following a significant phase of fini-Brasiliano strike-slip displacement (extensional tectonics, 545 to 500 Ma). During the rest of the Phanerozoic the boundaries of these crustal segments may have been disturbed slightly, but not enough to modify the earlier features of the main domains (Fig. 2).

In spite of the fact that most of the domain boundaries are lineaments or shear zones, this was not the fundamental basis for our subdivisions. Even though the lineaments chosen have most or all the characteristics necessary on which to define the limits of terranes or super terranes (sense of Howell, 1995), we prefer to use the informal designation of domains. We chose this approach with a view to maintaining maximum flexibility in defining terranes and super terranes for the future since there are many shear zones and faults within our designated domains that may also be terrane boundaries. Only for the subdivision of the Transversal Zone (TZ Domain), the informal designation of terranes will be used, following a previous and useful subdivision of Santos (1996), a usual practice followed by those working in the Borborema Province.

The Médio Coreaú Domain (MCO)

This domain in the northwestern part of the State of Ceará and the northeastern part of the State of Piauí, is situated between the reworked margin of São Luis-West Africa Craton and the Transbrasiliano/Kandi Lineament (figs. 1 and 2). The Médio Coreaú Domain consists of basement with juvenile 2.35 Ga (pre-Transamazonian) high-grade metamorphic rocks and entrapped segments of Neoproterozoic cratonic volcano-sedimentary (Martinópole Group) and pelitic-carbonate fold belts (Ubajara Group) that may be disrupted parts of the major trans-Saharan Mobile Belt. Along the Transbrasiliano Lineament there occur a series of transtensional basins and post-orogenic plutons occupying pull-apart spaces; many of these are covered by the Phanerozoic sedimentary rocks of the Parnaíba Basin of (Gusmão, 1998).

The Central Ceará Domain (CC)

This domain is situated between the Transbrasiliano-Kandi Lineament and the Senador Pompeu-Ile Ife Lineament (Figs. 1 and 2). In addition to the striking prominence of these boundaries (which can be extrapolated well beyond the formal limits of the Borborema Province), the Central Ceará Domain consists of gneissic basement formed during the Transamazonian collage, with the inclusion of an important Archean nucleus (Tróia-Tauá Massif).

The 2.2 to 2.1 Ga (U/Pb in zircon, some Rb/Sr isochrons) Transamazonian basement is distinguished by nearly juvenile Sm/Nd crustal signatures (T_{DM} = 2.4 to 2.3 Ga) for their high grade rocks. This domain also contains a series of middle Neoproterozoic supracrustal sequences or remnants of fold belts (quartzite, pelite, minor carbonate units) and expressive Brasiliano (late Neoproterozoic) plutonism. One of these granitic-migmatitic plutonic complexes, the NE trending Santa Quitéria Massif in the northwestern part of this domain, displays a series of isotope and geophysical characteristics that, although preliminarly, suggest that it is a Brasiliano continental magmatic arc (Fetter, 1999). Along the northern margin of the Archean Tróia-Tauá Massif, in contact with the Guaramiranga-Canindé Fold Belt, there is a unique occurrence of blue schist (Rio Canindé area) that could represent a collisional suture.

The separation of this domain from the adjacent Rio Grande do Norte Domain is done with caution at the present stage of knowledge. It is possible that the Sm/Nd isotope contrast across the Senador Pompeu-Ile Ife Lineament (Fetter, 1999) is actually gradual and these two domains have always been part of a single, large Paleoproterozoic Transamazonian crustal block (a tectono-stratigraphic terrane, according to Howell, 1995)

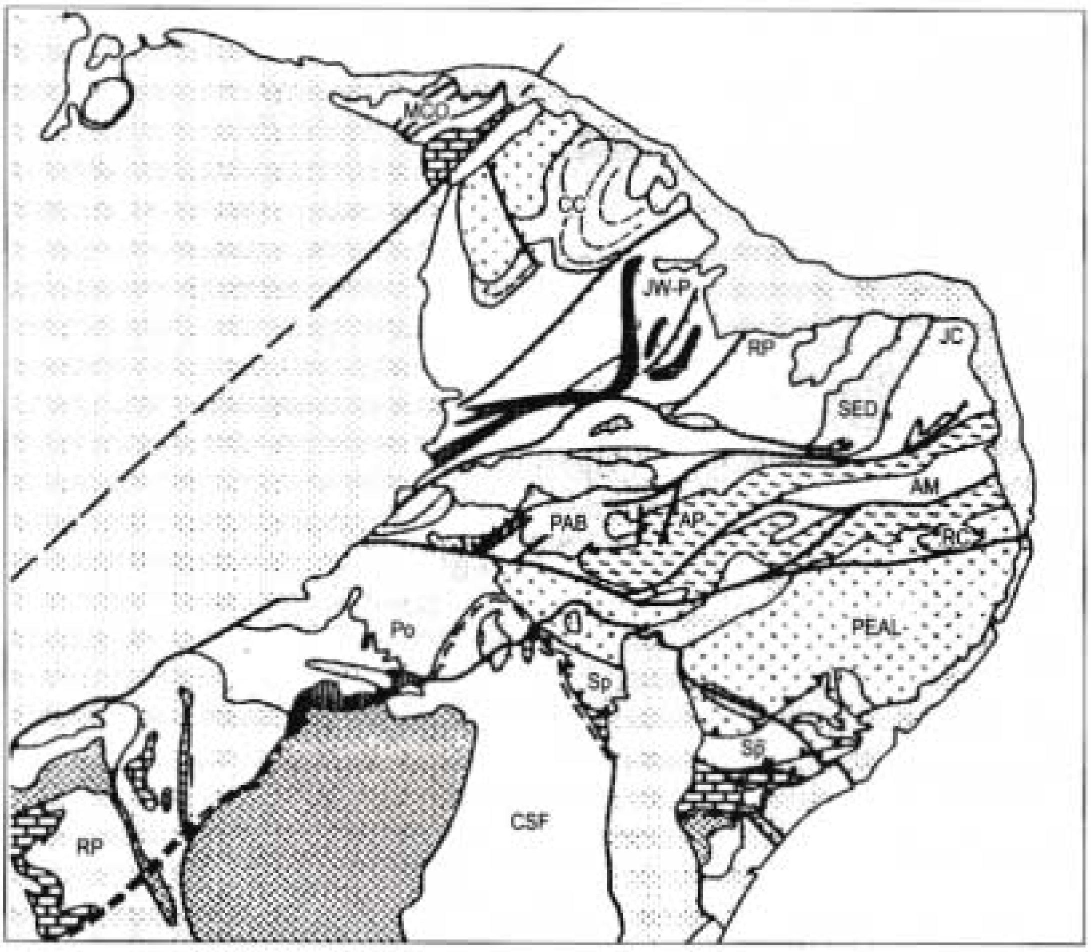
The Rio Grande do Norte Domain (RGND)

This domain is situated between the NE-SW trending Senador Pompeu-Ile Ife Lineament on the W, and the E-W trending Patos Lineament to the S. To the N and E its extension is limited by the Atlantic Ocean. The Rio Grande do Norte Domain (Fig. 2) includes several smaller tectonic zones or sub-domains, from W to E these are the Jaguaribeano-Encanto Fold Belt (J-WP) and its basement (between the Senador Pompeu and Portalegre lineaments), the Rio Piranhas Massif (RP), the Seridó Fold Belt (SED, the classical area of the province) and its basement, and the São José do Campestre Massif (JC).

The basement of all these massifs (RP, JC) and fold belts (J-WP) of the RGND comprise a major framework of the Transamazonian tectonic collage, which includes some local Archean nuclei. Throughout the RGND the Transamazonian basement gneiss yield late Archean T_{DM} ages (2.6 to 2.5 Ga; Van Schmus *et al.*, 1995a, b; Dantas, 1997; Dantas *et al.*, 1998), in contrast to the younger T_{DM} ages (Paleoproterozoic) found in the Central Ceará Domain to the W.

Upon the predominant Transamazonian basement, to the W of the RGND, there occur the remnants of schist belts (Jaguaribe, Orós, Peixe Gordo, Encanto, equivalent to the J-WP) with late Paleoproterozoic (1.8 to 1.7 Ga) volcano-sedimentary sequences. Neoproterozoic metasedimentary belts occur to the S (Caipu, Lavras, Mangabeira, Iara-Quimami), and the main Seridó Belt is in the central part of the domain (with possible extensions to the E).

The geophysical and geological data indicate that the RGND acted mostly as a monolithic block since the end of the Transamazonian Orogeny. No internal sutures have been found yet, and in a pre-drift (Pangea or Gondwana) reconstruction we suspect that this domain continues into the southern part of Nigeria. The Central Ceará and RGN domains together could have been a fragment of the Atlantica Supercontinent of Rogers (1996) that docked into its present position between the Patos and Sobral lineaments



MCO		Medio Coreau
CC		Central Ceará
RGND	JW-P = Jaguaribe-W potiguar RP = Rio Piranhas SED = Seridó Belt JC = São José Campestre	R. Grande do Norte
TZ	PBA = Piancó-Alto Brígida AP = Alto Pajeu AM = Alto Moxotó RC = Rio Capibaribe	Transversal Zone
SD	Rp = Rio Preto Po = Riacho do Pontal Sp = Sergipano	Southern Domain
PEAL	= Pernambuco-Alagoas Massif	
CSF		São Francisco Craton

FIGURE 2 - The main geological domains of the Borborema Province (MCO, CC, RGND, TZ, SD and PEAL). These domains may informally be subdivided into terranes, based on their main litho-structural and geochronological characteristics.

during the Brasiliano Orogeny, thus constituting a true tectono-stratigraphic terrane that has until now been overlooked in all the main reconstructions of paleo-supercontinents (Weil *et al.*, 1998; Unrug, 1996).

The central domain or Transversal Zone Domain - TZ

This domain is situated between the Patos and Pernambuco lineaments (shear zones) and displays general ENE-WSW to E-W structural trend (Figs.4 and 6). The name Transversal Zone Domain will be used since it gives a more graphic description relevant to the prevailing structures within the domain. In particular, the domain contains many internal fault blocks that have been deformed and rotated clockwise as a result of the dextral shear couple between the right-lateral shearing of the Patos and Pernambuco lineaments. This domain is also the type area for the definition of the *c.* 1.0 Ga to 950 Ma Cariris Velhos Orogen (Brito Neves *et al.*, 1995a, b; Van Schmus *et al.*, 1995a, b; Kozuch *et al.*, 1997).

The TZ contains several regions or lithotectonic elements previously referred by various names, including the Pajeú-Paraíba and Piancó-Alto Brígida fold belts of Brito Neves (1983), the latter also called the Cachoeirinha-Salgueiro Fold Belt by Sial (1987). Santos (1996) and Santos *et al.* (1997) have proposed three main parallel terranes for the area of the eastern and central area of the TZ domain (dismissing the designation of the Pajeú-Paraíba Fold Belt). These are the Alto Pajeú Terrane - APT (with most of the supracrustal rocks and orthogneiss), Alto Moxotó terrane - AMT (with a predominance of exposures of the reworked Transamazonian basement and very few Brasiliano plutons), and the Rio Capibaribe Terrane - RCT (with different supracrustal sequences, Neoproterozoic and older ones and large Brasiliano plutons).

This subdivision will be followed in this text because it seems practical and useful, but without advancing on the usual arguments about the uses and concepts of the term terrane. Recent studies and new data have shown that the geographical area previously represented by the Piancó-Alto Brígida Fold Belt (in different works and maps) is structurally, lithologically, and temporally not a single entity. There is a southern and narrow belt known as the Riacho Gravatá, situated to the S of the Serra do Caboclo Shear Zone, that was formed with the Cariris Velhos Orogeny, and so it should be added to AP.

Adjacent to and NW of the AP there is a late Neoproterozoic fold belt that contains low-grade psammitic and pelitic rhythmite beds with a few intercalations of mafic to felsic volcanic rocks; for which original name of the Piancó-Alto Brígida Fold Belt - PAB, will be maintained, with restrictions on the areal extent. The PAB originally may have had stratigraphic continuity with some of the Neoproterozoic supracrustal sequences (Rio Preto, Sergipana) of the Southern Domain. The PAB is also noted for the abundant Brasiliano plutonism that it contains, in the form of numerous stocks and plutons.

The continuation of the TZ to the African side of West Gondwana, between the Adamaoua and Garoua shear zones in Cameroun, has been suggested by a number of workers including Trompette (1994). However, this matter requires further investigation.

The Southern Domain (SD)

This domain is situated between the Pernambuco Lineament and the northern periphery of the São Francisco-Congo Craton (SFC). It is composed of continental margin and distal parts of the Rio Preto (Rp) and Riacho do Pontal (Po) fold belts (Fig. 2) to the NW and N of the São Francisco Craton, and the Sergipano Fold Belt (Sp) to the N and NE of the São Francisco Craton. This domain extends eastwards to Africa as the Central African-Oubanguides Fold Belt to the N of the Congo Craton. The Southern Domain includes parts of the cratonic foreland and hinterland of these fold belts (Rp, Po, Sp), which now consist of reworked high grade metamorphic rocks that define the original extent of the cratonic basement.

The Neoproterozoic supracrustal rocks of continental margin affinity in the southern parts of the Sergipano and Rio Preto fold belts, have similar stratigraphic sequences (diamictite, quartzite, pelite, carbonate) that are characteristic of undeformed or slightly deformed cratonic cover (the São Francisco-Bambuí Supergroup). The more distal sequences of the Sergipano and Rio Preto fold belts include a variety of volcano-sedimentary and plutonic rock assemblages, including deeper marine sediments and some suggestions for the presence of Neoproterozoic magmatic arcs. All these aspects are presently the subjects of ongoing research.

The Pernambuco-Alagoas Massif (PEAL)

This subdomain (of the southern domain) is a triangular shaped region of over 70 000 km² consisting of granite-migmatite basement with numerous Brasiliano plutons, some of which reach the size of large batholiths. In some parts, particularly in the eastern half, the gneiss-migmatite basement includes remnants of Transamazonian crust as well as some Archean relicts. In contrast, the migmatite gneiss and Brasiliano plutons in the western part of the PEAL have mainly late Mesoproterozoic Sm/Nd crustal formation ages (T_{DM}, Silva Filho *et al.*, 1999). The southern margin of the PEAL forms part of the hinterland for the Sergipano Fold Belt and includes a series of granitic rocks that could have been generated by northward-directed late Neoproterozoic (Brasiliano) subduction. The detailed structure and composition of the PEAL and its role in the tectonic evolution of the Borborema Province still requires extensive geological mapping, petrological, geochemical, and isotope studies.

Brasiliano Plutons

They constitute a ubiquitous part of the Borborema Province, comprising about 30% of the exposed bedrock (Sial, 1986, 1987). The distribution and chemistry of these plutons were recently reviewed by Ferreira *et al.* (1998) and Sial (1999). These plutons play several major roles in the tectonic history of the Borborema Province, and the results will be given throughout subsequent sections.

Chronostratigraphic Framework of the Borborema Province

One of the major problems in sorting out stratigraphic correlations within or among Precambrian terranes is the lack of the precise biostratigraphic markers that permit global or regional correlation. In the absence of other chronological markers, many early studies of Precambrian terranes relied on the usage of traditional field relationships to establish relative chronologies in local areas as well as the usage of lithostratigraphy for broader correlation. In many instances such correlation was very inaccurate, since there are relatively few lithostratigraphic markers that have unique time significance. Furthermore, estimates of elapsed time have often been too long or too short. For this reason, geologists working in Precambrian terranes must rely on some independent type of chronological tool, and the most common tools are those of radiometric geochronology and isotope geochemistry.

During the third quarter of the 20th century there was a rapid development of several methods of geochronology, most notably those related to the K/Ar, Rb/Sr, U/Pb, and Th/Pb decay systems. Of these, the K/Ar and Rb/Sr methods were (prior 1990) the most widely applied, due largely to the relative ease with which they could be used. As a result, many of the broad regional and global chronostratigraphic frameworks developed at that time (and up to the present) were based on data from these two systems. Unfortunately, more recent experience has shown that the ages obtained by these methods may be inaccurate, commonly being younger than the actual rock-formation ages needed for precise and accurate chronostratigraphic correlation.

During the last quarter of the 20th century there were major developments in three geochronological systems that have greatly enhanced the precision and accuracy of modern age determinations on Precambrian rocks. These were: (a) improvements in laboratory and instrumental techniques for U/Pb dating of zircon and other minerals that allowed this technique to be widely used on very small samples; (b) development of the $^{39}Ar/^{40}Ar$ methods for obtaining thermally and chronologically meaningful ages from the K/Ar system; and (c) development of methods based on the $^{147}Sm/^{143}Nd$ decay system, which allowed not only broad evaluation of regional crustal histories, but also the ability to obtain high quality ages on selected rock or mineral systems that were previously not possible to data.

Geochronology of the Borborema Province

In the case of the Borborema Province, the U/Pb and Sm/Nd systems were applied only in this last decade. For the fifteen years prior to 1990 data were obtained primarily using the K/Ar and Rb/Sr systems. Although these data provided many useful insights into the tectonic history of Northeastern Brazil, they were subject to many uncertainties. Thus, prior to 1990 the chronostratigraphic framework consisted of two major suites of rocks and events: (a) Transamazonian to Archean basement, and (b) Brasiliano *sensu lato* igneous, deformational, and metamorphic events. The corresponding history of the Transamazonian to Archean basement spanned over 1 billion years (*c.* >2.8 to <1.8 Ga), whereas that of the Brasiliano Cycle spanned several hundred million years (*c.* 900 to 400 Ma). As a result of geochronological research in the 1990s, we now know that this scenario was far too simple and orogenic cycles were too broadly defined. Table 1 summarizes the chronostratigrahic framework of the Borborema Province, as inferred from modem geochronological data and ongoing field mapping, using the following guidelines.

U/Pb geochronology

In the Borborema Province, as well as in most of the world, the U/Pb methods of geochronology using zircon or other minerals such as monazite and titanite (sphene) are the preferred methods for defining the original crystallization ages of igneous rocks, for defining ages of high-grade metamorphic events, or for detailed examination of sediment provenance. Although these minerals often yield discordant results for either the $^{207}Pb/^{235}U$ or the $^{206}Pb/^{238}U$ system, the combined use of both systems allows a precision and accuracy not obtainable by other methods. Consequently, the chronostratigraphic framework shown in Table 1 is based primarily on an extensive body of U/Pb data that has appeared over the past 10 years. In many instances the data were obtained from single zircon or monazite crystals, either through conventional isotope dilution analyses, Pb-evaporation techniques, or ion microprobe (SHRIMP) analyses, and include volcanic, plutonic, metamorphic, or detrital minerals in a variety of rock systems.

Sm/Nd isotope geochemistry

The Sm/Nd methods can often be used to define precise ages, particularly in the case of unaltered mafic rocks or garnet-bearing metamorphic rocks. However, their greatest utility in the Borborema Province is for defining the limits and broad crustal history of the many tectonic blocks, which occur throughout the province. Because this method is based on two rare earths elements, normal crustal processes do not create much (if any) elemental fractionation. The Sm/Nd radiometric clock established when crustal material is first extracted from the mantle continues to evolve through subsequent igneous, metamorphic, or sedimentary events, which affect the crustal material formed earlier

Some caution must be used in the interpretation of Sm/Nd results as commonly tabulated. The first involves use of so-called crustal residence ages, or T_{DM} following the depleted-mantle model of De Paolo (1981). These ages are indicative only of the general antiquity of the material in the rocks involved and cannot be used as precise ages, since model-dependent uncertainties can often exceed 100 to 200 million years for Precambrian rocks. Furthermore, if the Sm/Nd ratio of the rocks is close to mantle or bulk-earth values, then the age defined by extrapolation of modem Sm/Nd relationships back in time to a mantle evolution curve can be very inaccurate, with large uncertainties.

Secondly, as pointed out by Arndt and Goldstein (1987), Sm/Nd whole-rock relationships often are derived from the

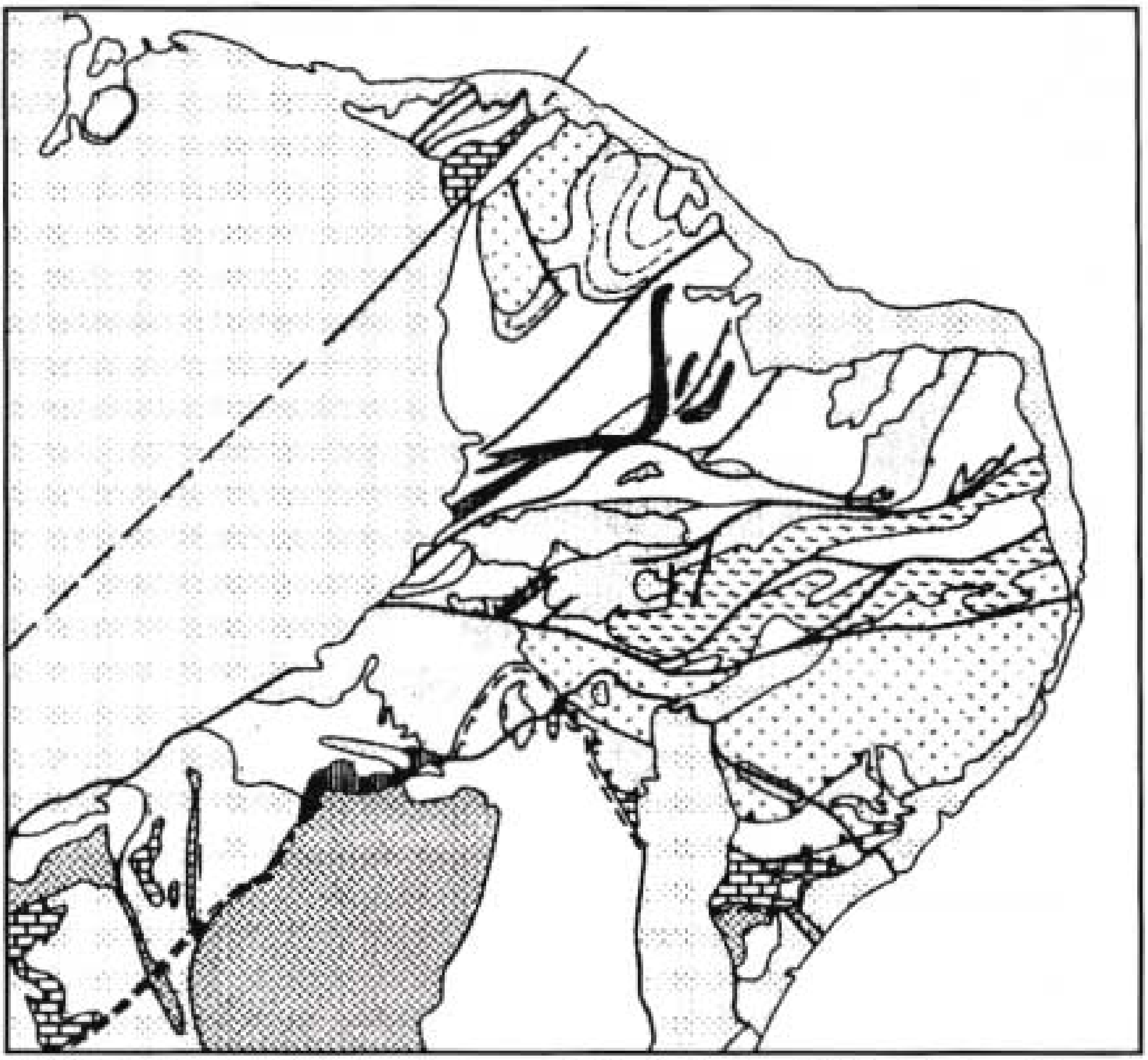


FIGURE 3 - The Archean nuclei of the Borborema Province are outlined, having as background the Figure 1. For all cases, these Archean nuclei are trapped within the structural framework of the Paleoproterozoic collage.

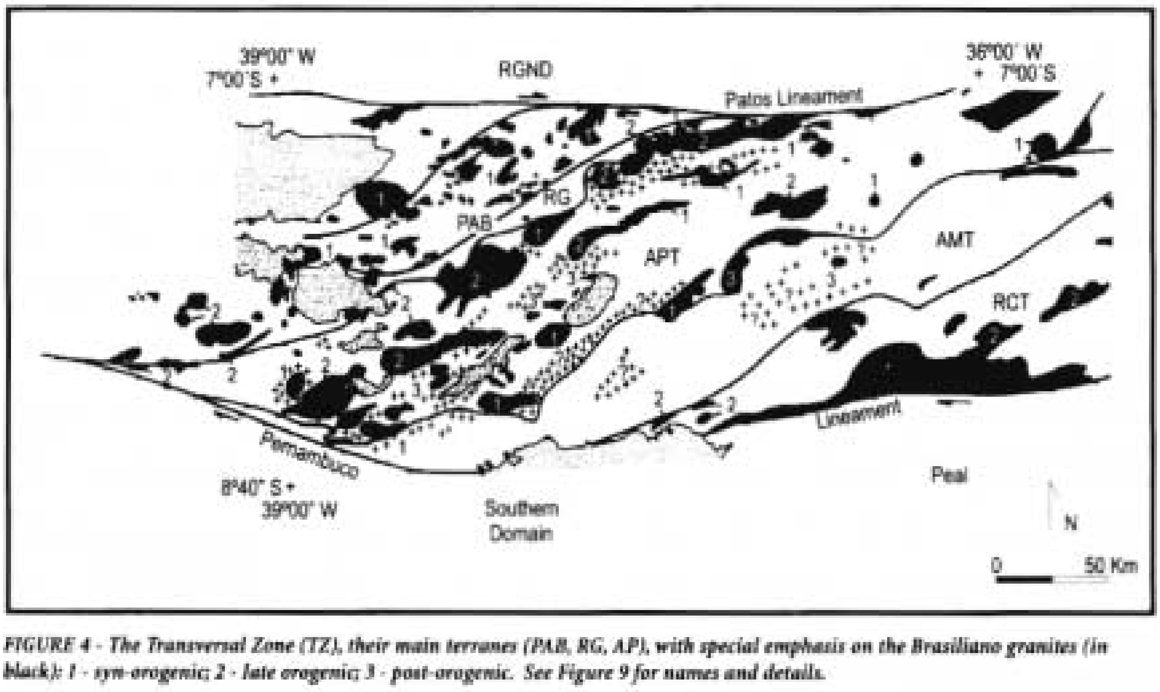


FIGURE 4 - The Transversal Zone (TZ), their main terranes (PAB, RG, AP), with special emphasis on the Brasiliano granites (in black): 1 - syn-orogenic; 2 - late orogenic; 3 - post-orogenic. See Figure 9 for names and details.

mixing of two or more components of differing ages, especially if the protoliths are sedimentary rocks or if crustal assimilation is involved. In these cases the T_{DM} ages reported may have no significance in terms of regional or local chronostratigraphy, although they can be very useful in setting limits on the ages of the rocks involved. For example, a rock having a T_{DM} of 1.2 Ga cannot have formed prior to that date, since that is the age that would be obtained if the rock represent a purely juvenile sample (extracted from the mantle at 1.2 Ga). That rock could, of course, have formed much more recently since recycling of originally juvenile material would not destroy that age. It could represent a rock formed as a result of magma genesis or sedimentation subsequent to 1.2 Ga. Thus, Sm/Nd T_{DM} ages should normally only be used to establish maximum ages of crustal rocks and used in conjunction with other data (such as U/Pb ages) to evaluate whether they represent juvenile, reworked, or mixed systems.

Sm/Nd whole-rock analyses are able to establish quickly whether a crustal unit contains material that has had a long, a short, or a mixed crustal residence age. In the Borborema Province, use of Sm/Nd analyses allowed the definition of major crustal blocks that include substantial amounts of post-Transamazonian material (Van Schmus *et al.*, 1995a, b), showing that the Brasiliano Cycle involved much more than just reworking Transamazonian to Archean basement. Continued application of these methods is providing further refinement of the broad crustal structure in the province.

Rb/Sr geochronology

There has been a tendency in recent years to avoid using Rb/Sr data, particularly in regions that have complex thermal and structural histories. There are numerous cases in the literature where seemingly good isochron ages have been shown subsequently to be wrong. Most commonly they are too young due to open system behavior subsequent to formation of the rocks concerned. However, in some instances the Rb/Sr isotope system remains closed or nearly closed even during high-grade events; this seems to occur most commonly when the systems are dry. Many Rb/Sr data obtained in the Borborema Province prior to 1990, and which were thought to represent partial resetting of Transamazonian to Archean rocks during the Brasiliano Orogeny, have subsequently been shown to represent essentially correct igneous crystallization ages of *c.* 950 Ma. Thus, Rb/Sr analyses are used here for establishing regional correlation of certain types of orthogneiss, particularly those related to the *c.* 1.0 Ga Cariris Velhos event (Brito Neves *et al.*, 1995a, b).

K/Ar and Ar/Ar geochronology

Most of the early K/Ar age measurements on rocks of the Borborema Province are too young due to subsequent Ar loss. So, Neoproterozoic ages and others due to partial rejuvenation are by far predominant. However, in recent years, $^{39}Ar/^{40}Ar$ methods have been used in the province to provide a control on the timing of late tectonic events related to the Brasiliano Orogeny (Monié *et al.*, 1997; Corsini *et al.*, 1998), including cooling ages of plutonic and metamorphic complexes and ages of strike-slip displacements. Continued application of these methods will be invaluable for detailed understanding of the final stages of the assembly of Western Gondwana.

Borborema Chronostratigraphy

The chronostratigraphic framework shown in Table 1 is based primarily on U/Pb geochronology of zircon, monazite, and sphene from a variety of rock types, supplemented with selected Sm/Nd and K/Ar ages of various metamorphic minerals. Use of these methods allows a level of accuracy and precision not readily attainable with other techniques, as reflected by the many individual events or stages listed in the table. This table includes several major advances in our understanding of the history of the region, as described below.

Brasiliano Orogeny

We recognize several distinct phases of the Brasiliano Orogeny and the relative short duration of the main orogenic cycle (640 to 570 Ma). This is in contrast to the traditionally used, relatively long Brasiliano Tectonic Cycle. So, it is necessary to start to define individual orogenic cycles more narrowly, in a manner consistent with Phanerozoic conventions and with the accepted tectonic behavior of the Earth.

Pre-Brasiliano supracrustal rocks

U/Pb methods are showing conclusively that there are several Neoproterozoic supracrustal sequences in the Borborema Province that formed prior to the main phases of the Brasiliano Orogeny. The occurrences that have been described may be only remnants of a more extensive system of middle Neoproterozoic supracrustal rocks, suggesting a significant interval of basin formation, volcanism, and sedimentation.

Cariris Velhos Orogeny

The recognition of the *c.* 1.0 Ga to 950 Ma Cariris Velhos Orogeny is a direct consequence of applying modern geochronological methods in a progressively well-mapped region. The significance of this event in regional and global history is presented later.

Mesoproterozoic

There is an almost total absence in the Borborema Province of igneous or high-grade metamorphic rocks with ages in the interval between 1.75 and 1.05 Ga, indicating that the pre-Carriris Velhos crust was probably part of a long-lived stable continent. Even though many of the Neoproterozoic supracrustal or plutonic rocks have Sm/Nd T_{DM} ages in this range, they may only represent hybrid Neoproterozoic and Paleoproterozoic sources and not any significant Mesoproterozoic crustal formation or reworking events.

Paleoproterozoic events

Current data on reworked terranes (basement and supracrustal rocks) of the Borborema Province allow resolution of several events throughout the Paleoproterozoic, which previously might have been grouped under a Transamazonian Tectonic Cycle. This is in keeping with using tighter restrictions on the duration of orogenic events, and hence we recommend that term Transamazonian be used up to the middle Paleoproterozoic as a collage. The Transamazonian Cycle may be broken into two main parts:

a older phase of crustal genesis (2.2 to 2.1 Ga) and a younger phase of deformation and metamorphism (2.1 to 2.0 Ga). This has good equivalency with the Paleoproterozoic data for the São Francisco Craton, besides which it roughly parallels the Birrimian and Eburnean events, respectively, of West Africa (to which we believe the Transamazonian events are the Brazilian equivalents).

Archean events

Within the Paleoproterozoic terranes/belts there are some preserved Archean nuclei. We recognize at least three distinct pulses of Archean petrogenesis, including one (3.5 to 3.4 Ga) that includes some of the oldest crustal rocks known on the South American Continent (Dantas *et al.*, 1998). Other significant events of plutonism and metamorphism occcurred at *c.* 3.2 Ga and 2.7 - 2.5 Ga. At this time the limited nature of these domains and the limited isotope data available preclude more details subdivisions of these rocks.

In summary, application of modem methods of geochronology and radiogenic isotope geochemistry by numerous researchers have allowed development of a relatively detailed chronostratigraphy for the Borborema Province over the past decade. Although this picture is neither complete nor perfect, we have used it as the basis for the discussions presented in subsequent sections.

Archean remnants

Lithostructural rock assemblages of Archean age (Fig. 3) are present to some degree in almost all the major domains of the province (with the exception of the MCO), and for all of them two observations can be made.

a) All of these are part of the Transamazonian orogenic framework (recycled crustal material, basement inliers, structural highs), and they have been reworked to varying degrees at different crustal levels (migmatization, intrusion of plutons, heating, shearing, stretching) during subsequent tectonic cycles, in particular during the Brasiliano Cycle;

b) Although the geographical extent of these true Archean areas is relatively small (only a few percent of the area of the province) they are scattered within the province within the Transamazonian parts of the basement. The isotope signatures and other indirect evidence for Archean events are widespread.

"Early" Archean (prior to 3.5 Ga)

At this time, there is no unequivocal evidence for crustal rocks in the Borborema Province that are older than 3.5 Ga. Some 3.45 Ga middle Archean gneiss has Sm/Nd crustal formation ages (T_{DM}) as old as 3.8 Ga, indicating the presence of older crustal material when they formed, but coherent crust of that age has yet to be found. Furthermore, no detrital or xenocrystal zircon older than 3.5 Ga has been reported yet for Northeastern Brazil.

In some supracrustal rocks of the province, suggestions of Archean source rocks have occasionally been found, these mostly giving late Archean T_{DM} ages. Considering the present level of geological knowledge and isotope data, it is reasonable to expect that there are more Archean nuclei than those already found, and that some others probably existed, but were completely reworked by Paleoproterozoic and Neoproterozoic tectono-thermal events. The majority of late Archean T_{DM} ages obtained up to now may represent different degrees of mixing between more ancient Archean and younger juvenile materials, and not necessarily a predominance of late Archean sources. Thus, it is still possible that evidence will be found for proximal early Archean crust in or near the Borborema Province.

"Middle" Archean (3.5 Ga to 3.0 Ga)

One of the most important Archean nuclei in the Borborema Province is the São José do Campestre Massif (Bom Jesus-Presidente Juscelino TTG block) of the RGND, SW of Natal and probably extending eastwards under the Cenozoic coastal sediments. Outcrops of this approximately 8000 km^2 basement area display the most extensive group of rocks that include both middle and late Archean ages. Grey tonalitic gneiss of the Bom Jesus Complex have yielded U/Pb ages of *c.* 3.45 Ga, but with a remarkable overprinting of other tectono-thermal events in the Paleoproterozoic (*c.* 2.0 Ga) and Neoproterozoic (*c.* 600 to 570 Ma), according to Dantas (1997) and Dantas *et al.* (1998). These TTG assemblages yield slightly older T_{DM} ages (*c.* 3.8 Ga), suggesting the involvement of Early Archean crust in their genesis. Garnet-bearing trondhjemitic gneiss exposed at Brejinho (35 km S of the City of Natal), within the same nucleus, gave a U/Pb age of *c.* 3.2 Ga and T_{DM} indications of a juvenile source. A small, deformed syenogranite pluton exposed to the S of São José do Campestre (on the boundary with the enclosing Transamazonian gneiss) yielded a U/Pb age of 2.7 Ga with a T_{DM} age *c.* 3.2 Ga (with negative ε_{Nd}).

"Late" Archean (3.0 Ga to 2.5 Ga)

Besides the São José do Campestre syenogranite, the Tróia-Tauá nucleus, situated in the southern part of the Central Ceará Domain, has an areal extent of about 6000 km^2. It is sharply limited to the E by the Senador Pompeu Lineament, and it is divided into two parts by the Sabonete-Inharé Fault. The western part (Tróia-Pedra Branca Block) is mostly composed of a volcano-plutonic sequence, consisting of anorthositic gneiss, and amphibolite, with some calc-silicatic rocks that gave U/Pb ages of *c.* 2.81 Ga and Sm/Nd data defining a short crustal residence. The eastern block, composed of high grade TTG suites also yielded an age around 2.8 Ga. It consists of a rock series with similarities to many Archean anorthosite bodies elsewhere in the world besides interesting occurrences associated with ultramafic rocks (Cr, PGE and Be in pegmatite). Evidence of younger crustal reworking is common, like the many features of the Brasiliano deformation (dyke swarms, shearing, thrusts, granite plutonism).

Usually, there are other occurrences of Archean rocks in the RGND (Granjeiro-Várzea Alegre, Jaguaretama Block, Cajazeiras, E of Orós), in the Caldas Brandão Massif (easternmost part of AMT, near Aroeiras), in the PEAL Massif (Riacho Seco, Poço Verde), in the basement of both

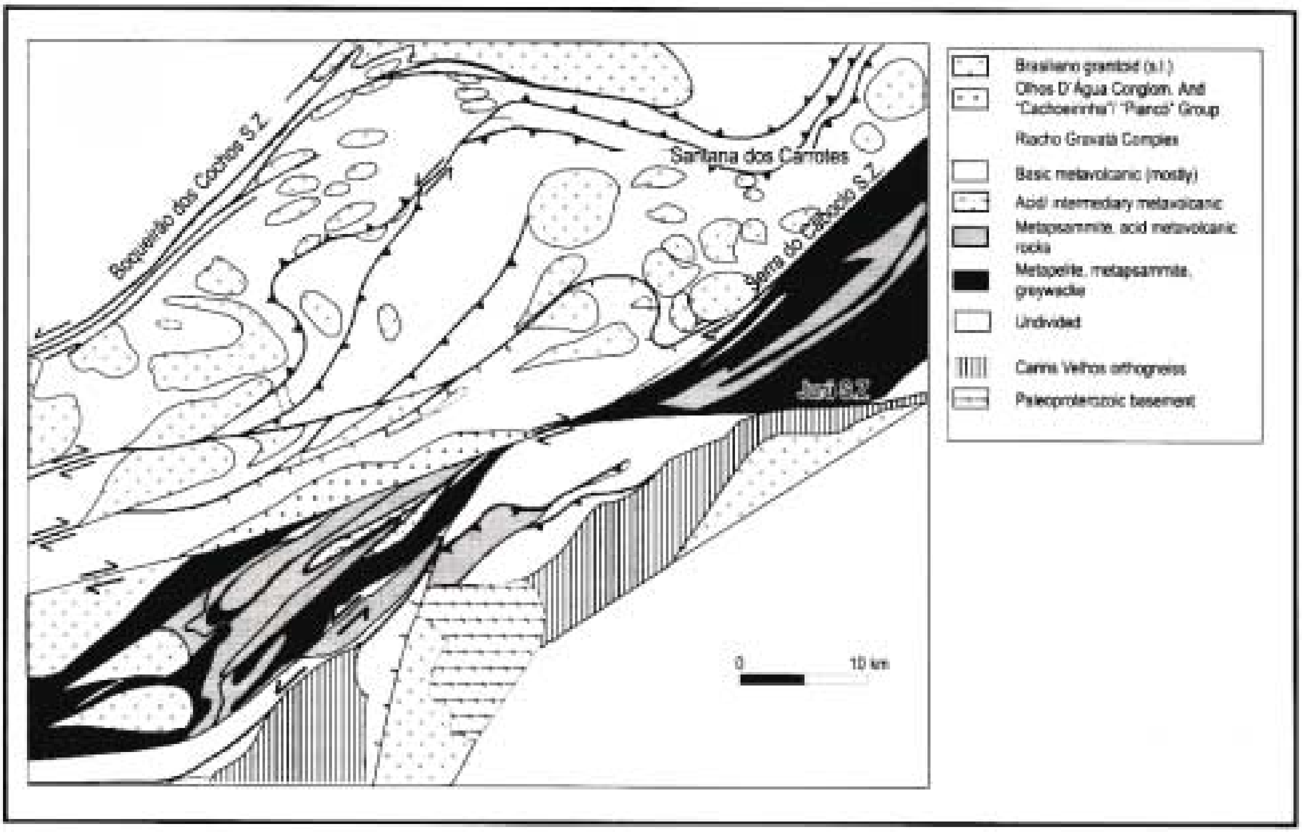


FIGURE 5 - The Serra do Caboclo shear zone, separating the Riacho Gravatá Terrane (RG) and the Piancó-Alto Brígida Fold Belt (PAB). Modified after Bittar (1998).

FIGURE 6 - The Transversal Zone and surroundings. Emphasis on the Paleoproterozoic basement ("massifs"), Neoproterozoic belts and the Cariris Velhos granitoid (augen-gneisses) along the Alto Pajeú terrane.

the Sergipano Belt (Jirau do Ponciano antiformal zone, Arapiraca Granulite, Simão Dias and Itabaiana domes) and the Riacho do Pontal-Rio Preto Belt (Sobradinho Massif, Uruás-Petrolina). All these rocks were submitted to overprinting by the subsequent Proterozoic events. They are mostly composed of high-grade gneiss (TTG suites, charnockitic-enderbitic suites) and migmatite. Most of these data (mostly with Rb/Sr method) suggest that late Neoarchean ages were dominant.

Archean Granitic Suites

The best described Archean TTG suites are exposed in that part of the São José do Campestre Massif, forming a preserved nucleus (Presidente Juscelino Block) inside the Santa Cruz Paleoproterozoic belt. This oldest granitic crust of the South American Platform consists of tonalite of the Bom Jesus unit that presents Sm/Nd model ages (T_{DM}) ranging from 3.5 to 3.8 Ga, having $\varepsilon_{Nd(T)}$ values at T = 3.4 Ga ranging from - 2 to - 4, suggesting an old sialic crust or enriched mantle derivation at >3.5 Ga. An episode of juvenile crust formation is known in the Mesoarchean (3.2 Ga tonalite and granodiorite of the Brejinho Complex), which was succeded by an anorogenic event that emplaced in the crust massifs of leuconorite, anorthosite and oligoclasite of granulitic protholiths around 3.03 Ga (Senador Eloi de Souza Complex; Dantas, 1997). Several minor Archean granitic units are also known in the Alto Moxotó Terrane, including the Itatuba and Barro Vermelho gabbro-anorthositic complexes within Paleoproterozoic orthogneiss.

Paleoproterozoic

The tectonic and lithostratigraphic history of the Borborema Province is very rich and has many chapters, some of them unsuspected in previous decades. It is now clear that all the basement of the different domains of this province represent the same kind of Paleoproterozoic orogenic system (collage) that characterizes the long-recognized basement of cratons that surround the province. This includes processes of sucessive accretionary and collisional orogeneses, forming tectonic collage and ending with continental accretion. Subsequent development of intracontinental extension and deposition of volcano-sedimentary sequences during the late Paleoproterozoic and Mesoproterozoic are also common factors. Even bearing in mind the many possibilities of displacements of the different domains (and internal terranes or sub-domains) of this province, mostly in Brasiliano times, it is necessary to think and speculate about a common supercontinent landmass that may have once included all these Paleoproterozoic orogenic systems. Thus, the idea of middle Paleoproterozoic continent called Atlantica, suggested by Ledru *et al.* (1994) and expanded by Rogers (1996), seems viable. This concept is improved when the basement of the Borborema Province, São Francisco Craton, São Luis-West Africa Craton, and the Parnaíba Basin are taken into consideration.

The same general concept of major Paleoproterozoic (Transamazonian) mobile belts with some scattered, re-structured Archean nuclei, may be observed both in the major exposures of the basement (massifs and blocks) and in many of the minor inliers (antiformal uplifts, tilted fault blocks). The basement of this province includes a particular characteristic (different from most Paleoproterozoic assemblages in the other continents), since it includes an event of juvenile crustal accretion between 2.4 and 2.3 Ga in the MCO basement (Fetter, 1999).

Early Paleoproterozoic (2.5 Ga to 2.2 Ga)

Occurrences of high grade rocks such as orthogneiss of tonalitic or quartz-dioritic composition, banded migmatite, kinzigite, khondalite, and enderbite have frequently been reported for the Granja Massif in the MCO (Brito Neves, 1983), where there is strong influence of NE-SW trending shearing along the Brasiliano lineaments. According to the recent data of Fetter (1999) such rocks yield U/Pb zircon ages of *c.* 2.35 Ga and Sm/Nd data that suggest a very short juvenile crustal history. The formation of this massif has been interpreted as the formation and amalgamation of island arcs during the early Paleoproterozoic. As such, this massif represents one of the few areas in the world where new crustal growth between 2.5 and 2.4 Ga has been found. So far no other examples of such crust have been found in the Borborema Province.

A similar set of Paleoproterozoic rock assemblages is mentioned by Dada (1998) in Ile-Ife grey gneiss of Nigeria, with U/Pb zircon data. In the adjacent CC domain, most of Nd model ages for high grade rocks of the basement also indicate values in the same time-span (Fetter, 1999).

Middle Paleoproterozoic (2.20 Ga to 1.90 Ga)

Special emphasis will be given here for the main intervals of the Transamazonian processes, 2.2 to 2.0 Ga, or more precisely, 2.15 ± 0.05 Ga for igneous crystallization ages (Hackspacher *et al.*, 1990; Dantas, 1997) and 2.05 Ga ± 0.05 Ga for granulite grade metamorphic ages (Dantas, 1997). The Mesoproterozoic orogenic processes (Transamazonian collage) are represented by widespread well-defined events in the basement of the Borborema Province and the basement of the bounding cratons, where the designation Transamazonian Cycle (imported form the eastern part of the Amazonian Craton) has traditionally been applied. The data for Transamazonian orogeneses (in Borborema and also in the São Francisco Provinces) are essentially coeval with the 2.15 ± 0.05 Ga Birrimian accretionary events in West Africa (Abouchamy *et al.*, 1990; Boher *et al.*, 1992; Dia *et al.*, 1997) and the 2.05 ± 0.05 Ga Eburnean deformational and metamorphic events (Toteu *et al.*, 1994) in West Africa.

The first of the Transamazonian orogeneses is the most important phase of Paleoproterozoic deformation and metamorphism in the basement of this province, with collisional tectonics apparently predominating over accretionary processes. Massifs or isolated crustal blocks representing this orogeny have been recognized in the basement of all domains and in nearly all segments of the province. In some cases the protoliths have Paleoproterozoic isotope signatures (Central Ceará Domain and João Câmara region, in the northeastern part of the RGNT), but in the majority of cases the Sm/Nd crustal formation ages (T_{DM}) fall in the late Archean (2.6 to 2.5 Ga), indicating significant involvement of older crustal material in the genesis of the Transamazonian protoliths.

To the SE of the Transbrasiliano Lineament (Sobral Fault in Ceará State), which forms the southeastern boundary of the MCO; Fetter (1999) concluded that the high grade orthogneiss and migmatite that comprise the basement in the Central Ceará Domain consist mainly of Transamazonian orogenic crust, although it also includes the deformed Tróia-Tauá late Archean nucleus. These crustal assemblages were formed and metamorphosed in the middle Paleoproterozoic (U/Pb ages between 2.2 - 2.1 Ga), but they also have Sm/Nd T_{DM} ages of 2.4 to 2.3 Ga indicating inclusion of some older crustal material during their formation. Fetter (1999) suggested that this area formed during a span of over 50 million years as a group of island arc terranes, like those of the slightly older Granja Massif with which they merged. These arcs were mainly isolated from the input of substantially older crustal materials.

Similar rock assemblages were found and described by Dantas (1997) and Dantas *et al.* (1998) surrounding the northeastern part of the the Presidente Juscelino- Bom Jesus Archean nucleus of the São José do Campestre Massif in the eastern part of the RGND. This João Camara occurrence is only a segment of a larger Transamazonian mobile belt, and it is composed of orthogneiss and migmatite that present general litho-structural and isotope characteristics (early Paleoproterozoic crustal formation ages) akin to those of the basement of Central Ceará.

Magmatic arc development with further collisional amalgamation has been mentioned in the Rio Piranhas and São José do Campestre Massif (Dantas, 1997; Hackspacher *et al.*, 1997). A very interesting process of volcanic island development was pointed out by Martins *et al.* (1998) for the area of Choró-Algodões, in the easternmost part of the Ceará Central Domain, where they described typical arc-related sequences (under high grade metamorphism) of volcanic, sedimentary and plutonic rocks with ages (formation and deformation) around 2.17 Ga.

With respect to the middle Paleoproterozoic it is necessary to discuss the isotope data that have given ages 2.05 - 1.95 Ga. Dozens of Rb/Sr isochrons (the majority of them from samples collected from different parts of the province), showed many zircon and monazite crystals with similar U/Pb ages (São José do Campestre, Caicó, Rio Piranhas, PEAL). These results have sometimes been obtained in samples from litho-structural units recognized elsewhere belonging to Transamazonian mobile belts. These could be recording a subsequent deformational episode (somewhat restricted) that occurred still in Mesoproterozoic times, and that has not yet been recognized in the basement rocks of the Borborema Province.

As previously stated the overall study of Transamazonian collage transcends the scope of the study of the Borborema Province. Several convergent island arcs, magmatic arcs, and collisional and traspressional events have been identified especially to the S of this province in the states of Bahia and Minas Gerais. In these localities, the overprinting of the Cariris Velhos and Brasiliano cycles is generally absent, and the age of 2.15 Ga and 2.05 - 1.95 Ga are very consistent. The same processes and ages are systematically being identified in the basement rocks of the Borborema Province, where the supremacy of this Paleoproterozoic event is being confirmed.

A reconstruction of the original shape, tectonic zones, arrangement and trends of the Transamazonian mobile belts is a difficult task. It has to take into account a series of superimposed tectonic movements (even those occurring in the late Paleoproterozoic), and the reworking at different crustal levels by two major Wilson Cycles (Cariris Velhos, Borborema). It requires the co-operation between workers on both the South American and African continents.

Some of the isotope data are of very good quality. Some were even preserved in rock units where the overprinting of the Brasiliano Orogeny was of considerable magnitude. In accordance with this statement and expectation, similar data are available for rocks occurring in the states of Bahia and Minas Gerais (São Francisco Craton) and they are consistent with the building of some fold belts (Teixeira *et al.*, 1997; Sabaté *et al.*, 1997), that were also described as Transamazonian.

When applying the data from Bahia and Minas Gerais (and other structural provinces), a concept of collage (Sengor, 1990) rather than one orogenetic process connected to the amalgamation of a continental landmass seems to be more appropriate to the Transamazonian.

Late Paleoproterozoic (1.90 to 1.60 Ga)

In the northern hemisphere the late Paleoproterozoic orogenies (Penokean, Trans-Hudson, Wopmay, Yavapai, Mazatzal, Labradorian, and so on) were characterized by arc accretion and continental collision (convergence). In contrast, the late Paleoproterozoic tectonic environments in the Borborema Province as well as in most of Atlantica during the late Paleoproterozoic (with some local continuation into Mesoproterozoic) was characterized by development of extensional basins and deposition of intracratonic volcano-sedimentary sequences. These extensional processes were synthesized by Brito Neves *et al.* (1995a, b), who described them from Venezuela to the N of Argentina as the Statherian Taphrogenesis. In the Borborema Province the main occurrence of such processes, and their products, is in the western part of the RGND, a rectangular area bounded by the Senador Pompeu Lineament to the W and the Portalegre Lineament to the E.

From the town of Pio IX in State of Piauí to northeastern Ceará and northwestern Rio Grande do Norte, there occur remnants of volcano-sedimentary sequences with bimodal volcanism (felsic>mafic), andesite, psammo-pelitic metasediments, and some carbonate-alkaline granite may occur locally in and around these occurrences. These occurrences have informally been referred as basins or belts, with names like Pio IX, Orós, Jaguaribe, Viraponga, Peixe Gordo, Quixopá, São José, Encanto, and W Potiguar (Jardim de Sá, 1994; Parente *et al.*, 1998; Cavalcanti, 1999), in the J-WP subdomain (of the RGND). Also, they have collectively been named as Jaguaribeano Belt (Brito Neves, 1983) or Orós-Jaguaribe Belt (Jardim de Sá, 1994; Cavalcanti, 1999). In many instances some local names may only designate isolated remnants of previously more widespread and more continuous cratonic cover represented by these rock assemblages. These rock assemblages (petrological and geochemical characteristics), their tectonic and paleogeographical environments, their isotope ages (*c.* 1.8 - 1.75 Ga), show many similarities with those of the Espinhaço Belt (western Chapada Diamantina) overlying the São Francisco Craton, and they probably were all together part of a major Pan-Atlantica cratonic cover. Transpressional movements associated with the intra-domain lineaments during the Brasiliano Cycle deformed these rocks. This means that for the best part they behaved as undeformed sequences for more than 1 Ga after the original crustal formation in late Paleoproterozoic times

Paleoproterozoic granitic suites

Paleoproterozoic granites suites are a major component of the basement blocks in the Borborema Province, now consisting of orthogneiss. The oldest episode corresponds to the generation of 2.5 Ga juvenile granitic crust in the Granja Massif. Another episode of 2.1 Ga juvenile crust formation occurred locally in the Rio Piranhas Massif (Dantas, 1992). Subsequent magmatic units appear dispersed in diverse Paleoproterozoic blocks, in general representing small gabbro-anorthositic intrusions and disrupted mafic dyke swarms. Other episodes of anorogenic granite emplacement and related rocks, such as the 1.77 Ga Serra do Deserto Granitic Suite consisting of alkaline granite in the aborted rift system of the Orós-Jaguaribe Belt, occur in the transition between the Paleoproterozoic and Mesoproterozoic, representing taphrogenic (intracratonic) events. These had hardly terminated before the onset of the more extensive rifting of the younger Cariris Velhos and Brasiliano cycles.

Although there are many Sm/Nd crustal formation ages (T_{DM} in parts of the Borborema Province that are between 1.6 Ga and 1.0 Ga; Van Schmus *et al.*, 1995a, b), virtually all represent Neoproterozoic igneous and sedimentary rocks that incorporated significant amounts of older crustal (Paleoproterozoic, Archean) material during their formation. It is possible that several cratonic sedimentary sequences were deposited on older crust of the Borborema Province during the Mesoproterozoic. This being the case, subsequent uplift, tectonism, and erosion have virtually destroyed that record.

Latest Paleoproterozoic to Middle Mesoproterozoic Stasis

Following the formation of the late Paleoproterozoic extensional basins there was a long hiatus in the tectonic evolution of the Borborema Province. There are very few igneous, sedimentary or metamorphic suites that can be shown to have formed between 1.7 and 1.1 Ga. The evidence for formation of supracrustal rocks between 1.1 and 1.0 Ga (as was previously assigned for the Cariris Velhos Orogeny in some papers) is still poorly developed, and still requires refinement. In fact, the only unequivocal Mesoproterozoic pluton known at present was reported by Sá *et al.* (1997), who showed from U/Pb zircon ages that the augen gneiss metapluton at Taquaritinga-PE originally formed about 1.5 Ga.

The A-type pluton of Serra de Taquaritinga in the Rio Capibaribe Terrane is characterized by a K-feldspar megacrystic augen-gneiss ranging in composition from granodiorite, through monzogranite, syenogranite to quartz syenite. It has Ba, Th, Zr, Nb and Y-enrichment and a high fractionated, light REE enriched REE pattern exhibiting a pronounced negative Eu anomaly. This pluton may represent *c.* 1.5 Ga extensional intra-plate phenomena preceding the subsequent 1.1 to 1.0 Ga extensional phases of the Cariris Velhos Orogeny in the eastern part of the TZ.

Thus, there is a major gap of about 600 million years in the documented history of terranes that now comprise the basement of the Borborema Province. In accordance with the Rogers (1996) concept of Atlantica, we suggest that the interior of this continental mass remained stable until near the end of the Mesoproterozoic.

Late Mesoproterozoic to early Neoproterozoic (Cariris Velhos Orogeny)

The assignment of the Cariris Velhos Orogeny to this time interval marks an important chapter in the history of geological research on the Borborema Province. However, this is done with some degree of uncertainty, since this proposal is based on rather limited data, and data indicating a late Mesoproterozoic ages requires refinement. However, the more robust data suggest that most, if not all, of the Cariris Velhos orogenic activity took place in the earliest Neoproterozoic (1.0 Ga to 940 Ma). The principal uncertainty concerns the beginning of presumed continental extension that created the various depositional basins. Geochronological control on this part of the orogenic cycle is still weak, and it may have begun in the latest Mesoproterozoic.

The development of this geodynamic cycle involved the interaction of the domains N of the Patos Shear Zone (CC+RGND) with a continental mass (northern parts of the present São Francisco-Congo cratons) to the S. The main products of this orogenesis are now situated mainly in the TZ, S of the Patos Lineament. Although they were also reworked to different degrees by the Brasiliano Orogeny about 600 Ma, the rocks formed during the Cariris Velhos Orogeny are recognized in the field or (more easily) from U/Pb data and, in many cases, Rb/Sr geochronology (Brito Neves *et al.*, 1995a, b). Up to now, there is no evidence for Cariris Velhos orogenesis N of the Patos Shear Zone. However, to the S of the TZ there are indications for its presence in the basement of the Southern Domain (SD), including the western part of the PEAL Massif and northernmost parts of the Sergipano Belt.

The products of this orogenic cycle have been attributed to the Brasiliano Pajeú-Paraíba orogenic system by Brito Neves (1975), and this regional designation was followed by several authors. However, Campos Neto *et al.* (1994), Santos (1995), Van Schmus *et al.* (1995a, b) and Brito Neves *et al.* (1995a, b), among others, brought new insight and began using the designation of Cariris Velhos Cycle. Santos (1996) and Santos *et al.* (1997) proposed the identification of individual terranes in the former Pajeú-Paraíba Fold Belt, as well as for the entire Borborema Province. The designations of Alto Pajeú (AP), Alto Moxotó (AM) and Rio

Capibaribe (RC) terranes or segments in the interior of ZT will here be followed (Fig. 2) without further discussion on this always controversial subject of terrane nomenclature. But, both the late Mesoproterozoic early Neoproterozoic Cariris Velhos Orogeny and the later reworking by the Brasiliano Orogeny were considered in defining these terranes. The litho-structural assemblages in these crustal segments record different aspects of the tectonic elements of the Cariris Velhos Orogen.

Alto Moxotó Terrane (AM)

The Alto Moxotó Terrane (Figs. 2 and 4) consists of large areas of Paleoproterozoic basement, including some Archean remnants. It is possible to follow the basement structures with varying degrees of Brasiliano overprinting from the northern part of the São Francisco Craton up to the coastal areas of eastern part of the State of Paraíba, where it was formerly called the Caldas Brandão Massif. Some early Neoproterozoic supracrustal sequences (Lagoa das Contendas, Sertania, Caroalina) and granite (calc-alkaline and trondhjemitic types) also occur in this segment. On the other hand, there are relatively few Brasiliano granite bodies in contrast to the AP to the N.

The Lagoa das Contendas supracrustal rocks occur exclusively in the western part of AM, particularly around the town of Floresta, PE. They represent arc-type volcano-sedimentary and exhalative deposits (Santos, 1995). Metapelite beds, probably deposited in pelagic environment, include chemical and exhalative iron facies interbedded with medium to high-K calc-alkaline volcanic sequences with geochemical patterns typical of a mature magmatic arc. Several types of gneiss, schist, (garnet-biotite, graphite, corindum), quartzite, calc-silicate rock, and banded-iron formation units occur in the sequence. Andesite, trachyandesite, andesitic basalt, alkali basalt, and dacite are the main volcanic components. They have meta-aluminous and calc-alkaline (high K) signatures, with REE patterns suggesting a mature arc environment. Recent U/Pb dating of zircon by SHRIMP (Van Schmus *et al.*, in preparation) indicate ages of *c.* 970 Ma for this volcanism. This is slightly younger than previously reported ages of *c.* 1.012 ± 0.018 Ga given by Van Schmus *et al.* (1995a, b), probably due to the influence of some detrital components in the zircon analyzed in the earlier work.

The Sertania Complex occurs extensively in AM as a thick and monotonous sequence of aluminous supracrustal rocks (Table 2), including sillimanite-garnet-biotite gneiss and schist, with minor calc-silicate and amphibolitic rocks. They display high-grade metamorphism with many occurrences of migmatitization. Some psammitic facies are represented locally by muscovite-bearing quartzite and muscovite-bearing gneiss, the former including some dark banding with iron-bearing minerals. The Sertania Complex is typical of the AMT terrane, even though there are some small similar occurrences in the northern APT. There are no direct isotope ages for rocks of the Sertania Complex yet, but they are attributed to the same early Neoproterozoic interval as the Lagoa das Contendas units. Usually, available Rb/Sr and K/Ar data have yielded late Neoproterozoic ages due to the Brasiliano reworking.

The Caroalina Complex crops out as vestiges (or allochtonous?) schist belts in the central southern part of AMT, near Custodia, PE. The field relationships suggest a stratigraphic position younger than that of the Sertania Complex. The Caroalina Complex consists of garnet-biotite-muscovite schist locally with sillimanite and kyanite with a basal quartzitic unit of local occurrence. Several stratigraphic and metamorphic arguments may be used to suggest a post-Cariris Velhos age for this unit, possibly middle Neoproterozoic like the Baixa Grande units to be mentioned later.

Calc-alkaline granite related to the magmatic arc environments of the Lagoa das Contendas units also occur in the AM, mostly in the area of Floresta, PE. There are also some trondhjemitic plutonic rocks that were considered by Santos (1995) as the second member of the bimodal sequence of Malhada Vermelha (mafic-ultramafic magmatism). This sequence mainly consists of severely deformed plutons, with gabbro, quartz-gabbro, monzogabbro, and anorthosite. The geochemical data suggest an oceanic affinity from an eclogitic source rock (Beurlen, 1988). This Malhada Vermelha Suite occurs in the interior of the AMT or in contacts with Lagoa das Contendas units. The ages of the calc-alkaline granite bodies are *c.* 1.0 Ga (Van Schmus *et al.*, 1995a), but these data require refining. The Malhada Vermelha Suite is presumed to be of the same general age, but there are no geochronological data available to confirm this.

Rio Capibaribe Terrane (CRC)

This is the informal designation for the southeastemmost part of the ZT (Santos and Medeiros, 1999; Medeiros, 1999) that is bounded by the Congo-Cruzeiro do Nordeste Lineament (to the N) and the Pernambuco Lineament (to the S), where the Transamazonian basement crops out only in some inliers, covered by two different sequences, both in age and lithological content.

The Vertentes Complex is of volcano-sedimentary nature, and probably related to the Cariris Velhos Cycle. It consists of metapelite, metaturbidite with intercalations of metabasalt, metavolcanic rocks of intermediate composition, and metavolcaniclastic sediments (including bombs) and sheet-like granitic bodies. Migmatitization is a common feature of this older complex. There are no geochronological data for this unit, and its relationships with the 1.5 Ga granite of the Serra Taquaritinga is still unknown, and the reference to the Cariris Velhos Orogeny is conjectural.

The Surubim Complex is mainly composed of metasediments including quartzite-pelite (biotite-schist is very common) and carbonate, of probable Neoproterozoic age. It lies unconformably on the basement units and on the Vertentes Complex (locally as a klippe).

Additional characteristics for this terrane are the Mesoproterozoic anorogenic features such as the gabbro-anorthosite units and mafic dykes of Limoeiro-Passira (1.7 Ga, Ignês Guimarães, unpublished) and the A-type granitoid of Taquaritinga (1.5 Ga, Sá *et al.*, 1998). Important contractional tectonic movements are also worthy of mention. These affected large Brasiliano batholiths as well as the anorogenic granite bodies (Serra de Taquaratinga overthrust), as well the strike-slip tectonics controlled by the lineaments that bound this triangular-shaped terrane.

Alto Pajeú Terrane (AP)

The Alto Pajeú Terrane is characterized mainly by the principal plutons, stocks, and sheets of Cariris Velhos orthogneiss and secondarily by the predominance of volcano-sedimentary supracrustal assemblages. The AP follows a WSW to ENE trend parallel to and N of the AM, diagonally within the TZ, and reaches the coastal area near the boundary between the states of Paraíba and Rio Grande do Norte (Figs. 2, 4, and 6). As an additional characteristic, the AP is intruded by many late Neoproterozoic plutons having synorogenic (high-K, calc-alkaline) and post-orogenic (high-K, shoshonitic) affinities. As a consequence, the AP was extensively reworked by Brasiliano structural deformation and metamorphism. To the N of the AP there is a complementary belt of a volcano-sedimentary nature, the Riacho Gravatá Complex or Group, that is a good candidate for a previous site of fore arc development, if the the Cariris Velho magmatic arc was developed on the northern edge of AP.

The main volcano-sedimentary assemblage present in AP is the São Caetano Complex, which consists of garnet-biotite gneiss and metagreywacke with several other types of metasedimentary or metavolcanic rock such as limestone, calc-silicate rock, quartzite, dacite, basalt, and chert. The regional metamorphism is in the amphibolite facies, with common formation of stromatic migmatite. In the southwest part of the AP, Santos (1995) described an important sequence, the Poço do Salgueiro Complex that consists of metapelite, dacite, rhyodacite, basalt, scattered gabbro, and chert. The metavolcanic rocks of these areas exhibit geochemical patterns of a high-K calc-alkaline series akin to those of mature magmatic arcs. The ages of this volcanism are still poorly constrained (younger than 1.08 Ga by U/Pb in zircon), but field relationships with the *c.* 970 Ma orthogneiss suggests a contemporaneous tectonic setting for both. Locally, some of these granites (now augen-gneiss) occur intruding the São Caetano paragneiss.

The Pedras Pretas Suite is composed of a group of metamorphosed mafic to ultramafic rocks and occur in the area between the APT and the AMT. They are important because of the presence of Fe-Ti-V mineralization. The geochemical data for this suite indicates a tholeitic affinity, similar to the MORB. The suite includes some garnet-bearing rocks that have been described as originally as eclogite (Beurlen, 1988).

The Riacho da Barreira Complex (Santos, 1999, unpublished) has a restricted occurrence in the area around Mirandiba, PE, near the boundary between the AP and AM. It is different from the São Caetano Complex because of the relative abundance of carbonate and quartzite, as well as having intercalations of meta-ultramafic and minor meta-mafic rock. Its volcano-sedimentary facies consists of metapelite with intercalations of tremolitite, actinolitite, chlorite, and talc-bearing schist, hornblendite, pyroxenite, and iron formation units. The sedimentary facies contains muscovite-bearing gneiss and cordierite-sillimanite-garnet schist with intercalations of quartzite, marble, and calc-silicate rocks. In the absence of isotope ages, field relationships suggest that this complex is middle Neoproterozoic, in like manner to the Baixa Grande assemblage and Caroalina Complex.

The Baixa Grande Complex is a small volcano-sedimentary sequence occurring in the southeastern part of Afogados do Ingazeira, PE area (*c.* 40 km N of the Caroalina Complex; Fig. 5). It is shown as part of the Paleoproterozoic Irajai Complex on the geological quadrangle map of Afogados do Ingazeira (Veiga Jr. and Ferreira, 1990), but recent U/Pb data on zircon from meta-tuff indicate that its rocks are of middle Neoproterozoic age (740 to 720 Ma). It contains immature pelitic metasediments, volcaniclastic rocks, meta-andesite, meta-basalt, tuff, metadacite, and some local intercalations of limestone, calc-silicate rock, and metachert; it has a low to medium grade of metamorphism (lower amphibolite facies). Limited geochemical data are interpreted in contradictory ways, with some authors claiming oceanic arc environments, but this is not supported by Sm/Nd data, which show significant contributions from older crust. The Baixa Grande, Caroalina, and Riacho da Barreira complexes may be remnants of former middle Neoproterozoic intracontinental extensional basins.

Riacho Gravatá Belt

The Riacho Gravatá Belt (Fig. 5) is a narrow and irregular belt positioned between the Serra do Caboclo Shear Zone (to the N) and the northern part of the AP, which constitutes an important linear zone of uplift (Teixeira-Terra Nova High; Tertiary in age, but with inheritance of the basement) that crosses diagonally the TZ. This belt is composed by five lithological units, always defined by tectonic contacts. It contains sedimentary, volcanic and plutonic rocks, mafic metavolcanic rocks, metachert and metamarl (A); metavolcanic and metaplutonic of acid to intermediate composition (B); metapsammite and acid metavolcanic units (C); metapelite and metapsammite, metarhytmite, metagreywacke and tuff (D) phyllite and quartzite, muscovite-schist, according to Bittar (1998). The grade of metamorphism is green-schist with some biotite and garnet, and a structural complexity due to shear zones. Geochronological data, from Rb/Sr and U/Pb methods suggest a Tonian age (950 ± 10 Ma) for this particular volcano-plutonic-sedimentary association. This is to say that it has the same age as the Cariris Velhos orthogneiss to the S. Some preliminary geochemical data (Melo *et al.*, 1996) suggests an intraplate tensional environment, but this is a subject that demands further investigation.

Cariris Velhos Plutonism

Granitic (orthogneiss) suite

The typical granitic suite of Cariris Velhos occur as a long plutonic belt (Fig. 6) situated along the AP and part of the SD, extending for more than 750 km, from the Riacho do Pontal Belt (states of Pernambuco and Piauí) to the Atlantic coastal area at the boundary between states of Paraíba and Rio Grande do Norte. Several types of granitoid pluton have been recognized, as batholiths (minor), stocks, sheet-like intrusions, with varied composition such as syenogranite (c. 60%), monzogranite (c. 18%), granodiorite and quartz-diorite (*c.* 11%) and quartz-syenite (*c.* 5%.), among the most common types up to now sampled.

Two basic textural types should be mentioned: 1) Tiacho do Forno two-mica migmatite and granitoid, and 2) Recanto biotite (minor muscovite) augen orthogneiss. Both types contain garnet, locally sillimanite, and show a narrow compositional range, especially in the syenogranitic and monzogranitic fields. They have identical REE and trace elements pattens that are similar to the enclosing metavolcaniclastic rocks of the São Caetano Complex, suggesting their generation by incongruent melting of mica of this hosting sequence (Santos, 1995). Usually they present penetrative foliation occurring in zones of high strain, and locally they occur as leucosomes in stromatic migmatite. According to Barbarin (1997) and Brown and Solar (1998) they contain many features typical of collisional granites.

Detailed U/Pb data, using single-crystal isotope dilution analyses (Kozuch *et al.*, 1997) and ion-microprobe analyses suggest ages for the orthogneiss in the range of 980 to 950 Ma. Rb/Sr isochron ages are similar or only slightly younger (*c.* 950 Ma, Brito Neves *et al.*, 1995a, b), suggesting that the Brasiliano deformation and metamorphism had relatively little effect on the Rb/Sr system. Sm/Nd isotope data yield T_{DM} ages between 1.7 and 1.2 Ga and $\varepsilon_{Nd(T)}$ from + 2.3 to - 1.8, suggesting that these plutons crystallized from melts the sources of which were neither exclusively juvenile nor exclusively older Transamazonian continental crust. The data are consistent with mixing of juvenile mantle-derived material and older continental crust, such as would occur for magmas formed in a continental magmatic arc

A major granitic suite of Cariris Velhos age, consisting mainly of orthogneiss, is characteristic of the ZT and parts of the Southern Domain. It extends for more than 700 km from the Riacho do Pontal Belt to the Atlantic Ocean. In the AP the dominant facies are peraluminous muscovite-bearing migmatite and deformed granitoid (augen orthogneiss) that occur generally as intrusive sheets stacked along the thrust surfaces of this orogenic event. The plutons show compositional ranges restricted to the monzogranitic and syenogranitic fields and show two distinct geochemical patterns. The peraluminous migmatite and leucogranite, dominant in the AP are similar to the CW sub-type of Barbarin (1996). The REE and trace element patterns are the same as the arc-type volcaniclastic sequence in the adjoining block, suggesting they represent crustal melts in an arc/continent collision scenario. The association of some juvenile mafic to ultramafic rocks (Serrote das Pedras Pretas Suite) indicates that crustal accretion may have represented an important role in parts of the Cariris Velhos Belt. On the other hand, the composition of certain leucogranite and trondhjemite bodies of the Alto Moxotó segment are similar to the lower continental crust (LCC) and bulk continental crust (BCC) in composition. These can be related to subduction fluid or enriched mantle sources. The subduction signatures and the steep fractionated REE patterns for some of these AMS leucogranite bodies resemble those of recently identified adakite, the origin of which can be related either to fusion of subducted oceanic crust or to the fusion of basaltic roots of an oceanic plateau or volcanic rifted margin above a subduction zone (Pearce, 1996).

Initial U/Pb data for zircon indicated ages ranging from 1.037 Ga to 999 Ma for intrusion of these granitoid plutons (Jardim de Sá, 1994; Santos, 1995; Van Schmus *et al.*, 1995a, b). More recent, detailed U/Pb data, using single-crystal isotope dilution analyses (Kozuch *et al.*, 1997) and ion-microprobe (SHRIMP) analyses (Van Schmus, unpublished) now point to ages of 980 to 950 Ma as being more representative of the Cariris Velhos Magmatism. The older ages obtained initially probably show some influence of older components in the populations analyzed. The Rb/Sr isochron ages are somewhat similar or only a little younger (*c.* 950 Ma; Brito Neves *et al.*, 1995a), suggesting that Brasiliano deformation and metamorphism had relatively little effect on the Rb/Sr system. Sm/Nd isotope data yield T_{DM} values between 1.3 and 1.6 Ga and $\varepsilon_{Nd(T)}$ from - 1.8 to + 0.8, suggesting that these plutons crystallized from melts the sources of which were neither exclusively juvenile nor exclusively older (Transamazonian) continental crustal. The data are consistent with mixing of juvenile mantle-derived material and older continental crust such as would occur for magmas formed in a continental magmatic arc.

With respect to the country rocks into which the granite bodies are intruded, it is possible to distinguish different types in the AM. The first are granite bodies hosted by the Lagoa dos Contendas Complex, apparently related to the genesis of the arc-type calc-alkaline magmas of this complex. The São Pedro-Riacho das Lages Suite consists of a calc-alkaline trondhjemite (TTG) series where the composition varies from tonalite to monzogranite. There are few available trace element patterns, which are compatible with plagiogranite (MORB), but a M-type volcanic arc association may be suggested. It is possible that these rocks may somehow have an association with the gabbro-diorite (minor anorthosite) Malhada Vermelha Suite. A second type corresponds to leucosomes of stromatic migmatite and biotite-granite intrusives in the Sertania Complex and in the Paleoproterozoic Floresta Complex. In both cases they are peraluminous monzogranite with indications of being syn-collisional plutons. But REE patterns exhibit high LREE/HREE ratios and a strongly negative Eu anomaly, probably reflecting distinct sources for the granite. The U/Pb isotope systems of these granite plutons of the AMT are different, showing zoned zircon with inherited nuclei. The Concordia diagram usually presents large errors and show the effects of the Brasiliano orogenies. The scarcity of Brasiliano rock units in the AMT suggests that the Cariris Velhos granite plutons could be formed by partial melting of older basement sources.

In the southeastern part of the Afogados da Ingazeira Shear Zone, granite with characteristics similar to Cariris Velhos plutons show U/Pb zircon ages, both by Concordia and dissolution methods, varying from 1.9 Ga to 1.0 Ga (Leite and Lima, 1997). Field data such as structural control and the presence of Mesoproterozoic supracrustal xenoliths, confirm these are Cariris Velhos granitoid plutons, demonstrating that these melts can in some cases preserve zircon of various ages, representing both Transamazonian and Cariris Velhos sources.

Mafic - ultramafic rocks

There is a discontinuous trend of mafic-ultramafic rocks, with relicts of high pressure granulite and eclogite facies along the APT-AMT boundary that has been attributed to the Cariris Velhos Belt (Santos, 1999). The most well preserved Serrote de Pedras Pretas Suite is exposed as interlayered sheets within the Cariris Velhos orthogneiss in the Lagoa das Pedras Complex, N of Floresta, PE. These rocks display geochemical patterns of a tholeiitic basalt series with MORB affinities (Santos, 1995). This suite includes pods of garnet-bearing rocks also within the *Icaiçara Block in the westernmost part of the AMT*, presenting similarities with eclogite associated with subduction zones (Beurlen *et al.*, 1992). Other geophysical and tectonic features are associated with that AMT-APT boundary to the E, where there occur scattered occurrences of rocks formed in high-pressure environments, interpreted as a disrupted ophiolitic suture. There is no direct geochronological data for these occurrences, but there are some local indications (Guimarães, unpublished) in the Icaiçara Block supporting a Cariris Velhos age.

The meaning of the Cariris Velhos Cycle

Much research remains to be carried out on the Cariris Velhos Orogen, and the eventual views may be very different from those held today. One must bear in mind that the Cariris Velhos Orogeny is a new concept in the Borborema Province. The first data on this cycle began to appear in the early years of this decade, following a series of earlier and somewhat contradictory suggestions of its meaning and age.

Nevertheless, all conclusion about the complete evolution of an early Neoproterozoic orogenic cycle are reliable and compatible with old and new geological and isotope data, much of which is still unpublished. There are many problems regarding detail. However, a better alternative hypothesis has not yet been suggested.

The Cariris Velhos Cycle seems to be complete, with all the main phases of a short-lived Wilson Cycle. The cycle began probably before 1.0 Ga with the accretion of a large continental landmass. Locally, this was the Paleoproterozoic supercontinent, Atlantica, although at the time of accretion *it may have been part of the larger Rodinian geodynamic* system. In any case, this accretion resulted in formation of several extensional basins, detached blocks of Transamazonian crust as fragments of Atlantica; and continental margins along larger remnants of Atlantica at the northern edge of the São Francisco-Congo Craton. Drifting and formation of oceanic crust presumably followed the rifting phase between some fragments and blocks, although evidence for this is sketchy. Some of the larger continental fragments subsequently converged, forming litho-tectonic assemblages of the type associate with *foreland and fore-arc basins*.

The Riacho da Barreira Complex (Santos, 1999) in the boundary AP-AM consists of pelitic and chemical sequences that contain a facies with abundant intercalations of ultramafic and mafic rocks, and iron formation units of unknown origin. The Caroalina and Surubim complexes form metasedimentary remnants quartzite-pelite, and subordinated carbonate sequences over older supracrustal rocks and over the basement of the AM and RC. They have preliminary been assigned Neoproterozoic ages (Table 2), but without the necessary geochronological support.

Other similar occurrences related to this continental break-up are possible and can be predicted. However, these are likely to be covered by younger Neoproterozoic sequences or completely reworked by the Brasiliano tectonics. In the Southern Domain (SD), in the same middle Criogenian times, there occurred sedimentation on the craton consisting of diamictite beds and rock assemblages related to the lowest stratigraphic levels of the Bambuí Group, showing widespread conditions of continental glaciation. These sediments (Bebedouro, Capitão-Palestina, Macaúbas, Canabravinha) and their subsequent pelitic-carbonate sequences were partially involved in the Brasiliano tectonics of all the marginal belts surrounding the São Francisco Craton, in like manner to the Rio Preto and Sergipano belts (Table 4, Fig. 7). Some glacial sediments of the same age have been reported on the African side of the MCO-Trans-Saharan Belt (Caby and Fabre, 1981), and again there remains the possibility that some of them were preserved in the northern domains of this province. Indeed, there are some candidates in the diamictite beds and banded iron formation units of the Seridó and Piancó belts.

After about 750 Ma the paleogeography of this region underwent major modification with the development of several continental proto-oceanic and oceanic basins, following widespread taphrogenesis. This led to the development of a series of concurrent zones of tectonic uplift as massifs, basement inliers, and tectonic highs that may correlate with the general breakup of Rodinia. These highs were the sites where Transamazonian and Cariris Velhos lithostructures (Granja-Senador Sá in the MCO; Tróia-Tauá in the CC, Rio Piranhas/RP, São Jose do Campestre/JC, and minor parts of AP+AM+ RC, PEAL) continued to be exposed providing the sedimentary source for the developing basins as well as those in which sedimentation was starting a new cycle.

Some of the Neoproterozoic basins were formed in part along previous structural trends of the Paleoproterozoic/Transamazonian belts (Ubajara, Seridó) and Cariris Velhos structural trends (Piancó-Alto Brígida). But others like the Rio Preto and Sergipana basins were aligned obliquely to previous tectonic features and do not seem to reflect any fundamental structural inheritance.

It is reasonable to expect some kind of paleogeographic link between these basins, mainly between those situated to the SE of the Senador Pompeu Lineament, covering the RGND basement, and there are some stratigraphic, isotope, geochemical and structural reasons favouring this point of view. In addition, there are some possible sites for triple junctions in the N of Sobradinho, BA and W of Granjeiro, CE. Although, for several reasons, the definition of this *paleogeographic framework is a problem that is still beyond* of the present level of geological knowledge, some preliminary inferences may be made, including:

a) The Sergipano continental margin, trending NW-SE (Fig.7), with the deposition of continental sediments on the craton in the S to the deposition of deep-sea sediments, and the development of oceanic environments in the N. This belt continued eastwards along the northern edge of the Congo Craton;

b) The SW-NE trending Rio Preto-Riacho do Pontal Belt (Table 4) was probably a continental margin with sedimentological polarity to the NW. However, irregular morphological features of the cratonic basement including the gneiss-migmatite highs of the Boqueirão and Estreito ranges complicate its definition. Furthermore, good indications for the development of oceanic crust have yet not been found. This marginal belt may have been continuous with the Sergipano-Central African Belt, all of which together define the northern limit of the São Francisco Craton.

c) The Pharusian/Dahomeyan belts (Affaton, 1990) define the continental margin along the eastern side of the West Africa Craton. On the Brasiliano side, the Ubajara Belt is the only likely counterpart; and it may represent the depositional relicts of a larger assemblage similar to that found on the African side.

d) The Piancó-Alto Brígida Basin, N of the Cariris Velhos Domain in TZ (N of Riacho Gravatá Belt) was a continental rift that evolved to an oceanic basin of unknown width. It may have formed above the Cariris Velhos structures as a Brasiliano fore-arc basin. It is bounded on the S by the Serra do Caboclo Shear Zone, and the source areas of the sediments were the AP and AM of the TZ. The northern limit and continental foreland was the RGND, but this boundary now lies along the extensively re-worked western side of the Patos Lineament. This belt may have continued into the Rio Preto continental margin, following a series of paleophysiographic accidents, although subsequent deformation caused by the Pernambuco Lineament has obscured these original links. In fact, it is also possible that the apparent alignment of these belts is fortuitous and the result of the Brasiliano deformation as a whole.

e) The Seridó Basin and associated basin remnants including Quimani-São Jose Piranhas, Lavras, Capiu, were developed within the RGND (Fig. 8). It may have started as a continental rift that has locally evolved in the subsequent thermal regime to an ensialic basin of attenuated continental crust, probably including a small proto-oceanic basin (Jardim de Sá, 1994). Recent studies on detrital zircon of this basin and on the Piancó-Alto Brígida Basin suggest that these basins may have once been contiguous, although each has different attributes and metamorphic-structural histories.

Late Neoproterozoic to Cambrian (Brasiliano Orogeny)

Rio Grande do Norte Domain

In the previous item some of the general characteristics of the Neoproterozoic rock assemblages of this Province have been briefly mentioned. In the case of the so-called pre-Brasiliano assemblages, it is necessary to reiterate that the late Paleoproterozoic rocks of the Jaguaribeano Belt got their present structural and metamorphic facies during the Brasiliano.

A stratigraphic overview for the lithostratigraphic pile of the Martinopole Group is shown in Table 5, where there is a predominance of phyllite and metafelsic rocks.

For the volcano-sedimentary sequences such as those of Rio Curu-Independência, Guaramiranga and Canindé reliable stratigraphic data are not available. Psammo-pelitic associations are predominant, with variable intercalations of calc-silicate and carbonate rocks and minor amounts of volcanic rock (Arthaud *et al.*, 1998). They present a great homogeneity in their stratigraphy as well as in their tectonothermal conditions, including amphiolite facies metamorphism with local anatexis and a general structural pattern of SW vergent thrusts and nappes.

The continental margins basins display a similar lithostratigraphic framework (diamictite + QPC rock assemblages), probable bulk chronostratigraphic correlation and their stratigraphical successions are akin to the classical models. For their marginal succession there are no major stratigraphic problems. Nevertheless, additional studies are required to refine the stratigraphy and structural relationships. There remain a number of problems such as the degree of participation of lithological units of earlier cycles such as the Cariris Velhos and Transamazonian, and the relative quantities of volcanic and plutonic components involved.

Piancó-Alto Brígida Belt

The Piancó-Alto Brígida Basin consists of a predominance of turbiditic sequences, psammo-pelitic rock assemblages with some local occurrences of felsic (andesitic to rhyodacitic and rhyolitic group, lava and tuff, peralkaline and peraluminous) and mafic magmatism (basalt, gabbro) suggestive of a magmatic arc environment (Sá *et al.*, 1998). The southern border of the basin, along the Serra do Caboclo Shear Zone, contains coarse-grained turbidite beds, conglomerate and breccia with volcanic pebbles, typical of those derived from the adjacent Cariris Velhos Terrane (Riacho Gravatá and AP). Rhythmic sedimentary features are very common in these low-grade rocks that appear to have been deposited mainly in deep-water environments. There are some paleogeographical indications for a basin with a previous and short history of continental rift that quickly evolved to a marine basin and subsequently became the site of the magmatic arc. Some local occurrences of mafic-ultramafic magmatism (Bodocó, PE) have been related (Beurlen *et al.*, 1992) to remnants of oceanic crust. Other evidence for the previous existence of a oceanic crust is indirectly obtained from the general features (geochemistry, petrology, I-type epidote-bearing plutons) of the plutonic rocks, a remarkable characteristic of this belt (Ferreira *et al.*, 1998) and of its continuation to the Rio Preto area to the S of the Patos Lineament.

Some flysch lithological associations, magmatism (Bittar, 1998) as well as ongoing isotope studies suggest that at least part of the sedimentation is syn-orogenic, *c.* 630 to 620 Ma. This is to say that it is coeval with the first phase of plutonism and deformation. This does not preclude the possibility of the presence of older Neoproterozoic successions such as those of Cryogenian age.

The Seridó Belt presents a sedimentological and tectonic history that is better schematized in time. The Jucurutu Formation with its paragneiss and calc-silicate rocks is mainly

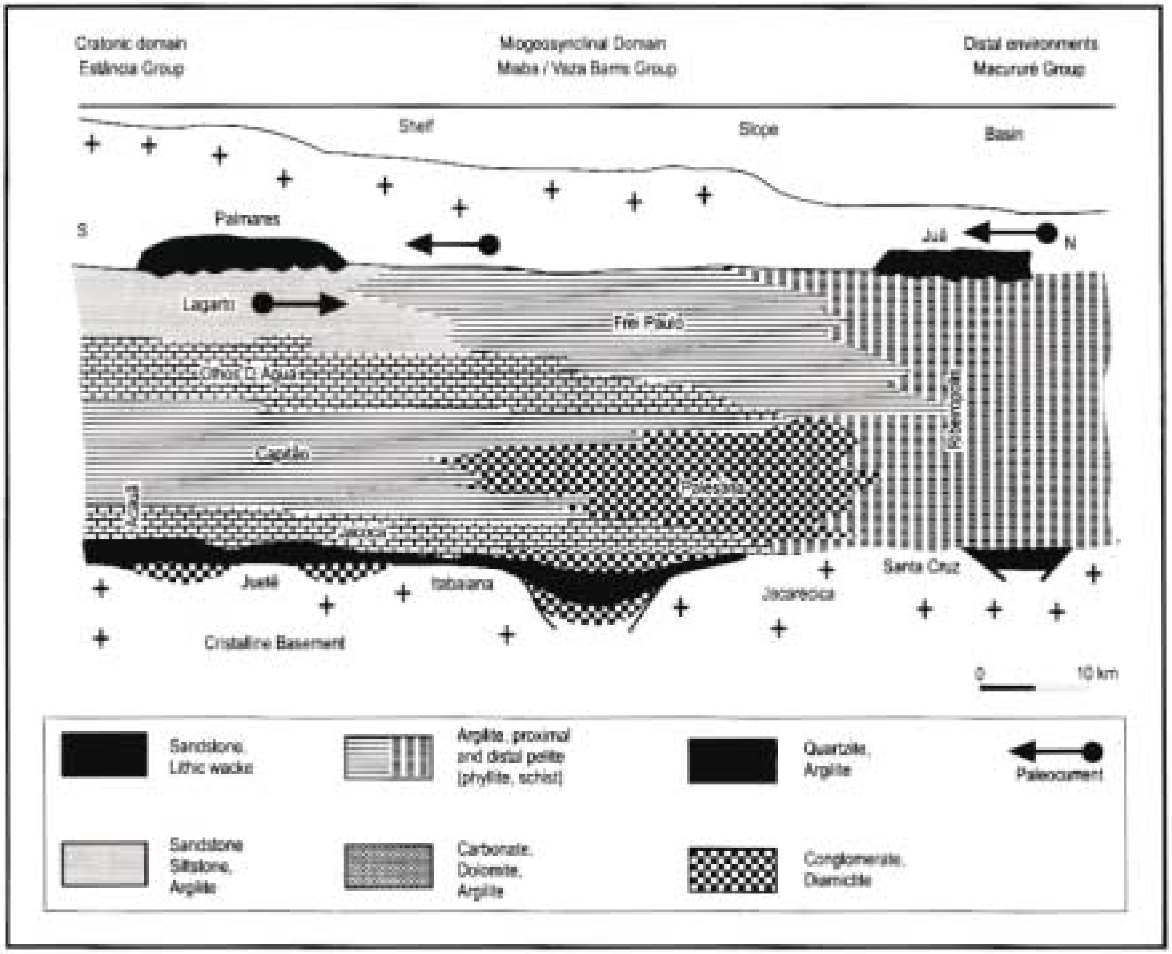


FIGURE 7 - Lithostratigraphic interpretation for the Sergipano Fold Belt (with the outline of its original environments), from the foreland (São Francisco Craton) to the deep marine basin (to the N). Modified after D'El Rey and McClay (1998).

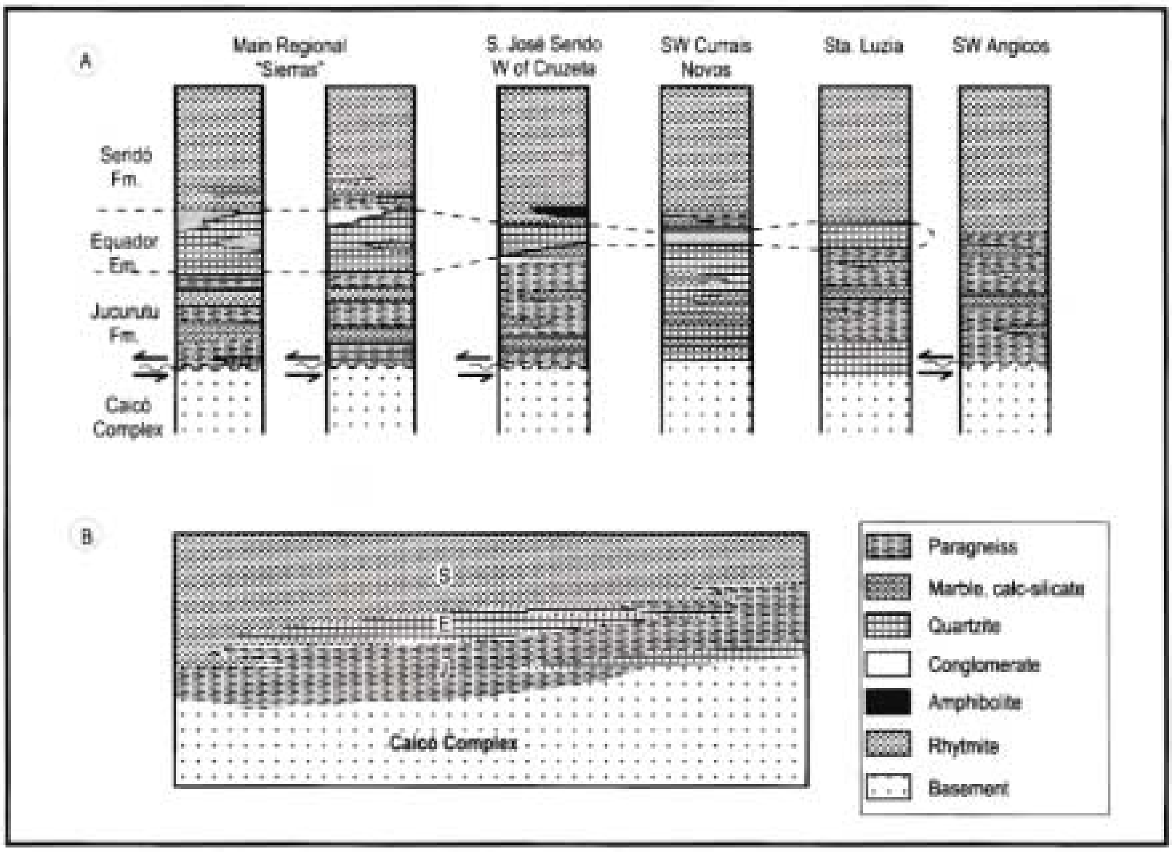


FIGURE 8 - Stratigraphic fence diagram (A) for the different sectors of the Seridó Fold Belt (SED) and a concluding interpretation (B). After Jardim de Sá (1994).

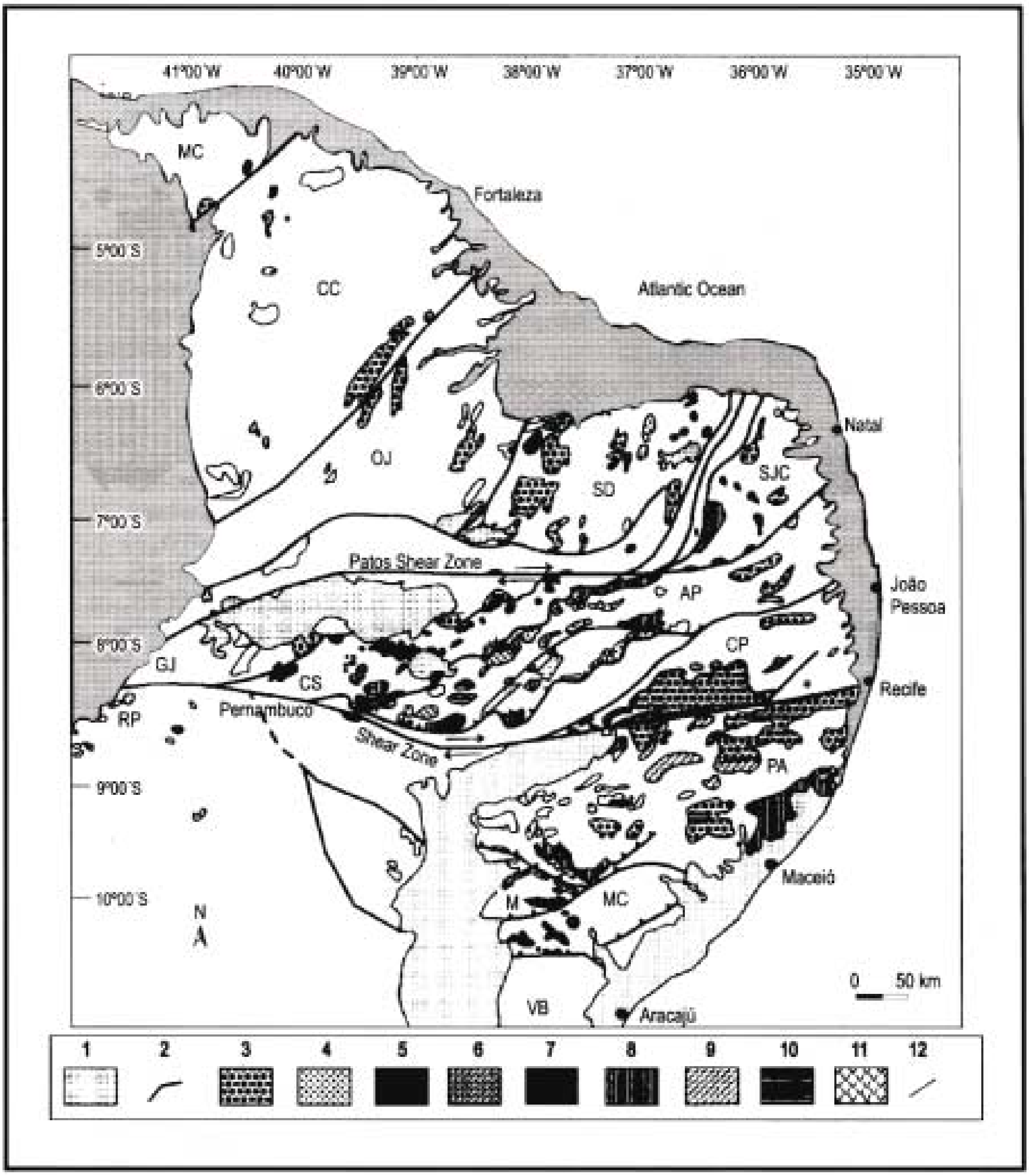


FIGURE 9 - Nomenclature for the main Brasiliano granitic plutons and suites:
1 - Meruoca, 2 - Mocambo, 3 - Serra da Barriga, 4 - Quixeramobim, 5 - Pereiro-São Miguel, 6 - Umarizal, 7 - Acari, 8 - Barcelona, 9 - Monte das Gameleiras, 10 - Serra da Lagoinha, 11 - Itaporanga, 12 - Conceição, 13 - Serrita, 14 - Salgueiro and Terra Nova, 15 - Triunfo, 16 - Solidão and Tabira, 17 - Teixeira, 18 - Catingueira, 19 - Esperança, 20 - Campina Grande-Serra Redonda, 21 - Prata, 22 - Fazenda Nova, 23 - Rajada, 24 - Serra da Esperança, 25 - Ouro Branco, 26 - Tará, 27 - Xingó, 28 - Coronel João Sá.
Super-suite I : 4, 5, 7, 8, 10, 11, 12, 13, 14, 23, 25, 26, 27, 28
Controversial: 18, 24
Supersuite II: 15, 16, 17, 19, 20, 22.
Supersuite III: 1, 2, 3, 6, 9, 21

a QPC sequence, and the volcanic rocks (Jardim de Sá, 1994) are indicating an extensional tectonic regime and continental rifting. The source area for the Jucurutu rocks varies considerably, even for rocks of middle Archean age, with contribution from all of the components of the Rio Piranhas foreland, including younger source-rocks of Neoproterozoic age, *c.* 650 Ma. The Equador Formation has some degree of association with the underlying Jucurutu Formation (Santos and Brito Neves, 1984) and it is also another QPC sequence, preceding a short hiatus and the deposition with widening of the original basin of the Seridó Formation.

Seridó Belt

The Seridó Formation with its predominantly psammo-pelitic and some local greywacke sequences (Fig. 8) shows the general characteristics including stratigraphy (flysch deposits), structural features (syn-depositional tectonic episodes), volcanism and isotope data (younger detrital zircon *c.* 647 Ma), suggesting that sedimentation, of the uppermost units at least, was coeval with the first phases of deformation of the whole basin. As the Seridó Formation shows metamorphic zoning with high-grade metamorphites to the E, and very low grade metamorphic facies (Cruzeta, RN) to the W it is possible to observe some similarities between these stratigraphic sections and those of the Piancó-Alto Brígida Belt. Recent isotope (SHRIMP, U/Pb) data show that partial stratigraphic correlation may be possible, and that some kind of link between these basins may have occurred, although they are the product of different evolutionary histories and classifications. Otherwise, it will be necessary to suggest a land bridge between the Seridó Basin (to the N) and the same source areas for the Piancó Basin (AP + AM) to the S. Considering the geochronological data of 590 - 580 Ma for the oldest late-tectonic plutons such as Acari, in the central part of the belt, according to Souza *et al.* (1998), the Seridó Belt developed over a period of some 70 Ma during the Neoproterozoic III.

Global correlation of the Borborema Brasiliano belts are to be found in several textbooks as well as in modern maps showing the reconstruction of Gondwana, although they are often not as precise as might be desirable. It is necessary to remember that the Brasiliano is a collective name for several different orogenic epochs. Thus, the Brasiliano orogenic system of the Borborema Province, to which there is not a special local name, has to be correlated only with those orogenies occurring in the time-span *c.* 630 - 570 ma. There are several examples on this continent as well as in Africa, of both older and younger orogenic systems also considered to be Brasiliano and occurring in the wide time-span between 900 and 500Ma. Actually, all these systems are parts of the same wide and general process of orogenic collage that were responsible for the accretion of Western Gondwana.

The Granitic Plutonism

Extensive and diverse granitic plutons were emplaced during the Brasiliano, and this is one of the main characteristics of the Borborema Province. This diversity is a reflection of the complex tectonic framework and heterogeneity of crustal features of this Province and of the overall hybrid character of the magmas there formed. A recent review on this subject was written by Ferreira *et al.* (1998), that distinguished about nine different petrogenetic groups. Studies carried out by Guimarães *et al.* (1998), in the ZT, show five distinct groups of granitoid plutons, corresponding to different tectonic events, ages and types of protolith.

The nomenclature applied to the granitoid plutons of this region is based mainly on the pioneer work of Almeida *et al.* (1967), which defined the main petrographic types. This was followed and amplified by Sial (1984, 1987) and other authors. Naturally, consensus about this complex subject is far from being reached. In this review we will use the concept of supersuites, following the recent proposal of Santos and Medeiros (1998), but avoiding discussion on the position of the granitoid plutons in the orogenic cycle, because of the complexity of the problem and the lack of the necessary multidisciplinary support. Three supersuites are distinguished, even allowing for the possibility that some transitional types may occur.

I - Hybrid and crustal suites. These are related to the more important magmatic episodes affecting this province, with pluton emplacement since the early phases of contractional tectonics up to the late phase of strike-slip displacements (Archanjo *et al.*, 1994; Jardim de Sá, 1994).

II - Enriched-mantle derived suites. This includes syn-kynematic and late-kynematic intrusives of the major strike-slip events, frequently emplaced along deep crustal structures, apparently corresponding to deep crustal discontinuities.

III - Hybrid suites. These are minor intrusive granites and dyke sets related to the post-closure uplift and the collapse phase of the orogenic structures.

Supersuite I

The supersuite I includes the following suites: two calc-alkaline, a trondhjemitic and a peraluminous suite. The high-K calc-alkaline suite, the most widespread in Borborema Province, shows some differences in mineralogical and geochemical composition of the plutons of different terranes (Ferreira *et al.*, 1998). It consists of large bodies of a megacrystic K-feldspar granite of the Itaporanga type, in addition to diorite and a mixing phase between them. This dioritic phase resembles the precursor and syn-plutonic basic pulses of calc-alkaline suites of continental magmatic arcs.

The epidote-bearing calc-alkaline suite (Conceição type) consists of tonalite and granodiorite with minor diorite and gabbro phases, occurring as high-level plutons, mainly in the PAB and in the northwestern part of the Po. They display typical rounded microgranular enclaves, mafic double enclaves and actinolite-rich clots, the latter being considered as fragments of a previous basaltic source (Sial, 1993).

According to Santos and Medeiros (1998), there is a relationship between the calc-alkaline suites. The Itaporanga-type shows similarities with the complex processes observed in the MASH zone (magma assimilation, segregation and homogenization of Pearce, 1996). Here the injection of mantle-derived basic magma at the base of the crust would produce a series of basic and felsic magmas, derived from basic magma and

TTNTH is sometimes included in the above-described supersuite. But according Guimarães *et al.* (1998), Teixeira, Solidão (SW of Teixeira in the APT) and the Serra Branca batholiths (SE of Teixeira, in the APT) are considered as transitional shoshonitic-alkaline types. They are composed of leucocratic pyroxene syenogranite to syenite, also including quatz-monzonite and quartz-monzodiorite (Teixeira and Solidão) and leucocratic biotite syenogranite (Serra Branca). They include rare mafic enclaves and large amounts of inherited zircon. U/Pb zircon ages of *c.* 570 Ma were obtained by Guimarães *et al.* (1988) and Kozuch (unpublished thesis).

Supersuite III

The main region of occurrence of this supersuite is the Sobral (Transbrasiliano Lineament) Groaíras-Tauá (along the MCO-CC boundary and in the western CC) extensional fault system. There occur the Meruoca, Mocambo and Serra da Barriga plutons and other sets of dykes of the same age (Sial, 1989, unpublished thesis). The Meruoca Pluton consists of two facies: a fayalite-bearing granitic facies, and a Fe-rich biotite alkali-feldspar granitic facies; and locally, a quartz syenite with aegirine or riebeckite. The Mocambo Batholith consists of granodiorite, granite and quartz syenite with hormblende and biotite. A set of dykes of latite, dacite, rhyolite and quartz diorite occurs around these plutons. U/Pb ages around 532 Ma were obtained, which is consistent with Rb/Sr data obtained previously by Fetter *et al.* (1997). Melting of amphibole and biotite in the lower crust is suggested for this plutonism.

A magnetite-bearing fayalite and iron-hyperstene alkaline granite occur in Umarizal, in the western part of the RP (between the Portalegre Shear Zone and Frutuoso Gomes Shear Zone). According Galindo *et al.* (1995), it shows a Rb/Sr age of *c.* 545 Ma.

Some stocks in the southern part of the JC (eastern RGND), such as like Dona Ignês and Monte das Gameleiras can be assigned to this supersuite.

In the TZ, there occur some post-tectonic granite plutons (Prata, A-type suite), of batholithic dimensions, stocks and dykes that are apparently controlled by a late brittle tectonic episode of shearing. The Prata Complex is formed by syenogranite, comagmatic with basalt and dacite (Melo *et al.*, 1996; Guimarães *et al.*, 1998). They are metaluminous sub-alkaline to slightly alkaline rocks, less depleted in Nb than other calc-alkaline suites of this province, being considered as representative of intraplate granite. Their Nd model ages suggest an Archean crustal source for this suite.

Summary and global implications

Any approach to describe the history of the Borborema Province must start with the Archean events; which rock-types crop out in the majority of the exposures of the basement of this province, in addition to isotope indicators in Archean rocks elsewhere. About 1 Ga of Archean history are somehow recorded in this province, even though all the known Archean fragments played only a limited role as local components of some Paleoproterozoic mobile belt. The importance of the Brasiliano Cycle is reflected in many different ways including the composition, formation, shape, and the reworking of older rock assemblages, in addition to general tectonics and predominant structural trends. The final shape was the result of late Neoproterozoic (collision) and Cambrian post-collisional (strike-slip) movements related to escape tectonics. Secondly, the Borborema Province owes its evolution to such phenomena as rift and drift and some transform displacements of the Equatorial and Central Atlantic Ocean, and the consequent formation of the Coastal Province.

Regarding the development of the Borborema Province it is necessary to note that it only represents about one third of a complete Wilson Cycle (Brasiliano/Pan African Cycle), in terms of extent, space and time. The complete geodynamic history of this cycle includes divergent processes that began as early as 950 Ma in central-western Brazil, as well as convergent processes with arc magmatism as early as 900 - 850 Ma, also in the central-western and also in southern Brazil. These processes have diachronously continued from one province to another, with some orogenic belts only closing their histories in the Early Ordovician period. The history of the Borborema Province has to be understood in these terms.

From the middle to the late Paleoproterozoic, during and after the processes of the Transamazonian collage, the area of the Borborema Province was situated in the remote interior of the Atlantica Supercontinent (Rogers, 1996). Probably the basement of this province was then representing the continuity of the structures of the São Francisco-Congo Craton situated to the S, with the São Luis-Parnaíba-West Africa Craton to the N and NW. A hypothetical supercontinent provides an almost ideal way of containing in a single landmass the remarkable group/set of the accretionary and collisional fold belts built and coalesced by the Transamazonian orogeneses. So well represented they are that they shear common characteristics in Brazil and in Africa.

Cartoon a: Atlantica, showing the three areas of common Paleoproterozoic basement (S. Luis West Africa + Borborema + São Francisco Congo).

A series of important events of extensional tectonics affected the Atlantica Supercontinent in middle Paleoproterozoic times, as a global phenomenon. This tectonism was accompanied by the formation of many volcano-sedimentary basins, volcanic traps and flood basalt flows, which are well represented all over the South American Continent. Of special importance are those in the São Francisco (Western Chapada Diamantina, Espinhaço), Parnaíba and Borborema provinces. Generally, these extensional tectonic movements did not develop rates that were sufficiently large to bring about the formation of oceanic realms, the Amazonas Basin being an exception. In the Borborema Province, these rock assemblages formed during the Stathefian taphrogenesis and are present mainly in the area of the Jaguaribeano Fold Belt. However, some scattered occurrences of plutonism with gabbro anorthosite and alkaline granite have been described elsewhere. Volcano-sedimentary sequences may have developed in other fold belts, but they will be difficult to define due to the superimposition of later basins and structures during the Cariris Velhos and Brasiliano cycles.

The volcano-sedimentary sequences of the Jaguaribeano Fold Belt are composed mainly of psammo-pelitic rocks and felsic magmatism, some local calc-silicate units. They remained for more than 1.0 Ga as cratonic cover, not deformed or barely deformed units up until the start of orogenic transpressional processes of the Brasiliano Cycle, existing no evidence for the Cariris Velhos Orogeny in that area.

Cartoon b :Generation of extensional structures in late Paleoproterozoic times reaching the areas of São Francisco + Borborema + São Luis-West Africa provinces.

From the late Mesoproterozoic to the very early Neoproterozoic (<1.05 Ga) the above discriminated part of Atlantica was once again affected by new important phase of extensional tectonics, with the formation of many rift basins, where thick volcano-sedimentary sequences were deposited and some anorogenic plutonism. This extensional regime evolved up to the drift stage, with the separation of different lithosphere fragments and the formation of oceanic floor. The volcanic contribution was varied. This reached major proportions in some belts (Riacho Gravatá, Piancó Alto Brígida), while in others it was modest or even absent (Sertania and São Caetano formations; Pajeú-Paraíba Fold Belt). These rock assemblages varied according with the development of the tectonic realm, since BVAC types (bimodal volcanics, arkose, and conglomerate) in rift systems to QPC (quartzite, pelite, carbonate) in continental margins. In so doing, there are some local occurrences of mafic rocks of tholeitic affinities (Pedra Preta, Floresta, Custódia), good candidates for dismembered ophiolite sequences, all of them deeply deformed in the Cairis Velhos Cycle and during the subsequent Brasiliano orogenesis. Evidence for arc-related volcanism (Lagoa das Contendas) and plutonism (Floresta area) developed during the phases of convergence (*c.* 970 - 950 Ma), confirming the presence of deep-sea sediments and oceanic floor.

Cartoon c: The late Mesoproterozoic-early Neoproterozoic extensional regime with the basin formation tectonics of the Cariris Velhos Cycle, accompanied by pronounced rifting and drifting (and associated rocks) that led to the consolidation of the Atlantica Supercontinent. It also led to the first signs of the development of two major segments. The CC+ RGN to the N, and the São Francisco-Congo block/plate to the S.

The major fragments, descendants of Atlantica, have interacted from the beginnings of the Neoproterozoic (early Tonian), some dozen of millions of years after consolidation. These convergent interaction, including subduction, collision and transpression were responsible for the formation of the Cariris Velhos Fold Belt (Pajeú-Paraíba and related systems) and the individualization of their main terranes like the APT, AMT, and RCT.

This development of the Cariris Velhos orogenic system at *c.* 970 - 950 Ma with the closing of continental and oceanic basins has provide the central part of the province with a thickened and heated crustal zone, with some branched systems (to the SW, Riacho do Pontal, to SE, Macururé-Marranco), which was very important for the development of the subsequent cycle and the whole history of the province.

The Cariris Velhos plutonism is outstanding and the leuco-orthogneiss formed during this cycle, as dozens of sheets, stocks and plutons, are distributed along a belt 800 km long from southeastern Piauí to the SE of the State of Rio Grande do Norte. The petrological and geochemical data are still the subject of ongoing research, but it is possible that part of this set of meta-plutonic rocks was somehow connected with subduction or some of them could be part of the former extensional regime. The western part of the granite-migmatitic PEAL Massif has been of substantial importance for the granitization processes of Cariris Velhos times, not yet wholly known and understood.

The evolution of the Cariris Velhos Orogeny is not only a provincial fact, but it is part of a global phenomenon, for it has to be included among those classical groups of sutures of the Grenville collage that were responsible for the consolidation of Rodinia Supercontinent.

Cartoon c: The early Neoproterozoic outline of the Cariris Velhos Orogenic Belt positioned between the northern domains (Ceará Central + Rio Grande do Norte domains) and southern domains (SD + São Francisco-Congo). Additionally, is outlined a minor fragment of the Transamazonian Belt (Eastern part of the PEAL) to the E. All these crustal fragments represent only a small area inside Rodinia . Some coeval extensional processes (dyke swarms) have been recorded in the southern domain.

During the middle (*c.* 850 - 750 Ma) and the late Neoproterozoic (660 - 620 Ma) , at least two major groups of taphrogenetic processes affected the above-mentioned part of Rodinia. Once again such processes were participants in a wide and diachronic global phenomenon that led to the dispersion of the previously consolidated supercontinent. These were processes were very important for the Borborema region, with the individualization of several lithosphere segments of different size, a characteristic of the Brasiliano tectogenesis, and continental and oceanic basins. Such processes were widespread in that part of Rodinia in question (cartoon d).

In the Southern Domain, the periphery of the São Francisco-Congo Plate begins to assume an outline with the formation of an Atlantic-type continental margin (Rio Preto, Riacho Pontal, Sergipana-Mbalmayo/Oubanguides), there including some types of oceanic environment. In the eastern part the São Luis-West Africa Craton was then being defined with identical formation of continental margins (Pharusian, Dahomeyan).

In the interior of the province several basins were formed, without preferential sites relating the nature of the basement. The rates of basin-forming tectonics governed subsequent environments and only locally the conditions that permitted the development of oceanic basin were attained. Only some of these oceanic basins are already confirmed, most of them are conjectural and involve the hypotheses about wide oceans traversing this portion of consolidated Rodinia. However, this hypothesis does not have the support of the tectonic and volcano-sedimentary record. Psammo-pelitic rocks, rhytmite beds having varying degrees of maturity and felsic volcanics are the predominant rock assemblage in most of the interior basins/belts. Coeval with the first phase of extension there is evidence for glacial (Sturtian) deposits and cap dolomite, especially in the Southern Domain

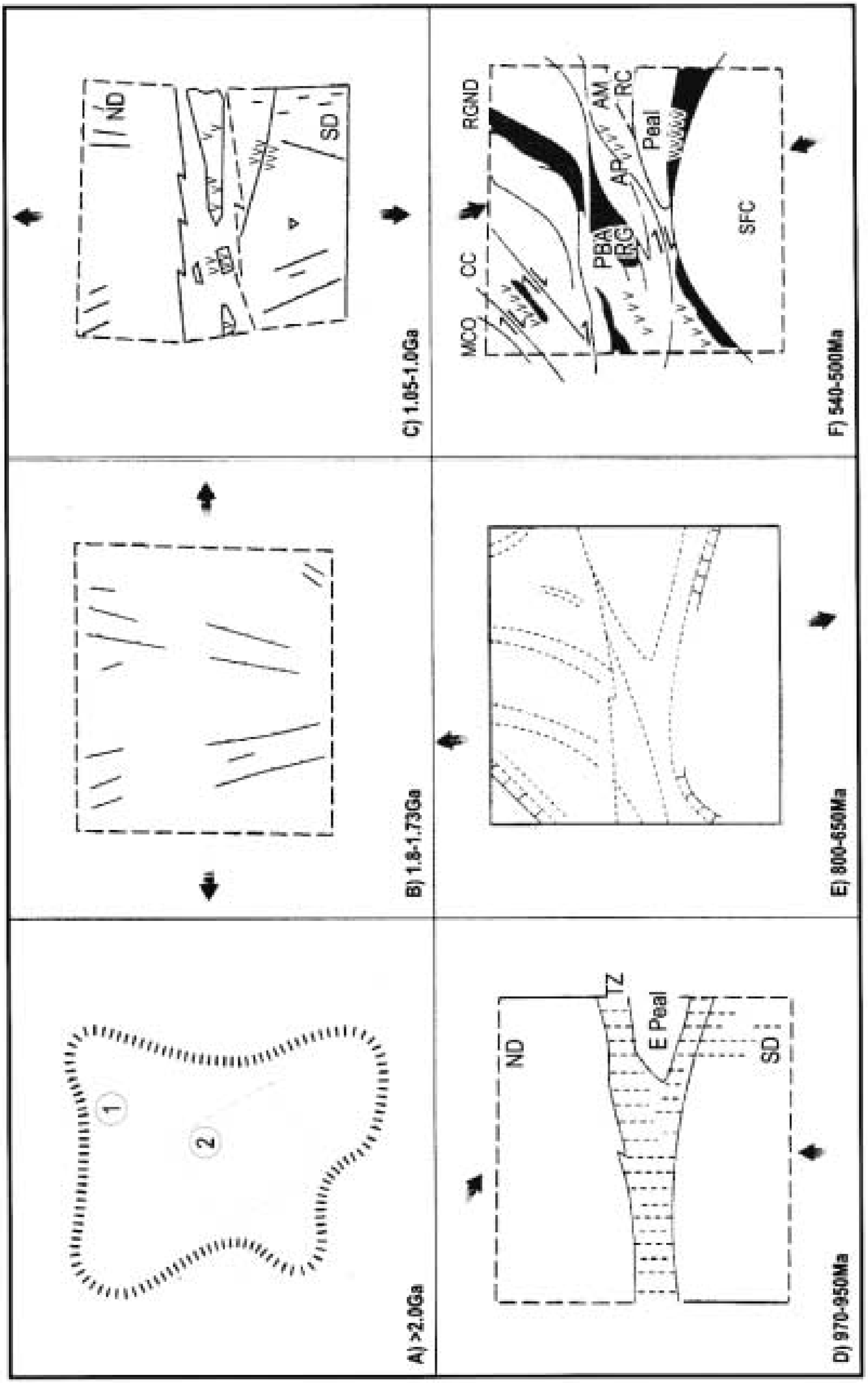


FIGURE 10 - Cartoons illustrating the main steps of evolution of the Borborema Province.
a - Atlantica supercontinent (l) and the area for the further evolution of the Borborema and surrouding provinces (2). b - The Statherian Taphrogenesis. Rift sedimentation, anorogenic volcanism and plutonism. c - The Atlantica break-up and the formation of the main realms for the Cariris Velhos Orogeny, rifts and marine basins. In this way, two major continental segments were defined: the northern (ND) and the southern (SD). d - The Cariris Velhos Orogeny caused by the interaction between the ND and SD segments. Final events related with the fusion of Rodinia Supercontinent (c. 950 Ma). e - Extensional events (820 - 750, <650 Ma) related with the Brasiliano Cycle. Development of continental margins (symbols for limestone) and interior volcanic-sedimentary (and oceanic) depositional sites (broken lines). f - The latest structural evolution of the Province, after the Brasiliano Orogeny (640 - 600 Ma) and the later extrusional activities. Final shaping of domains and terranes that took place during Early Ordovician times.

Carton e: Dense systems of rifts and related basins (some of with oceanic environment) created by the extensional processes of the middle and late Neoproterozoic. The basins/belts in the central part may have received some tectonic inheritance from the Cariris Velhos trends.

Cartoon f: The convergent events of the late Neoproterozoic were diachronic from belt to belt, leading to intense folding that maintained distinctive characteristics from one belt to another, but which terminated over the whole province between 630 - 610 Ma. However, our records on this subject are poor and/or obscured by the important granitic plutonism then developed. Granitic plutonism is the essential characteristic of the Borborema Province. This range of age is typical for most of the main deformational phase and syn-tectonic plutonsim. Late tectonic plutonism is also widespread with age range *c.* 580 Ma . A third and important phase of plutonism (with associated tectonism), granite intrusives with alkaline affinities (+fissural intrusives + hydrotermal veining) occurred over a period of approximately 50 Ma, from 550 to 500 Ma, from one place to another of the province, in accordance with local circumstances. This was especially true during the development of extensional, post-collisional movements along the main lineaments (inter and intra-domains of the province) for which there are several group of ages (several methods), some from 580 Ma, but with climax between 540 and 500 Ma.

These lateral movements of escape tectonics were very important, and they gave the last structural shape and arrangement to the basement of the province. This network of shear belts, which was just one of the tools used to identify domains and terranes, had considerably modified many of the earlier structural trends of the Brasiliano event, and even of the former orogenic cycles (Cariris Velhos, Transamazonian). So, all exercises on reconstitution of the previous scenarios, tectonic and paleogeographical are difficult and usually at risk. Furthermore, slight modifications in this general shape and framework then acquired were the result of Mesozoic continental drift. However, until this began, the influence of the Brasiliano trends, and particularly of the lineaments, was remarkable.

Cartoon g: The Brasiliano Borborema Province; the juxtaposition (mostly Cambrian in age) of different domains and terranes by the net work of lineaments, resulting in the final shape of the fold belts (Brasiliano and Cariris Velhos). However, by the end of this time the external outlines of the São Francisco Craton (in the S) and the S. Luis Craton (in the NW) started to take shape.

Table 1. Major chronostratigraphic events of the Borborema Province

AGE	DESCRIPTION
Cambrian	
545 - 500 Ma	Post-tectonic plutons; strike-slip faulting
Late Neoproterozoic (Brasiliano orogeny *sensu strictu*)	
c. 580 Ma	Late Brasiliano plutonism
610 - 600 Ma	Main Brasiliano deformation and metamorphism
640 - 620 Ma	Early Brasiliano magmatism and sedimentation
Middle Neoproterozoic (Pre-Brasiliano supracrustal rocks)	
850 - 790 Ma	Supracrustal suites (break-up of Rodinia?)
Early Neoproterozoic (Cariris Velhos Orogeny *sensu strictu*)	
1.0 Ga - 950 Ma	Plutonism, volcanism, sedimentation
Mesoproterozoic (Stasis)	
(1.7 - 1.1 Ga)	(Little to no record)
Late Paleoproterozoic (Intracratonic events: Intra-Atlantica?)	
1.8 - 1.75 Ga	Supracrustal suites
Middle Paleoproterozoic (Transamazonian Cycle *sensu lato*)	
2.1 - 2.0 Ga	Late Transamazonian plutonism and metamorphism
2.2 - 2.1 Ga	Main Transamazonian crustal formation
Early Paleoproterozoic (Pre-Transamazonian crustal fomation)	
2.4 - 2.3 Ga	Basement of Middle Coreau Domain
Archean crustal formation events	
2.8 - 2.5 Ga	Gneissic basement of Pre-Transamazonian Archean nuclei
3.2 - 3.0 Ga	Gneissic basement of Pre-Transamazonian Archean nuclei
3.5 - 3.4 Ga	TTG Gneissic basement of Pre-Transamazonian Archean nuclei
Obs: Chronology based primarily on U-Pb ages of zircon and monazite	

TABLE 2. Lithostratigraphic Units of the Transversal Zone

"TERRANES"		Piancó-Alto Brígida (PAB)	Alto Pajeú (APT)	Alto Moxotó (AMT)	Rio Capibaribe (RCT)
NEOPROTEROZOIC 1.0Ga	BRASILIANO OROGENY (540-570Ma)	Enriched mantle granitic supersuite II ++ Hybrid granitic supersuite I Cachoeirinha Group Santana dos Garrotes Formation OO Serra do Olho D'Água Formation ●●●●● ●●●●●●●●● Fernandes Vieira Shear Zone	Enriched mantle granitic supersuite II A-type post-orogenic granitic supersuite III ++ Hybrid and crustal granitic supersuite I vv Irajaí Complex (720 Ma) (2) Riacho da Barreira Complex Prece do Rodrigues Unit ΛΛ +	Caroalina Formation OOO Jaramantaia Unit (2)	Enriched mantle granitic supersuite II Hybrid and crustal granitic supersuite I Surubim Formation OOO (2) +
MESOPROTEROZOIC 1.6Ga	CARIRIS VELHOS EVENT (1.0Ga-950Ma)	Riacho Gravatá Complex (1.2 Ga)	XX Crustal collisional granitic suite (1.03 Ga) São Caetano Complex (~1.09Ga) Serrote das Pedras Pretas Suite (1) Serra de Jabitacá Nappe (cryptic suture?)	XX Crustal collisional granitic suite(~1.0Ga) vv Lagoa das Contendas complex (~1.01Ga) ** Sertânia Complex Cruzeiro do Nordeste/Congo Shear Zone	XX Crustal collisional granitic suite Vertentes Complex Serra de Taquaritinga A-Type granite (1.5 Ga) (1)
PALEOPROTEROZOIC "ARCHEAN"		Paleoproterozoic orthogneiss (2.0 Ga) (1)	Archean and Paleoproterozoic orthogneisses (1)	Gabbro-anorthositic Suite Floresta Granulitic Complex (2,15Ga) Caldas Brandão Leucogranitoids(3.1Ga)	Gabbro-anorthositic Suite (1,7Ga) Paleoproterozoic Orthogneisses (1)

+ Brasiliano granite
v Volcanic arc sequence
Turbidite sequence
OO Psephite and rudite
Rift sequence
Platformal sequence
Λ Mafic-ultramafic rocks
High pressure mafic-ultramafic rocks
Greywacke-volcaniclastic sequence
Undiscriminated volcano-sedimentary sequence
X Metagranitoid
* Anorogenic granite
Anorogenic gabbro-anorthositic suite
Orthognessic basement
●●● Stratigraphic unconformity
(1) Allochtonous contact related to the Cariris Velhos (1) and Brasiliano (2) episodes
Possibility of age change
Allochtoneity

Note: The indicated ages are U/Pb concordia in zircon, except for that labelled as (*), dated by Pb/Pb dissolution method.

TABLE 3. Lithostratigraphy of the Sergipano Belt

Lagarto/Tobias Barreto Foredeep (São Francisco Craton)	Sergipano Belt (S) and Itabaiana Dome (Proximal)	Distals Facies (N)
Palmares Fm.: Greywacke, sandstone. Local conglomerates and breccias	Juá\Serra do Cágado Fm.: (Intradeeps) Polymitic conglomerate, arkose Intrusive granite and granodiorite.	
Estancia Group Lagarto Fm.: Siltstone, mudstone, green and red sandstone. Acauá Fm.: Meta limestone, meta-dolomite, siltstone, mudstone. Juetê Fm. Meta limestone, meta-dolomite, siltstone, mudstone.	Vasa-Barris Group Frei Paulo - Ribeirópolis Fm.: Philyte, meta-turbidite, slate, micashist, meta-graywacke. Olhos D'água Fm.: Meta-limestone, calc-phylite, phyllite, meta-siltstone. Capitão-Palestine Fm.: Philyte, slate, Diamictite, greywacke (Patestina). Miaba Group Jococa Fm.: Meta-limestone, slate, mudstone. Jacarecia Fm. (=Ribeirópolis ?): Meta- graywacke with blocks, meta-mudstone and meta-siltstone. Itabaiana Fm.: Quartzite, meta-siltstone, local conglomerate.	Macururé Group Meta-rhytmite. Banded micashist. Intercalations of: Quartzite, Arkose, Calc-silicatic, Orthogneiss. Basal quartztite (Santa Cruz)

After Silva Filho et al, 1978 (modified)

TABLE 4. Rio Preto - Riacho do Pontal lithostratigraphy

Rio Preto Fold Belt		Riacho Pontal Fold Belt
Cratonic Domain	Pericratonic Domain	Distal Domain / Allochtonous Distal Domain
S. Desidério Fm. (= Sete Lagoas, L. Jacaré, Salitre Fm.) ? ? ? ? Canabravinha Fm.	Riacho Neves Fm. Greenish meta-arkose, siltstones, feldspatic sandstone, Limestone lenses. Serra da Mamona Fm. Meta-marls, fine-grained sandstone, siltstone S. Desiderio Fm. Bluish limestone and marl, Siltstone. Local oolitic limestone. Canabravinha Fm. Feldspatic quartzite, schist Local diamictite	Casa Nova Complex Phyllites, calc-phyllite ? Vargem Grande - Barra do Bonito ? ? ? ? (Monte Orebe Un.) Volcano-Sedimentary Amphibolite, actinolite-tremolite schist, meta-chert, tuff, breccia garnet-biotite schist. (Mandacaru Un.) Deep sea sediments Two-mica schist, quartzite lenses, meta-turbidite, meta-greywacke, sheets of two mica granite (Rajada orthogneiss). (Barra Bonita Un.) Terrigeneous shelf sediments Garnet-muscovite-biotite schist lenses of limestone and quartzite. Sheets of orthogneiss. (Rajada tonalite and syenogranite)
Espinhaço Supergroup: Rio Preto Group	Rio Preto Group* Quartzite, schist (Paleo-Mesoproterozoic?) High-grade Gneisses (Paleoproterozoic?) (*Occuring in the distal domains, reworked)	Espinhaço Supergroup: Chapada Diamantina Group (>1.3 Ga)
High-grade Gneiss (Paleoproterozoic)	High-grade Gneiss (Paleoproterozoic)	High-grade Gneiss and Transamazonian Belts (Paleoproterozoic) High-grade Gneiss: (Archean) TTG suites, migmatite, local quartzite, calc-silicate, mafic-ultramafic bodies.
From: Silva, M.E., 1987 Andrade Filho, *et al.* 1994		From: Angelim, 1988 Santos and Silva Filho, 1990

TABLE 5. Medio Coreaú - Ceará Central lithostratigraphy

Médio Coreaú (MCo) Domain		Ceará(Cc) Domain
Jaibaras Group	Aprazível Fm.: Polimictic conglomerate, coarse-grained matrix Parapuí Volcanism: Basalt, Andesite, Rhyolite, Volcanic Breccia Massapê-Pacujá Fm.: Breccias, conglomerate, Polimictic rock units. Sandstone, siltstone	Ceará Group a) Independência - Rio Curú Metapelite Marble Intercalations of quartzite and felsic volcanics b) Guasmiringa - Guanindé Metapelite Quartzite Intercalations of carbonate rocks C) Quixeramobim Metapelite Quartzite Carbonate intercalations
Martinópole Group Santa Terezinha Fm.: (São José Group) Phyllite, Quartzite,Schist, Metarhyolite Marble, local diamictite. Covão Fm.: Biotite-quartz schist, muscovite–schist with kyanite. Comon retrograde facies. São Joaquim Fm.: Quartzite with kyanite, staurolite, sillimanite. Calc-silicatic intercalations. Goiabeiras Fm.: Staurolite-garnet schist, kyanite schist. Muscovite schist.	Ubajara Group Coreau Fm.: Arcose, greywacke, sandstone with conglomeratic intercalations, basal siltstone Frecheirinha Fm.: Black and bluish limestone. Intercalation of phyllite and marl. Trapiá Fm.: Mica-bearing sandstone, siltstone, phyllite. Caiçaras Fm.: Slates, sandstone, siltstone Local conglomerate	
Paleoproterozoic Basement (High-grade Gneiss, Kinzigite)		Paleoproterozoic Basement Local Archean Nuclei

REFERENCES

Abouchamy, W., Boher, M., Michard, A. and Albared, F. (1990). A major 2.1 Ga event of mafic magmatism in west-Africa: an early stage of crustal accretion. *Journal of Geophysical Research*, **95**, 17605-17629.

Affaton, P. (1990). *Le Bassin des Volta (Afyique de L'Ouest): une marge passsive, d'age Proterozoique Supérieur, tectonisée au panafricain (600 ± 50 Ma).* Thesis. Faculté des Sciences et Techniques, Université d'Aix Marseille III, France, 2v., 309 p. (unpublished).

Almeida, F.F.M. and Hasui, Y., coords. (1984). *O Pré-Cambriano do Brasil.* Edgard Blücher Ltda., São Paulo, Brasil, 378 p.

Almeida, F.F.M., Hasui, Y., Brito Neves, B.B. and Fuck, R.A. (1977). Provincias Estruturais Brasileiras. *Atas, VIII Simpósio de Geologia do Nordeste.* Campina Grande, 363-391.

Almeida, F.F.M., Hasui, Y., Brito Neves, B.B. and Fuck, P.A. (1981). Brazilian Structural Provinces: An introduction. *Earth-Science Reviews*, **17**, 1-19.

Almeida, F.F.M., Leonardos, O.H. and Valença, J. (1967). Review on Granitic Rocks of Northeast South America. *Simposium on Granites, IUGS/UNESCO*, Recife-PE, 41 p. (Special Publication).

Andrade Filho, E., Neves, J.P. and Guimarães, J. (1994). *Programa Levantamentos Geológicos Básicos do Brasil. Santa Rita de Cássia, Formosa do Rio Preto. Folhas SC.23-ZC e SC.23-YD. Estados da Bahia e Tocantins. Escala 1:250.000.* MME/CPRM, Brasília, Special Publication, 68 p.

Angelim, L.A.A. (1988). *Projeto Levantamentos Geológicos Básicos do Brasil. Santa Filomena. Folha SC.24-V-A-III- Escala 1:100.000. Estados de Pernambuco e Piauí.* MME/CPRM. Brasília, Special Publication, 146 p.

Archanjo, C.J., Bouchez, J.L., Corsini, M. and Vauchez, A. (1994). The Pombal granite pluton: magnetic fabric, emplacement and relationships with the Brasiliano strike-slip setting of NE Brazil (Paraiba State). *Journal of Structural Geology*, **16**, 323-335.

Arndt, N.T. and Goldstein, S.L. (1987). Use and abuse of crust-formation ages. *Geology*, **15**, 893-895.

Arthaud, J.M., Vasconcelos, A.M. and Oliveira, F.V.C. (1998). As Seqüências Metassedimentares do Ceará Central. *Anais XL Congresso Brasileiro de Geologia, Belo Horizonte-MG*, p. 16

Barbarin, B. (1997). From the mineralogical composition of granitoids to their origins and tectonic settings. In: *International Symposium on Granites and Associated Mineralizations, ISGAM, 2. Salvador-BA. Extended Abstract and Program*, 95-98.

Barbarin, B. (1996). Genesis of the two main types of peraluminous granitoids. *Geology*, **24**(2), 295-298.

Beurlen, H. (1988). *Fazenda Esperança (Bodocó) e Riacho da Posse (Floresta): duas ocorrências atípicas de Fe-Ti no Estado de Pernambuco.* Thesis (Full Professor thesis). Universidade Federal de Pernambuco, Recife-PE, 72 p. (unpublished).

Beurlen, H., Silva Filho, A., Guimarães, I.P. and Brito, S. (1992). Proterozoic C-type eclogites hosting ununsual Ti-Fe± Cr ± CU mineralization in northeastern Brazil. *Precambrian Research*, **58**, 195-214.

Bittar, S.M.B. (1998). *Faixa Piancó-Alto Brígida: Terrenos tectono-estratigráficos sob regimes metamórficos e deformacionais contrastantes.* Thesis. Universidade de São Paulo, São Paulo, 126 p. (unpublished).

Bober, M., Abouchamy, W., Michard, A., Albarade, F. and Arndt, N.T. (1992). Crustal growth in west Africa at 2.1 Ga. *Journal of Geophysical Research*, **97**, 345-369.

Brito Neves, B.B. (1975). *Regionalização Tectônica do Precambriano Nordestino.* Thesis, Universidade de São Paulo, São Paulo, 198 p. (unpublished).

Brito Neves, B.B. (1983). *O Mapa Geológico do Nordeste Oriental do Brasil, Escala 1:1.000.000.* Thesis (Livre docência). Universidade de São Paulo, São Paulo, 177 p. (unpublished).

Brito Neves, B.B., Sá, J.M., Nilson, A.A. and Botelho, N.F. (1995a). A Tafrogênese Estateriana nos blocos Paleoproterozóicos da América do Sul e processes subseqüentes. *Geonomos*, **3**(2), 1-21

Brito Neves, B.B., Van Schmus, W.R., Santos, E.J., Campos Neto, M.C.C. and Kozuch, M. (1995b). O Evento Cariris Velhos na Província Borborema. Integração de dados, implicações e perspectivas. *Revista Brasileira de Geociências*, **25**(4), 279-296.

Brito Neves, B.B. and Cordani, U.G. (1991). Tectonic Evolution of South America during the Late Proterozoic. *Precambrian Research*, **53**(112): 23-40

Brown, M. and Solar, G.S. (1998). Granite ascent and emplacement during contracional deformationin convergent orogens. *Journal of Structural Geology*, **20**(9/10), 1365-1393.

Caby, R. and Arthaud, M. (1986). Major Precambrian Nappes of the Brazilian belt, Ceará, Northeast Brazil. *Geology*, **14**, 871-874

Caby, R. and Fabre, (1981). Late Proterozoic to Early Paleozoic diamictites, tillites and associated glaciogenic sediments in the Sé Pourprée of Western Hoogar, Algeria. In: *Earth's pre-Pleistocene glacial record*, eds. Hambrey, M.J. and Harland, W.B., pp.140-145. Cambridge University Press.

Campos Neto, M.C., Bittar, S.M.B. and Brito Neves, B.B. (1994). Domínio Tectônico Rio Pajeú - Província Borborema - Orogêneses superpostas no Ciclo Brasiliano/ Pan-Africano. *Boletim de Resumos Expandidos, XXXVIII Congresso Brasileiro de Geologia, Balneário Camboriú-SC*, **1**, 221-222.

Cavalcanti, J.C., Sá, J.M. and Macedo, M.H.F. (1998). A delimitação da Faixa Jaguaribeana: Novos Dados Geocronológicos. *Anais XL Congresso Brasileiro de Geologia, Belo Horizonte-MG*, p. 11

Cavalcanti, J.C., Medeiros, W., Castro, D. and Sá, J.M. (1998). A Faixa Jaguaribeana (parcial): Integração Geológico-gravimétrica. *Anais XL Congresso Brasileiro de Geologia, Belo Horizonte-MG*, p. 13

Cavalcanti, J.C. (1999). *Limites e Evolução Geodinâmica do sistema Jaguaribeano, Província Borborema, Nordeste do Brasil.* Thesis, Universidade Federal do Rio Grande do Norte, Natal-RN, 169 p. (unpublished).

Corsini, M., Lambert de Figueiredo, L., Caby, R., Féraud, G., Ruffet, G. and Vauchez, A. (1998). Thermal history of the Pan-African/ Brasiliano Borborema province of Northeast Brazil deduced from Ar^{40}/Ar^{39} analysis. *Tectonophysics*, **285**, 103-117.

Coutinho, M.G.N. and Alberton, D.H.M. (1998). Proterozoic Lode Gold Deposits in the Borborema Province, Northeast Brazil and their Exploration Significance. *Anais da Academia Brasileira de Ciências*, 10 (3), 429-439.

Dada, S. (1998). Crust-forming ages and Proterozoic crustal evolution in Nigeria: a reappraisal of current interpretations. *Precambrian Research*, **87**, 65-74.

Dantas, E.L. (1992). *Evolução tectono-magmática do maciço polidiapírico São Vicente-Florânia, RN.* Master thesis, Universidade Estadual Paulista-UNESP, Rio Claro-SP, 272 p. (unpublished).

Dantas, E.L. (1997). *Geocronologia U/Pb e Sm/Nd de terrenos Arqueanos e Paleoproterozóicos do Maciço Caldas Brandão, NE Brasil.* Thesis, Universidade Estadual Paulista-UNESP, Rio Claro-SP, 201 p. (unpublished).

Dantas, E.L., Hackspacher, P.C., Van Schmus, W.R. and Brito Neves, B.B. (1998). Archean accretion in the São José do Campestre Massif, Northeast Brazil. *Revista Brasileira de Geociências*, **28**(2), 221-228.

DePaolo, D.J. (1981). A Neodymium and Strontium isotope study of the Mesozoic calc-akaline granitic batholiths of the Sierra Nevada and peninsular Ranges, California. *Journal of Geophysical*

Research, **86**, 10470-10488.

Dia, A., Van Schmus, W.R. and Kröner, A. (1997). Isotopic constraints on the age and formation of a paleoproterozoic volcanic arc complex in the Kedougou Inlier, eastern Senegal, West Africa. *Journal of African Earth Sciences*, **24**, 197-213.

Fernandes, T.M.G. (1997). *Estudos Geológicos e Geocronológicos Complementares da parte sul do Maciço Caldas Brandão-PB.* Master thesis. Universidade Estadual Paulista-UNESP. Rio Claro-SP, 106 p. (unpublished).

Ferreira, V.P. (1991). *Petrology and geochemistry of the Late Proterozoic ultrapotassic peralkaline Triunfo Batholith and related dikes, State of Pernambuco, Northeast Brazil.* Thesis, University of Georgia, Athens, 269 p. (unpublished).

Ferreira, V.P., Sial, A.N. and Jardim de Sá, E.F. (1998). Geochemical and isotope signatures of Proterozoic granitoids in terranes of the Borborema structural province, Northeast Brazil. *Journal of South American Earth Sciences*, **11**(5), 439-455.

Fetter, A.H., Van Schmus, W.R., Santos, T.J.S., Arthaud, M. and Nogueira Neto, J. (1997). Geocronologia e Estruturação do Estado do Ceará: NW da Província Borborema, NE Brasil. In: SBG/Núcleo Nordeste, Simpósio de Geologia do Nordeste, 17. Fortaleza -CE, *Resumos Expandidos, Boletim*, **15**, 3233.

Fetter, AH. (1999). *Geochronological evolution of the Ceará State-Borborema Province-Northeast Brazil.* Thesis. University of Kansas, Lawrence, Kansas, 150 p. (unpublished).

Galindo, A.C., Dall'Agnol, R. MeReath, I. Lafon, J.M. and Teixeira, N. (1995). Evolution of Brasiliano-age granitoid types in a shear zone environment, Umarizal-Caraúbas region, Rio Grande do Norte, northeast Brazil. *Journal of South American Earth Sciences*, **8**, 79-95

Guimarães, I. P. 1989. *The petrological evolution and tectonic associations of the Bom Jardim Complex Pernambuco State, NE Brazil.* Thesis. University of London/Imperial College, London, 426 p. (unpublished).

Guimarães, I.P. and Silva Filho, A. (1998). Nd- and Sr- Isotopic and U-Pb Geochronologic constraints for the Evolution of the shoshonitic Brasiliano Bom Jardim and Toritama Complexes, Evidences for a Transamazonian Enriched Mantle Under Borborema Tectonic Province, Brazil. *International Geology Review*, **40**, 500-527.

Guimarães, I.P. and Silva Filho, A.F. (1995). Shoshonitic granitoids of the Borborema Province, NE, Brazil. *Abstracts, The III Hutton Symposium on the origin of granites and related rocks, Washington DC USA*, 61-63.

Guimarães, I.P., Silva Filho, A.F., Almeida, C.N., Araújo, J.M.M., Sales, A. and Melo, S.C. (1998). The Brasiliano granitoids from the Pajeú-Paraíba Belt and Teixeira High: Sm-Nd Isotope Geochemistry and U/Pb in zircon ages. *Anais XL Congresso Brasileiro de Geologia, Belo Horizonte-MG*, p. 48.

Guimarães, I.P., Silva Filho, A.F., Almeida, C.N., Melo, E.B., Melo, S.C., Araújo, J.M.M. and Sales, A. (1999). Sm-Nd isotope geochemistry and U-Pb zircon ages of the Brasiliano granitoids from the Pajeú-Paraíba Terrain, Borborema Province, Northeast Brazil. *Actas II South American Symposium on Isotope Geology-SAGI, Cordoba, Argentina*, 203-205.

Gusmão, R. (1998). *Arcabouço Geotectônico da região da Faixa Riacho do Pontal, Nordeste do Brasil. Dados aeromagnéticos e gravimétricos.* Master thesis, Universidade de São Paulo, São Paulo-SP, 157 p. (unpublished).

Hackspacher, P.C., Van Schmus, W.R. and Dantas, E.L. (1990). Um embasamento Transamazônico na Província Borborema. *Anais, XXXVI Congresso Brasileiro de Geologia, Natal-RN*, **6**, 2683-2696.

Hackspacher, P.C., Dantas, E.L., Brito Neves, B.B. and Legrand, J.M. (1997). Northwestern Overthrusting and Related Lateral Escape during the Brasiliano Orogeny North of the Patos Lineament, Borborema Province, Northeast Brazil. *International Geology Review*, **39**, 609-620.

Howell, D.G. (1995). *Principles of Terrane Analysis. New Application for Global Tectonics.* Chapman & Hall, London, U.K., 245 p.

Jardim de Sá, E.F. (1994). *A Faixa Seridó (Província Borborema, Nordeste do Brasil) e seu significado geodinâmico na Cadeia Brasiliano - Pan-Africana.* Thesis. Universidade de Brasília. Brasília, 803 p. (unpublished).

Jardim de Sá, E.F., Macedo, M.H.F., Kawashita, K., Peucat, J.J., Leterrier, J. and Fuck, R.A. (1996). A suite Serra da Esperança: intrusões alcalinas sintectônicas aos nappes brasilianos na Faixa Riacho do Pontal, NE do Brasil. *Anais, XXXIX Congresso Brasileiro de Geologia, Salvador-Ba*, **7**, 499-501.

Jardim de Sá, E.F., Trindade, R.I.F., Hollanda, M.H.B.M., Araújo, J.M.M. Galindo, A.C., Amaro, V.E., Souza, Z., Vigneresse, J.L. and Lardeaux, J.M. (1999). Brasiliano Syntectonic Alkaline Granites Emplaced in a strike slip/extensional Setting. *Anais da Academia Brasileira de Ciências*, **71**(1), 17-28.

Kozuch, M., Bittar, S.M.B., Van Schmus, W.R. and Brito Neves, B.B. (1997). Late Mesoproterozoic and Middle Neoproterozoic magmatism in the Zona Transversal of Borborema Province, NE-Brazil. In: SBG/Núcleo Nordeste, Simpósio de Geologia do Nordeste, 17. Fortaleza-CE. *Resumos Expandidos, Boletim*, **15**, 47-49.

Ledru, P., Johan, V., Milisi, J.P. and Tegyey, M. (1994). Markers of the last stages of the Paleoproterozoic collision: evidence for a 2 Ga continent involving circum-South Atlantic Provinces. *Precambrian Research*, **69**, 169-191.

Leite, P.R.B. and Lima, E.S. (1997). Caracteristicas geoquimicas dos granitóides proterozóicos da Folha Monteiro, terreno Alto Pajeú, nordeste do *Brasil. Resumos Expandidos, XVII Simpósio de Geologia do Nordeste, Fortaleza*-CE, 282-287.

Martins, G., Oliveira, E.P., Souza Filho, C.R. and Lafon, J.M. (1998). Geochemistry and Geochronology of the Algodões Sequence, Ceará, NE Brazil, A Paleoproterozoic Magmatic Arc in the Central Ceará Domain of the Borborema Province? *Anais, XL Congresso Brasileiro de Geologia, Belo Horizonte-MG*, p. 28.

Melo, S.C., Guimarães, I.P., Silva Filho, A.F., Macambira, M.J.B. and Lima, E.S.L. (1996). Geochemical and Rb-Sr data of the Neoproterozoic bimodal anorogenic magmatism in the Borborema Province, NE Brazil: the Prata Complex. *Resumos Expandidos, XXXIX Congresso Brasileiro de Geologia, Balneário Camboriú-SC*, **6**, 607-610.

Monié, P. Caby, R. and Arthaud, M. (1997). The Neoproterozoic Brasiliano orogeny in Northeast Brazil: $^{40}Ar/^{39}Ar$ and petrostructural data from Ceará. *Precambrian Research*, **81**, 241-264.

Parente, C.V. and Arthaud, M.H. (1995). O Sistema Orós-Jaguaribe no Ceará-NE *Brazil. Revista Brasileira de Geociências*, **25**(4), 297-305.

Parente, C.V. Guillou, J.J. and Artahud, M.-H. (1998). Geologia e Geoquimica dos Elementos Maiores dos Depósitos de Magnesita Pré-Cambriana (*c*.1.8 Ga) da Faixa Móvel Orós (*Ceará*). *Revista Brasileira de Geociências*, **28**(4), 439-448.

Pearce, J. (1996). Sources and Settings of Granitic Rocks. *Episodes*, **19**(4), 120-125.

Pitcher, W. 1993. Granites: Tipology, geological environment, and melting relationships. In: *Migmatites, melting and ultrametamorphism*, eds. Atherton, M.P. and Gribble, C.D., pp. 277-285. Shiva Publication, Cheshire.

Porada, H. (1989). Pan African Rifting and orogenesis in southern to Equatorial Africa and Eastern Brazil. *Precambrian Research*, **44**, 103-136.

Powell, H. (1995). *Principles of Terrane Analysis. New Apilication for Global Tectonics.* London, Chapman & Hall, 245 p.

Reagan, P.F. (1995). The Early Basic Intrusions. In: *Magmatism at* *a plate edge: The Peruvian Andes*, eds. Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D., pp. 72-89. Black & Sun Ltd.

Rogers, J.J.W. (1996). A History of continents in the past three billion years. *The Journal of Geology*, **104**, 91-107.

Sá, J.M. (1991). *Evolution Geodynamique de La ceinture Noterozoique d'Oros, Nord-Est Brésil.* Faculté de Sciences, Université de Nancy 1. Thesis. Nancy, 177 p. (unpublished).

Sá, J.M., Macedo, M.H.F. and Fernandes, N.R. (1998). Ambiente Tectônico da parte oriental da Sequência Salgueiro-Cachoeirinha, Provincia Borborema. *Anais XL Congresso Brasileiro de Geologia, Belo Horizonte-MG*, p. 15.

Sá, J.M., Bertrand, J.M. and Leterrier, J. (1997). Geocronologia U-Pb e geoquimica de ortognaisses Paleo e mesoproterozóicos da região de Taquaritinga-*PE. Resumos Expandidos, XVIII Simpósio de Geologia do Nordeste, Fortaleza-CE*, 108-112.

Saabaté, P., Barbosa, J.S.F. Marinho, M., Martin, H. and Teixeira, L. (1997). Granitóides in the northeastern São Francisco Craton, NE Brazil, and continental growth from 3.4 to 1.9 Ga. *Extended Abstract and Program, II International Symposium on Granites and Associated Mineralizations, Salvador-Ba*, 235-237.

Sampaio, T.Q. and Vasconcelos, A.M. (1991). *Programa Levantamentos Geológicos Básicos do Brasil. Afranio Folha SC.24-V-A-V. Estados de Pernambuco, Piauí e Bahia.* MME/CPRM. Brasília, Special Publication, 144 p.

Santos, E.J. (1995). *O Complexo Granitico Lagoa das Pedras: Acresção e Colisão na Região de Floresta (Pernambuco), Provincia Borborema.* Thesis. Universidade de São Paulo, São Paulo, 219 p. (unpublished).

Santos, E.J. (1996). Ensaio Preliminar sobre Terrenos e Tectônica Acrescionária na Provincia Borborema. Anais, XXXIX *Congresso Brasileiro de Geologia. Salvador-BA*, **6**, 47-50.

Santos, E.J. and Brito Neves, B.B. (1993). Petrogenesis and Tectonic Setting of the Lagoa das Pedras Magmatism, floresta, State of Pernambuco, Borborema Province, Northeast Brazil. *Anais da Academia Brasileira de Ciências*, **65** (Suplemento 1), 131-139.

Santos, E.J. and Brito Neves, B.B. de (1984). Provincia Borborema. In: *O Pré-Cambriano do Brasil*, eds. Almeida, F.F.M. de and Hasui, Y., pp. 121-186. Edgard Blücher Ltda.

Santos, E.J. dos, Coutinho, M.G.N, Costa, M.P.A. and Ramalho, R. (1984). A Região de Dobramentos Nordeste e a Bacia do Parnaiba, incluindo o cráton de São Luis e as Bacias Marginais. In: *Geologia do Brasil*, eds. Schobbenhaus, C., Campos, D.A., Derze, G.R., Asmus, H.E., pp. 131-189. Ministério das Minas e Energia-Departamento Nacional da Produção Mineral.

Santos, E.J., Gusmão, R.G. and Paiva, I.P. (1997). Terrenos no dominio transversal da Provincia Borborema, controles sobre acresção e retrabalbamento crustais ao sul do Lineamento Patos. In: *SBG/Núcleo Nordeste, Simpósio de Geologia do Nordeste, 17. Fortaleza-CE. Resumos Expandidos, Boletim*, **15**, 140-144.

Santos, E.J. and Medeiros, V.C. (1998). New insights on Grenville-age and Brasiliano Granitic Plutonism of the Transverse Zone, Borborema Province, NE Brazil. *Actas X Congresso Lafinoamericano de Geologia, y VI Congreso National de Geologia Econômica, Mar del Plata, Uruguay*, **2**, 427-431.

Santos, C.A. and Silva Filho, M.A. (1990). *Programa Levantamentos Geológicos Básicos do Brasil. Folha Riacho do Caboclo - SC.24-VA-VI. Estado de Pernambuco. Escala 1:100.000.* MME/CPRM, Brasília. Special Publication, 113 p.

Santos, T. (1999). *Evolução Tectônica e Geocronológica do Extremo Noroeste da Provincia Borborema.* Thesis, Universidade Estadual Paulista - UNESP, Rio Claro-SP, 186 p. (unpublished).

Scheid, C. and Ferreira, C.A. (1993). *Programa de Levantamentos Geológicos básicos do Brasil, Juazeirinho - Folha SR24-Z-D-II Estado da Paraíba. Escala 1:100 000.* MME/CPRM/DNPM, Brasília. Special Publication, 164 p.

Sengör, A.M.C. (1990). Plate Tectonics and Orogenic Research after 25 vears, A tethyan perspective. *Earth-Science Reviews*, **27**, 1-201.

Sial, A.N. (1986). Granite-types in Northeast Brazil: Current Knowledge. *Revista Brasileira de Geociências*, 16(1), 54-72.

Sial, A.N. (1990). Epidote-bearing calc-alkalic granitoids in Northeast Brazil. *Revista Brasileira de Geociências*, **20**(1-4), 88-100.

Sial, A.N. (1987). Granitic Rocks of Northeast Brazil. *Annals, International symposium on granites and Associate mineralizations, Salvador-BA*, 61-69.

Sial, A.N. (1984). Litogeoquímica de Elementos Terras Raras na caracterização dos granitóides do Espaço Cachoeirinha. *Anais, XXXIII Congresso Brasileiro de Geologia, Rio de Janeiro-RJ*, 2697-2709.

Sial, A.N. (1993). Contrasting metaluminous magmatic epidote-bearing granitic suites from two Precambrian fold belts in Northeast Brazil. *Anais do Academia Brasileira de Ciências*, **65**(1), 141-162.

Silva, M.E. (1987). *O Sistema de dobramentos Rio Pardo e suas relações com o Cráton do São Francisco.* Thesis, Universidade de São Paulo, São Paulo, 135 p. (unpublished).

Silva Filho, A.F., Van Schmus, W.R., Guimarães, I.P. and Luna, E.B. (1997). The usage of Sm-Nd isotope signatures as a tool for defining sub-domains in the southern tectonic domain of the Borborema Province. Atas, *17 Simpósio de Geologia do Nordeste, Fortaleza- CE*, 128-131.

Silva Filho, A.F., Guimarães, I.P. and Van Schmus, W.R. (1999). The crustal evolution of the PE-AL terrane, Borborema Province, NE Brazil, based on the Sm-Nd data of their Proterozoic Granitic Batholiths. *Actas II South American Symposium on Isotope Geology, Cordoba-Argentina*, 300-301.

Souza, L.C., Hunziker, J., Verkaeren, J. and Legrand, J.M. (1998). A Termocronologia $^{40}Ar/^{39}Ar$ e a história termal do Maciço Granítico de Acati-RN durante pelo menos 98 Ma. *Anais XL Congresso Brasileiro de Geologia, Belo Horizonte-MG*, p. 14.

Teixeira, W., Noce, C.M., Quéméneur, J.J.G. and Martins, V.T.S. (1997). Sr, Nd and Pb isotope signatures of intrusive granitoids of the Paleoproterozoic Mineiro Magmatic Ar, southern São Francisco Craton, Brazil. *Extended Abstract and Program, II International Symposium on granites and Associated Mineralizations, Salvador-BA*, 288-289.

Toteu, S.F., Van Schmus, W.R., Penaye, J. and Nyobé, J.B. (1994). U-Pb and Sm-Nd evidence for eburnian and Pan-African high-grade metamorphism in cratonic rocks of southern Cameroon. *Precambrian Research*, **67**, 321-347.

Trompette, R. (1994). *Geology of Western Gondwana (200-500 Ma) Pan African-Brasiliano Aggregation of South America and Africa.* Bakelma, Rotterdam/ Brookrield, 350 p.

Unrug, R. (1996). The Assembly of Gondwanaland. *Episodes*, **19**(1/2), 11-20.

Van Schmus, W.R., Brito Neves, B.B., Hackspacher, P.C., Fetter, A.H., Kozuch, M. Dantas, E.L. and Babinski, M. (1998). The Borborema Province: A collage of Polycyclic Domains in Northeast Brazil. In: *International Conference on Precambyian Craton Tectonics, 14. Ouro Preto-MG. Abstracts*, 80-83.

Van Schmus, W.R., Brito Neves, B.B., Hacspacher, P.C., Babinski, M., Fetter, A.H. and Dantas, E.L. (1995a). Neoproterozoic and Late Mesoproterozoic sedimentary and volcanic sequences in the Borborema province, NE Brazil. In: *SBG/Núcleo Nordeste, Simpósio de Geologia do Nordeste, 14. Recife-PE, Atas, Boletim*, **14**(2), 391-393.

Van Schmus, W.R., Dantas, E.L., Fetter, A., Brito Neves, B.B., Hackspacher, P.C. and Babinski, M. (1996). Neoproterozoic age for Seridó Group, NE Broborema Province, Brazil. *Anais, XXXIX Congresso Brasileiro de Geologia, Salvador-BA*, **6**, 152-155.

Van Schmus, W.R., Brito Neves, B.B., Hackspacher, P.C. and Babinski, M. (1995b). U-Pb and Sm-Nd geochronologic studies of the Eastern Borborema Province, Northeast Brazil: initial conclusions. *Journal of South American Earth Sciences*, 8(3/4), 267-288.

Vauchez, A., Neves, S.P., Caby, R., Corsini, M., Egydio-Silva, M., Arthaud, M. and Amaro, V. (1995). The Borborema Shear Zone System. *Journal of South American Earth Sciences*, 8(314), 247-266.

Veiga Junior, J.P. and Ferreira, C.A. (1990). *Programa Levantamentos geológicos Básicos do Brasil, carta geológica, carta metalogenética e previsional Folha SR24-Z-C-VI, Afogados da Ingazeira. Estados de Pernambuco e da Paraiba. Escala* 1:100 000. MME/CPRM/DNPM. Brasília, special publication, 121 p.

Wanderley, A.A. (1990). *Programa levantamentos Geológicos Básicos do Brasil, carta geológica, carta metalogenética e previsional Folha SB.24-Z-D-IV Monteiro. Estados de Pernambuco e Paraiba. Escala 1:100000*. MME/CPRM/DNPM. Brasília, special publication, 100 p.

Weil, A.B., Van Der Voo, R., Mac Niocaill, C. and Meert, J.M. (1998). The Proterozoic supercontinent Rodinia: paleomagnetically derived reconstructions for 1100 to 800 Ma. *Earth and Planetary Science Letters*, 154, 13-24.

Windley, B. (1995). *The Evolving Continents*. John Wiley & Sons, New York, NY, USA, 526 p.

Windley, B. (1998). The Tectonic Development and Continental Reconstruction of Precambrian Orogens and Cratons. *Abstracts XIV International Conference on Basement Tectonics, Ouro Preto-MG*, p. 1.

Paraguay and Araguaia Belts

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Orogenic belts of the Neoproterozoic Brasiliano-Pan-African Cycle are one of the most important features of Precambrian shield regions of Brazil. In the western and central parts of the Country, Neoproterozoic orogenic belts developed on the border of the Amazonian Craton, being originally called the Paraguai-Araguaia Belt (Almeida, 1967). Silva *et al.* (1974) and Almeida (1974, 1984) suggested the sub-division of this fold belt in two units. The Araguaia Belt (or Araguaia-Tocantins) bordering the eastern edge of the Amazonian Craton, and the Paraguay Belt along the southeastern border of the craton. Sediments of the Bananal Basin cover the contact between these two fold belts, and a large portion of these belts lie below Paleozoic-Mesozoic sedimentary rocks (Parnaíba, Parecis and Paraná basins) and Cenozoic sedimentary sequences (Pantanal and Bananal basins) (Fig. 1).

The Paraguay Fold Belt

The Paraguay Belt and its corresponding cratonic cover were deposited on the southern margin of the Amazonian Craton/Rio Apa Block (Figs. 1 and 2). Deformation in the belt increased from almost imperceptible on the border of the craton to tight and isoclinal folds within the compressional belt (Alvarenga, 1990; Alvarenga and Trompette, 1993). This belt represents a young Brasiliano tectonic unit (Trompette, 1994; Pimentel *et al.*, 1996) formed by Vendian to Early Cambrian sediments, with deformation between 550 and 500 Ma, followed immediately by post-orogenic granite magmatism at about 500 Ma (Almeida and Mantovani, 1975). A SE-NW trending branch of the Paraguay Belt extends from Corumbá into Bolivia, being known as the Tucavaca Belt (Litherland *et al.*, 1986; O'Connor and Walde, 1986). It appears as a large synclinal structure in Bolivia (Figs. 1 and 2), and was interpreted as an aulacogen in contrast with the previous interpretation of a Brasiliano/Pan-African fold belt isolating the Amazonian Craton from the cratonic Rio Apa Block (Trompette, 1994).

Lithostratigraphy

Three major lithostratigraphic units can be distinguished (Figs. 2 and 3). The lowermost unit consists of glaciomarine beds and sediments of turbidite character with glacial affinities. This is overlain by a carbonate unit that marks the end of the glacial influence in the basin, and can be considered as a chronostratigraphic unit related to a period of relative sea-level rise (Fig. 3). The uppermost unit consists of a sequence of siliciclastic rocks.

The Lower Unit

The lower unit includes the Puga Formation, Jacadigo and Boqui groups on the craton, and the Cuiabá Group in the fold belt (Figs. 2 and 3). The Puga Formation consists of diamictite beds associated with conglomerate, sandstone, siltstone and shale. The widespread diamictite association shows a lateral transition from thin, coarse-grained beds near the Amazonian Craton (Puga Formation) to thick, finer-grained facies in the E (Cuiabá Group). Fine-grained detrital sediments associated with diamictite, conglomerate and sandstone units comprise the Cuiabá Group that differs in composition from N to S.

In the northern part of the belt (Cáceres, Cuiabá and Província Serrana regions; Figure 2), according to Alvarenga and Trompette (1992), the sedimentation model proposed for the lowest unit at the edge of the Amazonian Craton involves three main depositional systems. Glaciomarine sediments cover the Amazonian Craton on the western inner shelf, which were partially reworked by gravity-flows on the eastern, outer shelf. The second facies association, developed along the continental slope, is characterized by the supply of platform glaciomarine material, reworked by powerful gravity flows. This mechanism resulted in the down-slope accumulation of sediments (conglomerate, sandstone and diamictite), being interpreted as a submarine channel system. The third facies association, in the central part of the basin, represents the outer slope depositional system, dominated by fine-grained deposits (phyllite and metasiltstone) forming turbidite cycles assigned to the Cuiabá Group.

In the S of the fold belt, NE of Bonito (Fig. 2), the Cuiabá Group consists of folded rocks (limestone, laminated siltstone, and sandstone) with cleavage. There are no associated diamictite units and other glaciogenic sediments. In this region the Cuiabá Group may include part of the carbonate unit, situated in the inner domain of the fold belt. Towards the Amazonian-Rio Apa Craton, in the Serra da Bodoquena, Corumbá and also in Bolivia, the lower unit consists of diamictite (Puga Formation and the Boqui Group), arkose and banded iron formation units (Jacadigo and Boqui groups) (Figs. 2 and 3).

In the eastern part of the fold belt (Bom Jardim, Barra do Garças, Nova Xavantina) the Cuiabá Group consists of fine-grained metasedimentary rocks (phyllite) and associated quartzite (Seer, 1985) (Fig. 2). Near Bom Jardim de Goiás the Cuiabá Group is separated by a fault from an older metavolcano-metasedimentary sequence associated with magmatic arc terranes (Seer, 1985; Pimentel and Fuck, 1992). Mafic volcanic rocks, chemical sediments (banded

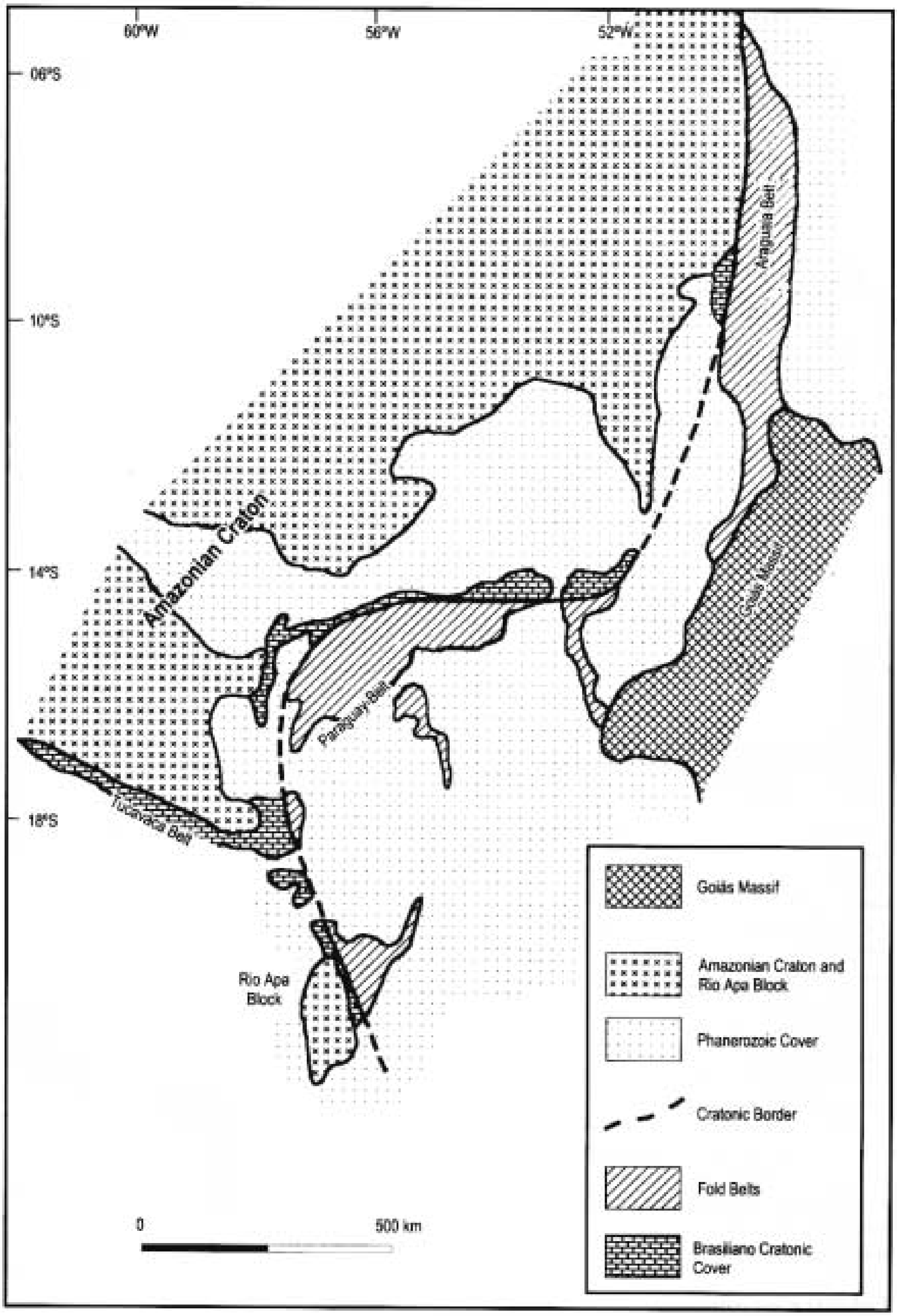


FIGURE 1 - Structural relationship between the Paraguay Belt and cratonic cover, including the Tucavaca aulacogen structure separating the Amazonian Craton and the Rio Apa Block (modified after Almeida, 1984; Litherland et al., 1986; Alvarenga and Trompette, 1993).

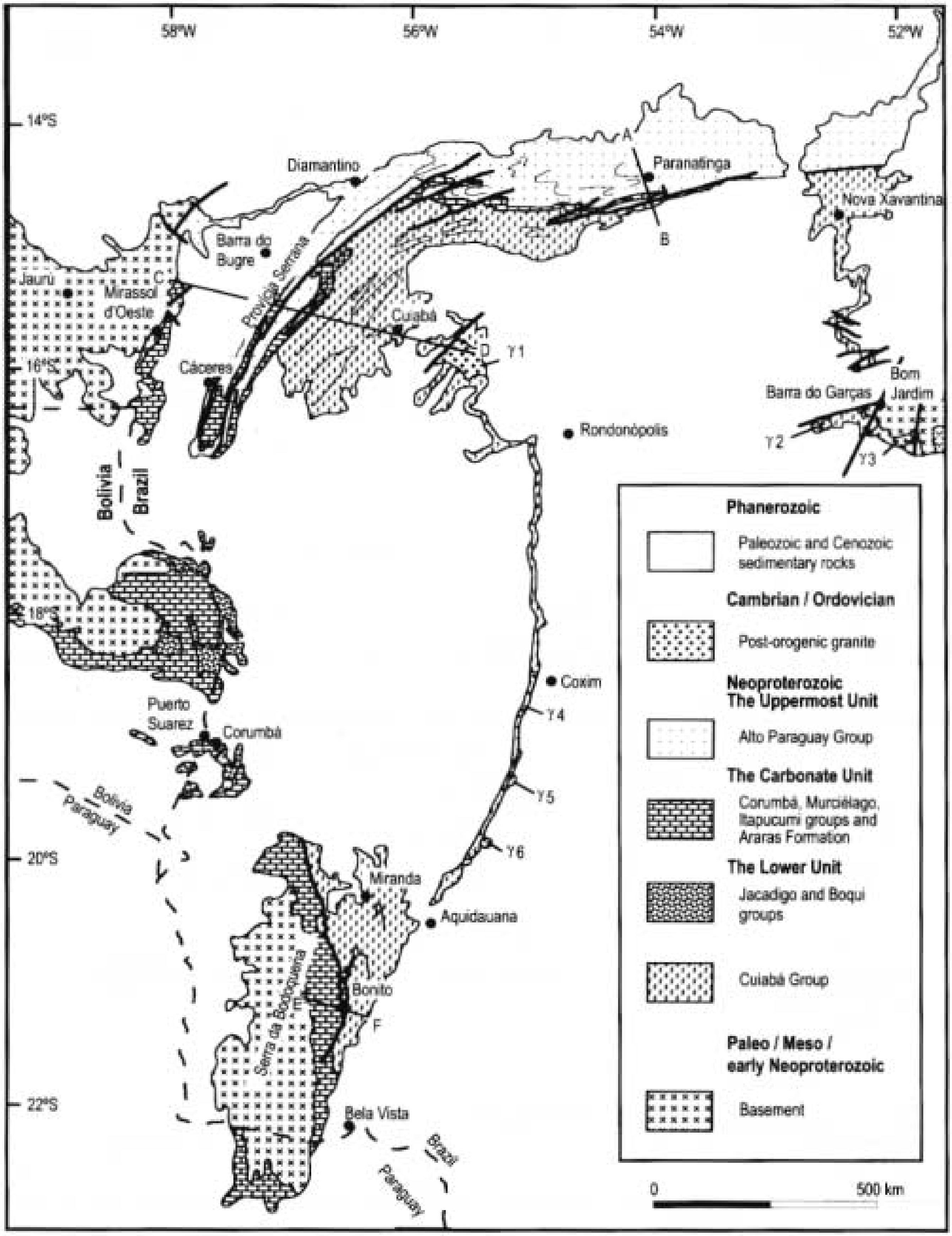
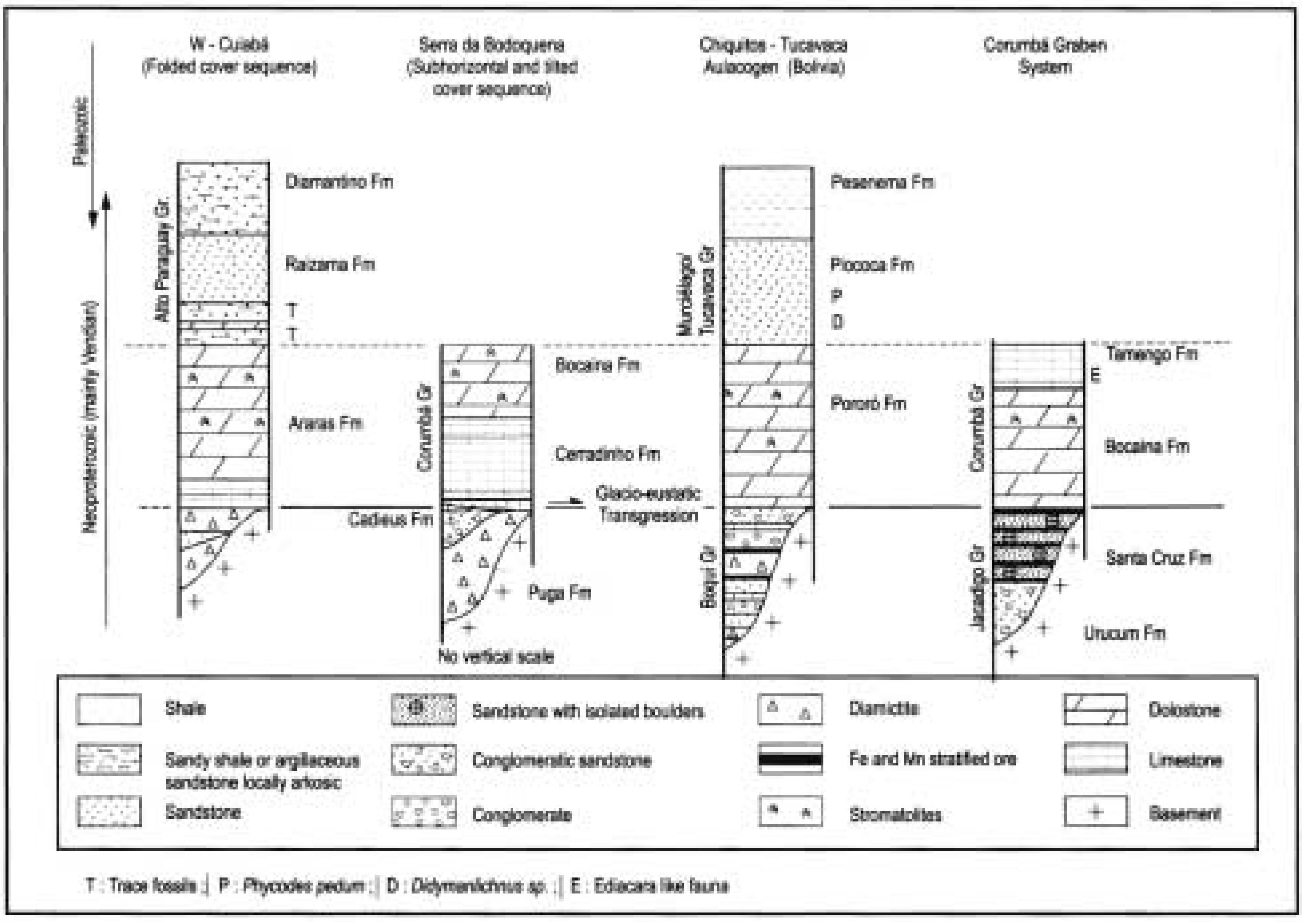


FIGURE 2 - Geological Map of Paraguay Belt (modified after Alvarenga and Trompette, 1993; Correa et al., 1979; Luz et al., 1980). Post-orogenic granites: γ1: São Vicente, γ2: Serra Verde, γ3: Piranhas, γ4: Coxim, γ5: Rio Negro, γ6: Taboco granite intrusives. AB, CD and EF are location of Fig. 4 sections.



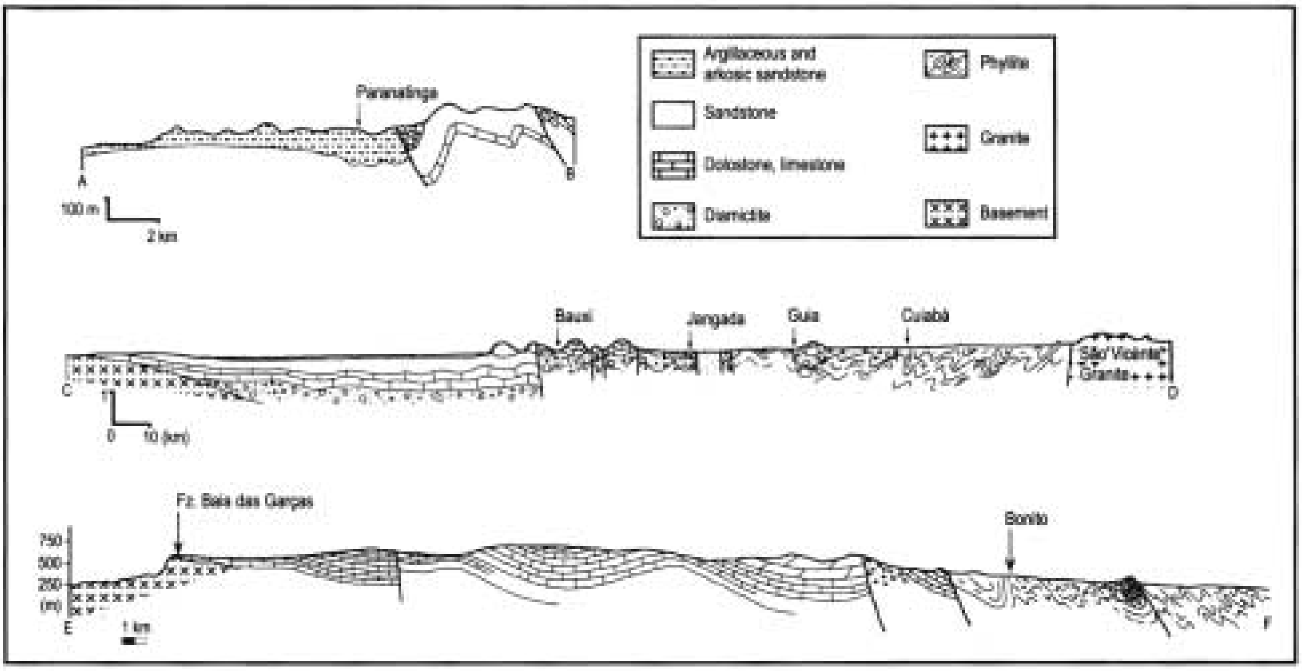


FIGURE 3 - Lithoestratigraphic correlation of Neoproterozoic to Early Cambrian sequences along the southeastern margin of the Amazonian Craton and Rio Apa Block. Transgression of the limestone and dolostone of the Araras Formation and Corumbá and Murcielago groups is of glacio-eustatic origin and supposedly synchronous (after Trompette et al., 1998).

FIGURE 4 - Sections from the cratonic platform to the inner zone of the Paraguay Belt. AB, CD and EF sections are located on the map of Figure 2.

iron formation units, chert) and carbonaceous phyllite beds limited by a shear zone are assigned to the Cuiabá Group around Nova Xavantina (Pinho, 1990; Martinelli, 1998), but this poorly known area has only a few outcrops.

The Carbonate Unit

Overlying the glaciogenic unit there is a carbonate sequence that marks the end of the glacial influence over the basin (Figs. 2 and 3). The upper limestone members of this sequence in Corumbá contain an Ediacaran-like fauna consisting of metazoans *(Cloudina lucianoi and Corumbella werneri)*, metaphytes *(Tyrasotaenia)* and microfossils *(Sphaerocongregus variabilis* sp.) (Fairchild, 1978, 1984; Hahn *et al.*, 1982; Walde *et al.*, 1982; Zaine and Fairchild, 1985, 1987; Zaine, 1991). This fossil content indicates a late Vendian age for the deposition of the carbonate rocks.

In the northern region around Cáceres, Cuiabá and Paranatinga, the Araras Formation consists of approximately 1200 m of limestone and dolostone (Luz *et al.*, 1980). These beds are finely stratified in the lower 200 m and are followed by dolograinstone and dolosparite/dolosparudite. To the E, the carbonate units pass laterally from a thick carbonate sequence on the border of the craton to a mud-rich limestone and laminated metasiltstone in the Paraguay Belt. These limestone and laminated metasiltstone beds included in the Cuiabá Group are described as the Guia Facies near the city of Cuiabá.

In the S the carbonate unit is known as the Corumbá (Brazil), Murcielago (Bolivia) and Itapucumi (Paraguay) groups (Figs. 2 and 3), which were deposited within and along the margin of the cratonic basement, and Cuiabá Group in the fold belt. Limestone and dolostone (grainstone) deposited in shallow water in the W grade to lime mudstone and mudstone towards the inner part of the basin in the E (Boggiani *et al.*, 1993) (Fig. 2). According to Boggiani (1997) the Corumbá Group reaches a thickness of 700m and consists (from bottom to top) of conglomerate, sandstone and pelitic sediments (Cadieus and Cerradinho formations), dolostone (Bocaina Formation), limestone and carbonaceous shale (Tamengo Formation).

The Uppermost Unit

The uppermost unit is known as the Alto Paraguay Group and consists of two siliciclastic formations; cross-bedded sandstone (fine to very coarse-grained sub-arkose) at the base (Raizama Formation) and red shale, siltstone and arkose at the top (Diamantino Formation). The contact between these two formations is transitional, and consists of sandstone and laminated argillite and siltstone intercalations. The unit is absent in the S, but is well exposed in the N and NE (Fig. 2). However, the thickness of these units is hard to estimate. Mapping suggests that the entire sequence is about 2800 m thick (Barros *et al.*, 1982). Shale of the lower part of the Diamantino Formation gave a Rb/Sr isochron age of 568 ± 20 Ma, interpreted as the age of diagenesis (Bonhomme *et al.* 1982; Cordani *et al.*, 1985).

Structural Geology

According to Trompette (1994), the Paraguay Belt represents a young feature of the Brasiliano/Pan-African Orogeny, aged between 550 and 500 Ma; being locally intruded by granite of about 500 Ma. Deformation and metamorphism increased from the craton towards the fold belt during the main phase (D_1), and isoclinal folds to the E and open folds to the W, with associated cleveage (Fig. 4).

Three other closely related phases of deformation (D_2 to D_4) are distinguished in the Paraguay Belt (Alvarenga 1990; Alvarenga and Trompette, 1993). Trends of D_1, D_2 and D_3 are almost identical (NE-SW), whereas D_4 is transverse (NW-SE). The first phase D_1 is the most prominent, and is contemporaneous with the regional metamorphism, whereas D_2 and D_3 phases are locally represented by crenulation cleavage (S_2 and S_3). Regional slight transverse folds characterize the D_4 phase.

The metamorphic evolution is determined by the illite crystallinity index determined by argillaceous minerals in S_1 cleavage, synchronous with the main phase (D_1). The regional vergence of the fold belt is towards the craton, as observed in rocks of the Cuiabá Group to the south (Fig. 4, section EF), and folds in the Alto Paraguay Group (Fig. 4, section AB). An opposite vergence in the Cuiabá Group near the City of Cuiabá is a matter of debate (Fig. 4, section CD).

Geological evolution

Stratigraphic, palaeontological and geocronological evidence suggests a depositional age of 600 - 540 Ma for the Paraguay and Tucavaca belts and for the cover rocks of the Amazonian Craton-Rio Apa Block. Deformation and metamorphism in the Paraguay Belt occurred between 550 and 500 Ma, and is attributed to the final stages of the Brasiliano/Pan-African Orogeny. Granite intrusions occurred locally at about 500 Ma, whereas folding in the Tucavaca Aulacogen occurred between 500 and 480 Ma (Trompette *et al.*, 1998).

The geodynamic setting of the Paraguay Belt is under discussion. Trompette (1994) and Trompette *et al.* (1998) suggested a foreland basin developed in front of the Brasilia Fold Belt. Pimentel *et al.* (1996) drew attention to the post-orogenic, extensional granitic magmatism (*c.* 580 Ma) following the Brasiliano/Pan-African deformation in the Brasilia Fold Belt at about 600 Ma. Therefore, they are in part coeval with the formation of the Paraguay and Tucavaca basins, demonstrating the regional importance of the rift episode that occurred during Neoproterozoic times.

The Araguaia Belt

The Araguaia Belt corresponds to the northern part of the Paraguay-Araguaia Belt (Almeida *et al.*, 1981). It has a general N-S orientation, and is, approximately, 1200 km long and more than 100 km wide. To the E, the Araguaia Belt is covered by the Phanerozoic rocks of the Parnaíba Basin, whereas on the western side, the anchimetamorphic to unmetamorphosed rocks of the belt rest unconformably, or are overthrust on the rocks of the Amazonian Craton (Fig. 5). In the SE, the Araguaia Belt is limited by the Goiás Massif (Almeida *et al.*, 1976). Basement inliers have been recognized in the core of dome-like structures along the eastern side of the northern part of the Araguaia Belt (Hasui *et al.*, 1984a; Herz *et al.*, 1989). The Araguaia Belt is composed of metamorphosed psamitic and pelitic sediments, with minor contributions of carbonate rocks, mafic and ultramafic rocks, and granitic bodies. The evolution of the Araguaia Belt is related to the Brasiliano thermo-tectonic event.

Lithostratigraphy

Basement Units

The basement inliers recognized in the Araguaia Belt have been grouped in two different lithostratigraphic units named the Colméia Complex (Costa, 1980) and the Cantão Gneiss (Souza *et al.*, 1985). In the Colméia Complex, Dall'Agnol *et al.* (1988) grouped orthogneiss with TTG affinity and minor amphibolite units (Fig. 5). Single zircon Pb-evaporation ages (Pb/Pb zircon ages) of 2.855 ± 0.012 Ga, 2.858 ± 0.02 Ga and 2.867 ± 0.012 Ga were obtained, respectively, for the orthogneiss of the Colméia, Grota Rica and Lontra structures (Moura and Gaudette, 1999). The Cantão Gneiss is a granitic rock with well-developed augen structures presenting a Pb/Pb zircon age of 1.858 ± 0.068 Ga (Moura and Gaudette, 1999). Archean TTGs (2.9 - 2.87 Ga) intruded by Paleoproterozoic (1.88 Ga) granitic bodies have been well described in the adjacent southeastern part of the Amazonian Craton (Macambira and Lafon, 1995). Thus, it has been suggested that the basement rocks of the northern part of the Araguaia Belt could represent inliers of the Amazonian Craton in this belt (Moura and Gaudette, 1999).

In the southern part of the Araguaia Belt, a small piece of the Goiás Massif, situated to the E of the Transbrasiliano Lineament (Fig. 5), has been considered as part of the basement of this belt (Hasui *et al.*, 1984b). A metavolcano-sedimentary sequence, named the Rio do Coco Group, has been interpreted as the remnant of an Archean greenstone belt (Barreira and Dardenne, 1981), although its age has not been determined yet. Tonalitic and calc-silicate gneiss, with Pb/Pb zircon ages between 2.1 - 2.0 Ga, were grouped in the Rio dos Mangues Complex (Moura and Souza, 1996). The Serrote Granite (Fig. 5), intrusive in the Rio dos Mangues Complex, has a Pb/Pb zircon age of 1.851 ± 0.041 Ga (Souza and Moura, 1995). Finally, the Matanga Granite, previously considered as part of the basement units, has a Rb/Sr whole-rock age of 510 ± 15 Ma (Barradas *et al.*, 1992), and may, in fact, represent granitic magmatism related to the Brasiliano thermo-tectonic event along the shear zones of the Transbrasiliano Lineament (Fig. 5). The dominance of Paleoproterozoic rocks in this region is stressed by the Pb/Pb zircon ages of 2.14 - 2.13 Ga of the granulitic rocks of the Porto Nacional Complex (Barros, 1998) (Fig. 5). These data suggest that this crustal segment, which has been named the Tocantins Shear Belt (Gorayeb, 1996), is quite distinct from the basement inliers recognized in the northern part of the Araguaia Belt.

Units of the Araguaia Belt

The metasedimentary rocks of the Araguaia Belt have been assigned to the Baixo Araguaia Supergroup. In addition to this supracrustal sequence there are alkaline felsic plutons, mafic and ultramafic bodies, and granitic rocks (Hasui *et al.*, 1984a, b; Dall'Agnol *et al.*, 1988; Herz *et al.*, 1989). The alkaline rocks of Monte Santo and Serra da Estrela suites occur in the southern part of the belt and in its adjacent basement (Fig. 5). They may represent alkaline magmatism associated with crustal rifting and formation of the Araguaia Basin, which received the sediments that ultimately formed the Baixo Araguaia Supergroup. Moura and Souza (1996) presented a Pb/Pb zircon age of 1.006 ± 0.086 Ga for a syenitic gneiss associated with the Serra das Estrelas Suite, and that may be interpreted as the age of this magmatic event.

The Baixo Araguaia Supergroup has been divided in the Estrondo and Tocantins groups. The first unit occurs in the eastern side of the Araguaia Belt (Fig. 5). It is composed of metaconglomerate, quartzite, micaschist, and schist with garnet, staurolite and kyanite (Morro do Campo Formation); feldspathic schist, with variable amounts of biotite and garnet (Canto da Vazante Formation); and micaschist, calc-schist, marble and amphibolite lenses (Xambioá Formation). The Tocantins Group lies on the western side of the Araguaia Belt (Fig. 5), and includes the Pequizeiro and Couto Magalhães formations. The former is composed mainly of chlorite-muscovite-quartz schist and minor intercalation of phyllite and quartzite. The Couto Magalhães Formation is dominated by phyllite and slate and minor amounts of quartzite, meta-arkose and metalimestone (Hasui *et al.*, 1984a; Dall'Agnol *et al.*, 1988).

Mafic and ultramafic rocks are associated with both the basement and supracrustal rocks, although the most impressive ultramafic bodies occur in the western part of the belt (Fig. 5). The tectonic emplacement of these ultramafic bodies has been demonstrated by Gorayeb (1989), who described these massifs as being composed of serpentinized peridotite and dunite and their metamorphic products (talc schist, tremolite-actinolite schist and chloritite), in addition to chromitite, chert and jaspilite. Pillow basalt has also been reported (Kotschoubey *et al.*, 1996). This rock association has been interpreted as the remnant of an ophiolite complex, suggesting the presence of an oceanic crust in the Araguaia Belt.

Granite bodies, associated with the highest metamorphic grade rocks of the Estrondo Group, occur along the Araguaia Belt (Fig. 5). They have been considered as the product of partial melting of the supracrustal sequences during the peak of the metamorphism (Dall'Agnol *et al.*, 1988; Abreu *et al.*, 1994). A Pb/Pb zircon age of 655 ± 24 Ma (Moura and Gaudette, 1993) for the Santa Luzia Granite (Hasui *et al.*, 1984b; Lamargo and Kotschoubey, 1996) indicates the age of emplacement of this granitic body.

Metamorphism

The regional metamorphism affecting the rocks of the Araguaia Belt increases gradually from incipient in the W to middle-high amphibolite facies in the E. The metamorphism is of the Barrovian type, and different metamorphic zones, oriented N-S, may be mapped along the belt. From W to E, the pelitic sequences show the following mineral assemblages: sericite-chlorite, muscovite-chlorite-epidote,

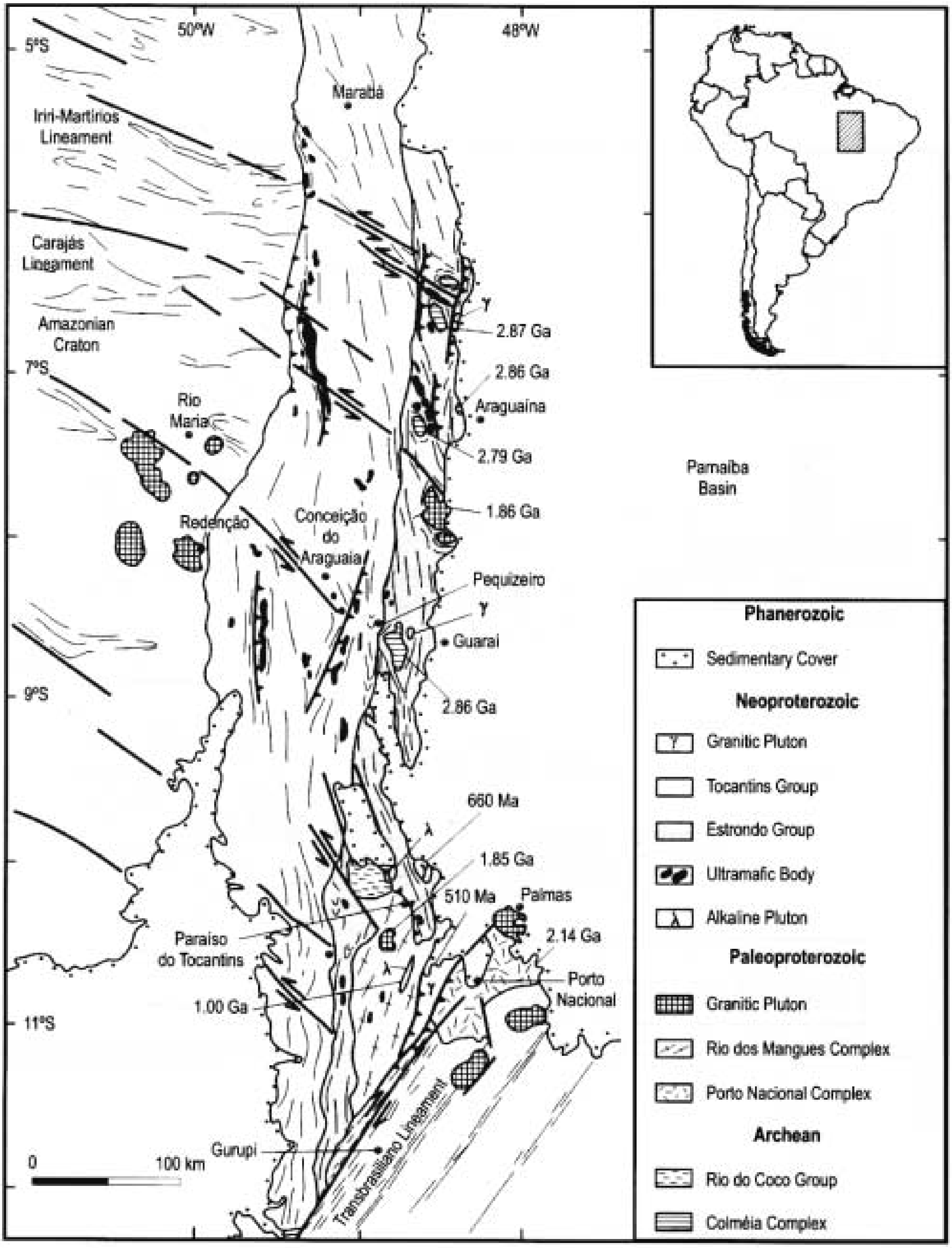


FIGURE 5 - Simplified geological map of the Araguaia Belt and its basement units. Geochronological data are referred to in the text.

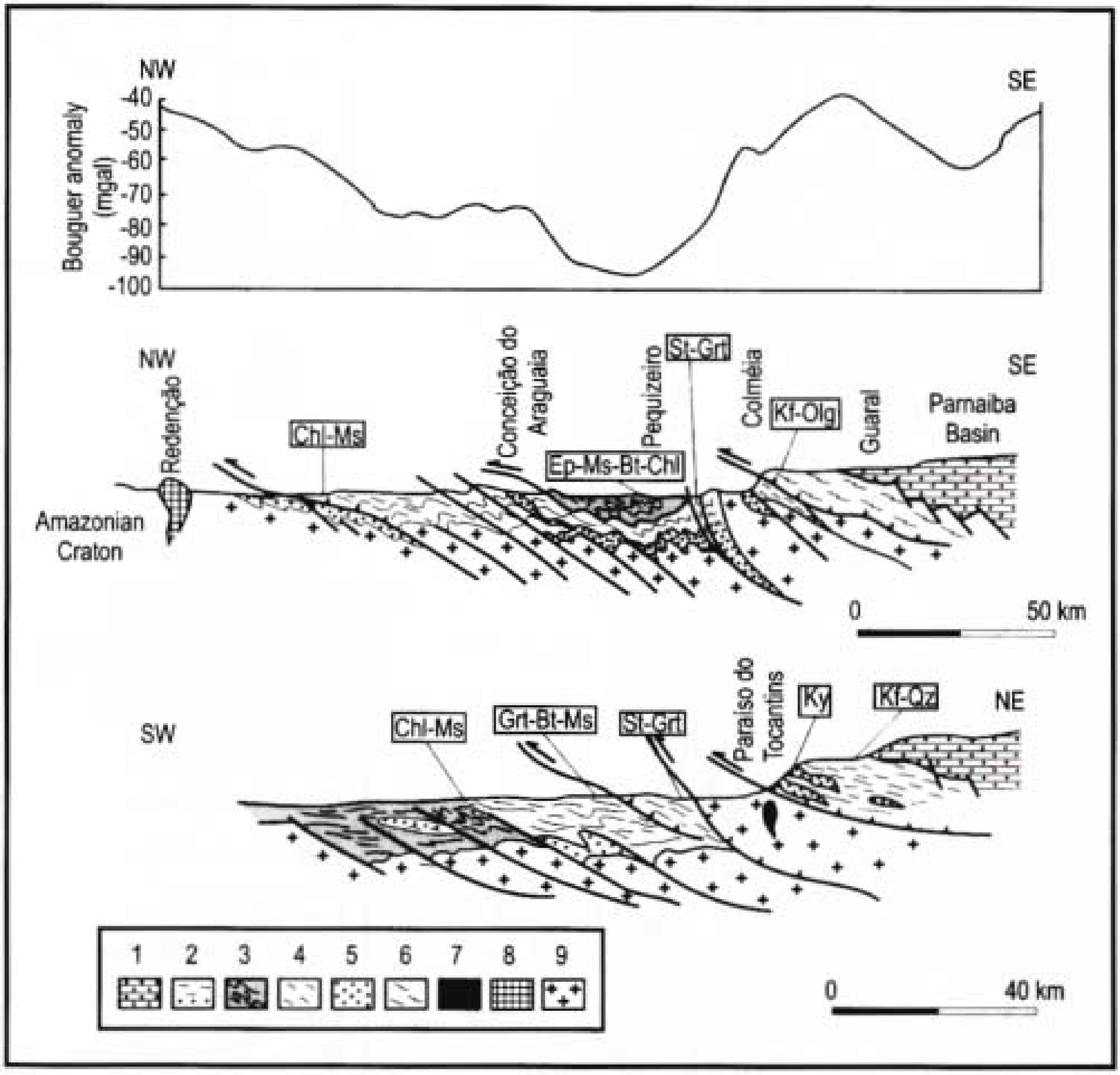


FIGURE 6 - Simplified geological sections across the Araguaia Belt, showing the relationships between supracrustal sequences and basement rocks, the metamorphic mineral assemblages and the gravimetric profile (modified after Abreu et al., 1994). 1- Phanerozoic cover; 2- Couto Magalhães Formation; 3- Pequizeiro Formation; 4- Xambioá Formation; 5- Morro do Campo Formation; 6-Canto da Vazante Formation; 7- Serra da Estrela Suite; 8- Redenção Granite; 9- Archean and Paleoproterozoic basement.

muscovite-biotite±chlorite, muscovite-biotite-garnet, biotite-muscovite-garnet-staurolite, biotite-garnet-kyanite and, finally, restricted areas of partial melting generating quartz-feldspar pockets and small granite bodies (Fig. 6). During the peak, the metamorphism has reached temperatures of approximately 650 °C, and pressures over 6 kbar (Abreu *et al.*, 1994). As the foliation of the metamorphic rocks, in general, dips eastwards and the metamorphic grade increases from W to E, Abreu *et al.* (1994) suggested an inversion of the metamorphic terrains, since the rocks of higher metamorphic grade overlie lower grade rocks (Figs. 6).

The age of emplacement of the syn-tectonic Santa Luzia Granite, around 660 Ma, probably dates the peak of this metamorphism and defines the age of the metamorphic event. Previous Rb/Sr ages obtained in the metasedimentary rocks of the Baixo-Araguaia Supergroup (Hasui *et al.*, 1980), ranging, roughly, between 1.0 Ga and 500 Ma, are not true isochronic ages, and probably reflect the non-equilibrium of the Rb/Sr system in these metamorphic rocks. K/Ar ages between 560 and 520 Ma obtained in biotite, muscovite and hornblende from schist and amphibolite of the Estrondo Group (Macambira, 1983) may record cooling ages of this metamorphism. The imprinting of this metamorphic event on the basement rocks of the Araguaia Belt is indicated by K/Ar ages around 530 Ma in biotite and muscovite of the Archean basement gneiss (Macambira, 1983); and by Rb/Sr mineral ages in the Paleoproterozoic granitoid plutons: 500 - 470 Ma for the Cantão Gneiss (Lafon *et al.*, 1990) and 536 ± 37 Ma for the Serrote Granite (Souza and Moura, 1995). All these data relate the metamorphism of the Araguaia Belt with the Brasiliano thermo-tectonic event.

Structural Geology

The approximate N-S trend of the Araguaia Belt is shown not only by the disposition of the lithological units, but also by the pervasive NNW-SSE and NNE-SSW trend of the foliation, impressed on both supracrustal and basement rocks. Variations in this foliation trend may be observed: 1) near the dome-like structures where the foliation follows the configuration of the structure, 2) near the NW-SE ductile and brittle-ductile shear zones that cross the Araguaia Belt and, 3) in the southern part of the belt where it is affected by the NNE-SSW transcurrent shear zones of the Transbrasiliano Lineament. Major N-S thrust faults indicate westward crustal shortening (Fig. 5). The dome-like structures, previously interpreted as related with granitic diapirism (Hasui *et al.*, 1984a), are now related with thrust faulting involving basement and the supracrustal sequence (Abreu *et al.*, 1994). Mineral and stretching lineations dip gently to the SE except around the dome-like structures, showing a 50 - 60 ° rake on the low-middle angle E-dipping foliation. Structural analysis of the planar and linear elements suggests tectonic transport towards the NW indicating an oblique collision with the Amazonian Craton (Costa *et al.* 1988; Abreu *et al.* 1994).

Evolution

The Araguaia Belt developed in the articulation zone of the Amazonian Craton with the Paleoproterozoic Tocantins Shear Belt. The geometry and the evolution of the basin were constrained by ancient NW-SE discontinuities, mainly on the Amazonian Craton, and NNE-SSW discontinuities on the limit of these geotectonic units. The original cross-section of the basin shows low-angle dip towards the E, where the deepest part of the basin developed. In its evolution, during the extensional rift phases, NW-SE discontinuities acted as transfer faults, developing different basin compartments with different depocentres. The sector of the basin to the S of the Carajás Lineament, where the majority of the ultramafic bodies and pillow basalts occur, was the most extended compartment, and evolved to a proto-oceanic stage of the continental rift evolution. Alkaline felsic magmatism (1.0 Ga Estrela Suite) took place within the basement at the eastern margin of the basin.

The basin was filled with a continental margin sedimentary sequence, clastic and minor carbonate, followed by fine-grained sediments characteristic of deeper marine environments. Meanwhile, there was the emplacement of mafic magma as flows, sills and dykes. The basin inversion was accompanied by the reactivation of the ancient basement structures, regional metamorphism and crustal anatexis (660 Ma Santa Luzia Granite). Thrusting, the formation of nappes, and tectonic imbrication of the lithostratigraphic-metamorphic units led to the inversion of the supracrustal pile.

The compressional stage involved tectonic transport toward NW indicating an oblique collision against the Amazonian Craton. The previous NW-SE transfer faults acted, at that moment, as sinistral transcurrent fault zones. The basement rocks were displaced in the process, overthrusting the supracrustal rocks, forming the basement inliers (Fig. 6). At the end of this evolution, strike-slip movement and granite emplacement (510 Ma Matança Granite) occurred along the Transbrasiliano Lineament.

Final Remarks

The Paraguay and Araguaia belts have more differences than similarities, and have been separated in two different tectonic units (Almeida, 1986; Trompette 1994). These two fold belts are placed on the border of the same cratonic block (Amazon Craton), and their evolution is related to the Brasiliano/Pan-African Orogeny.

The fill of the Paraguay Belt basin started with a glacially-influenced sedimentation consisting mainly of diamictite beds and coarse to fine-grained turbidite, and carbonate units with an Ediacaran-like fauna. Gradients of deformation and metamorphism are moderate, and sedimentary structures are preserved. The magmatic contribution is reduced to some post-orogenic granite intrusives in the inner part of the belt.

The Araguaia Fold Belt has a clastic filling in which the regional metamorphism increases gradually from incipient in the W to middle-high amphibolite facies in the E. Mafic and ultramafic rocks are associated with basement and supracrustal rocks. They have been interpreted as remnants of an ophiolite complex, suggesting the presence of an oceanic crust in the Araguaia Belt.

REFERENCES

Abreu, F.A.M., Gorayeb, P.S.S. and Hasui, Y. (1994). Tectônica de inversão metamórfica no Cinturão Araguaia. *Actas, IV Simpósio de Geologia da Amazônia*, Belém, *Brasil*, 1-4.

Almeida, F.F.M. (1967). Origem e evolução da plataforma brasileira. (DNPM), Div. Geol. Mineral., Rio de Janeiro, Brazil. *Boletim Departamento Nacional da Produção Mineral*, 241, 36 p.

Almeida, F.F.M. (1968). Evolução tectônica do centro-oeste brasileiro no Proterozóico Superior. Rio de Janeiro, Brasil. *Anais da Academia Brasileira de Ciências*, **40** (suplemento), 285-293.

Almeida, F.F.M. (1974). Sistema tectônico marginal do cráton do Guaporé. *Anais, XXVIII Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Porto Alegre*, **4**, 11-17.

Almeida, F.F.M. (1984). Província Tocantins, setor sudoeste. In: *O Pré-Cambriano do Brasil*, eds. F.F.M. de Almeida and Y. Hasui, pp. 265-281. Edgard Blücher, São Paulo.

Almeida, F.F.M. (1986). Alguns problemas das relações geológicas entre o Cráton Amazônico e as faixas de dobramentos marginais a leste. *Ata, II Simpósio Geologia do Centro-Oeste, Sociedade Brasileira de Geologia, Goiânia, 1986*, 3-14.

Almeida, F.F.M. and Mantovani, M.S.M. (1975). Geologia e geocronologia do granito de São Vicente, Mato Grosso. Rio de Janeiro, Brazil, *Anais da Academia Brasileira de Ciências*, **47**, 451-458.

Almeida, F.F.M., Hasui, Y. and Brito-Neves, B.B. (1976). The Upper Precambrian of South America. *Boletim do Instituto de Geociências, Universidade de São Paulo*, 7, 45-80.

Almeida, F.F.M., Hasui, Y., Brito-Neves, B.B. and Fuck, R.A. (1981). Brazilian structural provinces: an introduction. *Earth Science Review*, **17**, 1-29.

Alvarenga, C.J.S. (1990). *Phénomènes sédimentaires, structuraux et circulation de fluides developpes a la transition chaine-craton. Exemple de la chaine Paraguai d'age protérozoïque supérieur, Mato Grosso, Brésil*. Doctor thesis. Université d'Aix-Marseille III, France, 177 p.

Alvarenga, C.J.S. and Trompette, R. (1992). Glacial influenced turbidite sedimentation in the uppermost Proterozoic and Lower Cambrian of the Paraguay Belt (Mato Grosso, Brazil). *Palaeogeography, Palaeoclimatology, Palaeoecology*, **92**, 85-105.

Alvarenga, C.J.S. and Trompette, R. (1993). Evolução tectônica brasiliana da Faixa Paraguai na região de Cuiabá. *Revista Brasileira de Geociências*, **23**(1), 18-30.

Barradas, J.A., Lafon, J-M. and Kotschoubey, B. (1992). Geocronologia Pb-Pb e Rb-Sr da região de Monte do Carmo-Porto Nacional, TO. Novos resultados. *Anais, XXXVII Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, São Paulo*, **2**, 182-183.

Barros, G. (1998). Estudo geocronológico dos zircões dos granulitos de Porto Nacional (TO). *Actas, IX Seminário de Iniciação Científica da Universidade Federal do Pará*, Belém, 283.

Barreira C.P. and Dardenne M.A. (1981). A seqüência vulcano-sedimentar Rio do Coco. *Ata, I Simpósio de Geologia do Centro-Oeste*, Sociedade Brasileira de Geologia, Goiânia, 1, 241-264.

Barros, A.M., Silva, R.M., Cardoso, O.R.F.A., Freire, F.A., Souza Jr., J.J., Rivetti, M., Luz, D.S., Palmeira, R.C.B. and Tassinari, C.C.G. (1982). *Folha SD21 Cuiabá, Geologia. RADAMBRASIL, Levant. Rec. Naturais, Rio de Janeiro*, **26**, 25-192.

Boggiani, P.C. (1997). *Análise estratigráfica da Bacia Corumbá (Neoproterozóico) - Mato Grosso do Sul*. Doctor thesis, Instituto de Geociências, Universidade de São Paulo, São Paulo, Brasil, 181 p.

Boggiani, P.C., Fairchild, T.R. and Coimbra, A.M. (1993). O Grupo Corumbá (Neoproterozóico-Cambriano) na região central da Serra da Bodoquena (Faixa Paraguai), Mato Grosso do Sul. *Revista Brasileira de Geociências*, **23**, 301-305.

Bonhomme, M.G., Cordani, U.G., Kawashita, K., Macedo, M.H.F. and Thomaz Filho, A., (1982). Radiochronological age and correlation of Proterozoic sediments in Brazil. *Precambrian Research*, **18**, 103-118.

Cordani, U.G., Thomaz Filho, A., Brito-Neves, B.B. and Kawashita, K. (1985). On the applicability of the Rb-Sr method to argillaceous sedimentary rocks: some examples from Precambrian sequences of Brazil. *Journal of Geology*, **471**, 253-280.

Costa, J.B.S. (1980). Estratigrafia da região de Colméia. *Anais, XXXI Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Camboriú*, **2**, 720-728.

Costa, J.B.S., Hasui, Y., and Gorayeb, P.S.S. (1988). Relações estruturais entre a Faixa Araguaia e o Maciço goiano na região de Paraíso do Norte - Gurupi, GO. *Anais, XXXV Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Belém*, **5**, 2187-2196.

Correa, J.A., Correia Filho, F.C.L., Scisiewski, G., Neto, C., Cavallon, L.A., Cerqueira, N.L.S. and Nogueira, V.L. (1979). *Geologia das regiões Centro e oeste de Mato Grosso. Projeto Bodoquena. (Mapa Geol. Escala 1:250 000)*. Brasília, DNPM, Série Geológica Básica 3, 111 p.

Dall'Agnol, R, Teixeira, N.P., Macambira, J.B., Kotschoubey, B., Gorayeb, P.S.S., and Santos, M.D. (1988). Petrologia dos gnaisses e micaxistos da porção norte da faixa de dobramentos Araguaia, Goiás-Brasil. *Actas, VII Congresso Latino-Americano de Geologia, Belém*, 1, 1-1 9.

Fairchild, T. R. (1978). Evidências paleontológicas de uma possível idade ediacariana ou Cambriano inferior, para parte do Grupo Corumbá, Mato Grosso do Sul. *Boletim Especial, XXX Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Recife*, 181 p.

Fairchild, T. R. (1984). Caution: an ediecaran or early Cambrian age for the the Corumbá and Jacadigo Group (SW-Brazil) still requires definitive proof. Abstract, *XXVII International Geological Congress, Moscow*, 1, 38-39.

Gorayeb, P.S.S. (1989). Corpos serpentiníticos da Faixa Araguaia na Região de Araguacema-Pequizeiro-Conceição do Araguaia (Goiás-Pará). *Revista Brasileira de Geociências*, **19**, 51-62.

Gorayeb, P.S.S. (1996) *Petrologia e evolução crustal das rochas de alto grau de Porto Nacional (TO)*. Doctor thesis. Universidade Federal do Pará, Belém (Brazil), 262 p.

Hahn, G., Hahn, R., Leonardos, O.H., Pflug, H. D. and Waide, D.H.G. (1982). Körperlich erhaltene Scyphozoen - Reste aus dem Jungpräkambrium Brasiliens. *Goologica et Palaeontologica*, **16**, 1-18.

Hasui, Y., Tassinari, C.C.G., Siga Jr., O., Teixeira, W., Almeida, F.F.M. and Kawashita, K. (1980). Datações Rb-Sr e K-Ar do centro norte do Brasil e seu significado geológicogeotectônico. *Anais, XXXI Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Camboriu*, **5**, 2659-2667.

Hasui, Y., Costa, J.B.S. and Abreu, F.A.M. (1984a). Província Tocantins. Setor setentrional. In: *O Precambriano no Brasil*, eds. de Almeida, F.F.A. and Hasui, Y., pp. 137-204. Edgard Blücher, São Paulo.

Hasui, Y., Costa, J.B.S., Gorayeb, P.S.S., Lemos, R.L.L., Gama Jr., T. and Bemerguy, R.L. (1984b). Geologia da Região de Paraíso do Norte de Goiás, GO. *Anais, XXXIII Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Rio de Janeiro*, **5**, 2220-2233.

Herz, N., Hasui, Y., Costa, J.B.S. and Matta, M.A.S. (1989). The Araguaia Fold Beft, Brazil: A reactivated Brasiliano-Pan-Africano cycle (550 Ma) geosuture. *Precambrian Research*, **42**, 371-386.

Koutschoubey, B., Hieronimus, B., Rodrigues, O.B. and Amaral, R.T. (1996). Basaltos e serpenfinitos da área da Serra do Tapa (PA). Prováveis testemunhos de um complexo ofiolítico pouco evoluído e desmembrado. *Anais, XXXII Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Salvador*, 6, 25-28.

Lafon J.M., Macambira, J.B.; Macambira M.J.B., Moura, C.A.V., Gaudette, H.E. and Souza, A.C.C. (1990). A Faixa de Dobramentos Araguaia (TO): novos dados geocronológicos. *Anais, XXXIV Congresso Brasileiro de Geologia, Sociedade Brasileira de Geologia, Natal*, 5, 2550-2563.

Lamargo, C.N. and Kotschoubey, B. (1996). Granitóides Santa Luzia registro do magmatismo granítico brasiliano do cinturão Araguaia na região de Paraíso do Tocantins. *Revista Brasileira de Geociências*, 26, 277-288.

Litherland, M., Annells, R.N., Appleton, J.D., Berrangé, J.P., Bloonfield, K., Burton, C.C.J., Darbyshire, D.P.F., Fletcher, C.J.N., Hawkins, M.P., Klink, B.A., Llanos, A., Mitchell, W.I., O'Connor, E.A., Piffield, P.E.J., Power, G. and Weeb, B.C. (1986). The geology and mineral resources of the Bolivian Precambrian shield. *Overseas Memoir- British Geological Survey*, London, 9, 153 p.

Luz, J.S., Oliveira, A.M., Souza, J.O., Motta, J.F.M., Tanno, L.C., Carmo, L.S. and Souza, N.B. (1980). Projeto Coxipó. *Ministério das Minas Energia, Departamento Nacional da Produção Mineral, Companhia Pesquisa e Recursos Minerais, Relatório Final. Goiânia, Brasil*, 1, 1-136.

Macambira, J.B. (1983). *Geologia e ocorrências minerais do braquianticlinal do Lontra (GO)*. M.Sc. thesis. Universidade Federal do Pará, Belém, Brasil, p. 108.

Macambira, M.J.B. and Lafon, J.-M. (1995). Geocronologia da Província Mineral de Carajás: síntese dos dados e novos desafios. *Boletim do Museu Paraense Emilio Goeldi, série Ciências da Terra*, 7, 263-288.

Martinelli, C.D. (1998). *Petrografia, estrutural e fluidos da mineralização aurífera dos Araés-Nova Xavantina-MT*. Doctor thesis. Instituto de Geociências, Universidade Estadual Paulista, Rio Claro, São Paulo (Brazil), 183 p.

Moura, C.A.V. and Gaudette, H.E. (1993). Zircon ages of the basement orthogneiss of the Araguaia Belt, north-central Brazil. *Actas, IV Congresso Brasileiro de Geoquímica, Brasília, Brasil*, 232-235.

Moura, C.A.V. and Gaudette, H.E. (1999). Zircon Ages of the Basement Orthogneisses from the Northern Segment of the Araguaia Belt, Brazil. In: *Basement Tectonic 13*, ed. Sinha, A.K., pp. 155-178. Holanda: Kluwer Academic Publishers.

Moura, C.A.V. and Souza, S.H.P. (1996). Síntese dos dados geocronológicos das rochas do embasamento do Cinturão Araguaia e suas implicações estratigráficas. Anais, XXXIX *Congresso Brasileiro de Geologia*, Sociedade Brasileira de Geologia, Salvador, Brasil, 3, 31-34.

O'Connor, E.A. and Waide, D.H.G. (1986). Recognition of an Eocambrian orogenic cycle in SW Brazil and SE Bolivia. *Zentralblatt für Geologie und Paläontologie, Teil 1, Stuttgart*, 1985(H. 9/10), 1441-1456.

Pimentel, M.M. and Fuck, R.A. (1992). Neoproterozoic crustal accretion in central Brazil. *Geology*, 20, 375-379.

Pimentel, M.M., Fuck, R.A. and Alvarenga, C.J.S. de (1996). Post-Brasiliano (Pan-African) high-K granitic magmatism in central Brazil: the role of late Precambrian-early Paleozoic extension. *Precambrian Research*, 80, 217-238.

Pinho, F.E.C. (1990). *Estudo das rochas encaixantes e veios mineralizados a ouro do Grupo Cuiabá, na região denominada Garimpo do Araés, Nova Xavantina, Estado de Mato Grosso)*. M.Sc. thesis. Instituto de Geociências, Universidade Federal do Rio Grande do Sul, Porto Alegre, RS, Brasil, 114 p.

Silva, G.G., Lima, M.J.C., Andrade, A.R.F. and Issler, R.S. (1974). Folha SB-22 Araguaia e partes da Folha SC22 Tocantins. Geologia. *RADAMBRASIL, Levant. Rec. Naturais, Rio de Janeiro*, 4, 22-131.

Seer, H.J. (1985). *Geologia, deformação e mineralização de cobre no complexo vulcano sedimentar de Bom Jardim de Goiás*. M.Sc. thesis, Instituto de Geociências, Universidade de Brasília, Brasília, DF, Brasil, 181p.

Sousa, D.J.L. and Moura, C.A.V. (1995). Estudo geocronológico do Granito Serrote, Paraíso do Tocantins (TO). *Anais, V Congresso Brasileiro de Geoquímica, Niterói, Brasil*, CDROM.

Souza, A.C.C., Dall'Agnol, R., Teixeira, N.P. (1985). Petrologia do gnaisse Cantão: implicações na evolução da faixa de dobramentos Araguaia. Serra do Estrondo (GO). *Revista Brasileira de Geociências*, 15, 300-310.

Trompette, R. (1994). *Geology of Western Gondwana (2000-500 Ms). Pan-African-Brasiliano aggregation of South America and Africa*. Balkema, 350 pp.

Trompette, R., Alvarenga, C.J.S. de and Walde, D. (1998). Geological evolution of the Neoproterozoic Corumbá graben system (Brazil). Depositional context of the stratified Fe and Mn ores of the Jacadigo Group. *Joumal of South American Earth Science*, 11, 587-597.

Walde, D.H.G., Leonardos, O.H., Hahn, G. and Pflug, H.D. (1982). The first pre-Cambrian megafossils from South America, *Corumbelia wemeri. Anais da Academia Brasileira de Ciências*, 54, 461.

Zaine, M.F. (1991). *Análise dos fósseis de parts da faixa Paraguai (MS, MT) e seu contexto temporal e paleoambiental*. Doctor thesis, Instituto de Geociências, Universidade de São Paulo, São Paulo, Brazil, 215 p.

Zaine, M.F. and Fairchild, T.R. (1985). Comparaison of *Aulophycus lucianoi* Beurlen and Sommer from Ladário (MS) and the genus *Cloudina Germs*, Ediacaran of Namibia. *Anais da Academia Brasileira de Ciências*, 57, 180.

Zaine, M.F. and Fairchild, T.R. (1987). Novas considerações sobre os fósseis da formação Tamengo, grupo Corumbá, SW do Brasil. *Anais, X Congresso Brasileiro de Paleontologia*, Rio de Janeiro, Brazil, 2, 797-806.

The Basement of the Brasília Fold Belt and the Goiás Magmatic Arc

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The Tocantins Province, in the central part of Brazil, represents a large Brasiliano/Pan-African orogen formed between three major continental blocks: the Amazonian, São Francisco/Congo and Paraná cratons (Fig. 1). The province comprises three important supracrustal fold belts known as the Araguaia, Paraguai and Brasília belts. The Araguaia and Paraguai belts occupy the western part of the province, bordering, respectively, the eastern and southern margins of the Amazonian Craton. The Brasília Belt forms the eastern part of the Tocantins Province, flanking the western edge of the São Francisco Craton. This belt is about 1100 km long, extending N-S from western Minas Gerais, through Goiás to southeastern Tocantins. It constitutes, probably, the best preserved Neoproterozoic orogenic belt in Brazil, displaying features such as an ophiolitic mélange, calc-alkaline arc rocks, S-type collisional granites, and regional scale nappe structures indicating tectonic transport towards the E.

Separating the supracrustal fold belts, in the central part of the Tocantins Province, there is a large and geologically complex area where orthogneiss, granitoid, volcano-sedimentary sequences, layered complexes and granulite bodies of various ages and nature are exposed (area shown in black in Figure 1a). In the past, due to the scarcity of geochronological data, this was interpreted as the exposure of a large section of the Archean sialic basement of the supracrustal belts, representing a "median massif" (Almeida *et al.*, 1981; Marini *et al.*, 1984). The data available then did not indicate the presence of Neoproterozoic calc-alkaline rock units, which would be indicative of the consumption of oceanic lithosphere. During the last decade, however, field and geochronological studies of some areas of the central part of the Tocantins Province have revealed a rather different picture. Firstly, the new data indicate that this area comprises diverse geological/tectonic domains with different ages and tectonic significance. Furthermore, the data also point to the importance of Neoproterozoic granitic magmatism in the geological evolution of the orogen (for a review, see Pimentel *et al.*, 1999a). For example, the discovery of Neoproterozoic juvenile arc rocks, as well as of collision-related peraluminous granite intrusions associated with metasedimentary rock units in the southern part of the Brasília Belt, were fundamental to challenge previous ensialic models (Brito Neves and Cordani, 1991; Pimentel and Fuck, 1992; Pimentel *et al.*, 1997).

The different domains recognized in the central part of the Tocantins Province are: (i) the Archean terranes of Crixás-Goiás, interpreted as a small allocthonous continental block, (ii) Paleoproterozoic sialic basement of the supracrustal belts (*e.g.*, the orthogneiss and volcano-sedimentary sequences in the Almas-Dianópolis area or the Anápolis-Itauçu high-grade complex), (iii) the large Paleo-Mesoproterozoic mafic-ultramafic layered complexes of Barro Alto, Niquelândia and Cana Brava and associated oceanic volcano-sedimentary sequences, and (iv) the Neoproterozoic juvenile Goiás Magmatic Arc.

As detailed below, these rock associations have been affected, to different degrees, by the Neoproterozoic tectono-thermal events that built the Brasília Belt. They represent, therefore, important pieces of the tectonic puzzle of this orogen. Thus, understanding their geological history is fundamental to the clarification of the geodynamic evolution of the Tocantins Province.

In this chapter we present an up-to-date review of the geology and geochronology of the different segments of the central part of the Tocantins Province, and discuss their possible tectonic significance in relation to the history of the province.

The Archean terranes of Crixás (Goiás) - an allochtonous block

The Archean terranes of Goiás crop out within an area of about 50 000 km^2 in the central part of the State as a typical association of granite-gneiss complexes and greenstone belts. The Archean area is oval-shaped, NE-SW oriented, and its limits are entirely tectonic. To the N and W, it is in contact with Neoproterozoic rocks of the Goiás Magmatic Arc, to the S and E, with Proterozoic metasedimentary and metavolcanic units (Fig. 2).

Linear belts of supracrustal rocks in that area were first interpreted as greenstone belts by Danni and Ribeiro (1978) and Sabóia (1979), following the discovery of well-preserved spinifex-textured ultramafic rocks in Crixás. Geological investigation of the granite-greenstone association increased considerably during the last 20 years. Results were published in Kuyumjian (1981); Danni *et al.* (1982,1986); Jost *et al.* (1995, 1996); Lacerda and Lima Júnior (1996); Montalvão (1985); Montalvão *et al.* (1982); Jost and Oliveira (1991), and reviewed by Marini *et al.* (1984); Danni (1988); Jost *et al.* (1999).

Granite-gneiss domains comprise about 80% of the Archean segment and consist of orthogneiss and tonalite-granodiorite intrusions. Originally, Danni and Ribeiro

(1978) grouped the granitoid plutons of the northern portion of the Archean terranes into four independent blocks or complexes (Anta, Caiamar, Moquém, and Hidrolina; Figure 2). To the S, the Anta and Caiamar complexes extend towards the Caiçara Complex, previously called Itapuranga Complex, and later renamed as Caiçara Complex by Jost *et al.* (1999) to avoid confusion with the Itapuranga Suite of Oliveira (1993), which consists of alkali granite and syenite intrusion of possible Neoproterozoic age. The Caiçara Complex extends southwards to the northern limits of the Serra de Santa Rita greenstone belt. Gneiss and granite intrusions to the S of this belt are grouped into the Uvá Complex, which extends to the S up to the lower slope of the Serra Dourada range formed by Proterozoic metasedimentary rock units.

The Archean supracrustal rocks occur in five belts between 40 km and 100 km long and, in the average, 6 km wide. Three of them are in the northern part (Crixás, Guarinos and Pilar de Goiás) and the other two in the southern part (Serra de Santa Rita and Faina) of the Archean Block (Fig. 2). The former occur side by side, separated by granite-gneiss terranes and trend almost N-S, whereas the latter apparently form one single N60°W belt truncated by a dextral strike-slip fault zone.

The geometry of the greenstone belts is linear, curved or irregular and locally cuspate, which is dictated by the nature of the contacts with the adjacent granite-gneiss complexes (Fig. 2). Linear limits, as in the Guarinos Belt, are due to vertical shear zones or steep contact between the supracrustal rocks and the granitoid intrusions. Curved geometry, as in Pilar de Goiás and the northeast part of Guarinos, is determined by low to moderately dipping lateral ramps of thrust faults, subsequently, tilted or not, by uplift of the granite-gneiss complexes. Irregular geometry derives from two distinct factors. One is the interference of gneiss domes or granitoid intrusions that molded the supracrustal contacts into a sinuous, cuspate trace, as in Crixás. Another corresponds to the combined effect of flat-lying contacts between gneiss domes or intrusions and the supracrustal rocks, or mylonite zones underneath the supracrustals, as in the Serra de Santa Rita and Faina belts.

The Greenstone Belts

The preserved stratigraphic pile of the five belts contains a lower section of metavolcanic rocks, followed by an upper metasedimentary section. Initially, the stratigraphy of the belts in the northern section was described by Danni and Ribeiro (1978); Sabóia *et al.* (1979); Sabóia and Teixeira (1983), and was subsequently modified by Danni *et al.* (1982, 1986); Castro and Magalhães (1984); Jost and Oliveira (1991). In the southern belts, the first model was put forward by Danni *et al.* (1981), and later modified by Tomazzoli (1985); Resende *et al.* (1998).

Despite numerous stratigraphic aspects still to be resolved, it seems that each belt is an independent stratigraphic entity. Differences are more obvious in the upper sedimentary groups. To avoid unnecessary repetition, the common basal metavolcanic content of the five belts will be described, followed by the metasedimentary content of each belt.

Metavolcanic Sequences

The lower stratigraphic sections of the five greenstone belts consist of metavolcanic rocks, starting with metakomatiite followed by tholeiitic metabasalt. So far felsic metavolcanic rocks have been described only in the Serra de Santa Rita Belt.

Metakomatiite and metabasalt sections consist of packages of several flows. Each package is tens to hundreds of metres thick and separated from adjacent packages by chemical sediments or metapelite layers.

Metakomatiite is the main rock-type of the Córrego Alagadinho (Crixás Group), Serra do Cotovelo (Guarinos Group), Córrego Fundo (Pilar de Goiás Group), and Manoel Leocádio (Serra de Santa Rita and Faina) formations (Fig. 3). The komatiitic protoliths were peridotitic or pyroxenitic, with minor sills of olivine gabbro and pyroxenite. Due to deformation and metamorphism, these volcanics were transformed into rocks with variable proportions of serpentine, talc, chlorite, carbonate and actinolite.

In less deformed exposures, primary volcanic features of the metakomatiite flows may still be preserved. Cumulate and spinifex textures, polyhedral joints, pillowed flows, and flow breccias have been described from several exposures (Sabóia 1979; Sabóia and Teixeira 1983; Danni *et al.*, 1981; Tomazzoli, 1985; Kuyumjian and Teixeira, 1982).

Very little is known about the age of these metavolcanic rocks. The metakomatiite flows of Crixás have a Sm/Nd isochron age of 2.825 ± 0.098 Ga ($\varepsilon_{Nd(T)}$ of + 0.6), and a Pb/Pb whole-rock isochron age of 2.728 ± 0.14 Ga (Arndt *et al.*, 1989). These might be minimum ages because the granitoid intrusives of the Anta and Caiamar complexes, which intruded already deformed, metamorphosed and thrusted supracrustal rocks, has zircon U/Pb ages of 2.842 ± 0.006 Ga and 2.820 ± 0.006 Ga, respectively (Queiroz *et al.*, 1999). K/Ar and Ar/Ar isotope data, as well as Rb/Sr mineral isochrons, for rocks of the Crixás greenstone yielded ages between *c.* 750 - 505 Ma (Fortes *et al.*, 1995, 1997) indicating the strong imprint of the Brasiliano thermal events in the area. A K/Ar age of 557 ± 11 Ma is reported for hydrothermal biotite from the Maria Lázara gold deposit in Guarinos (Pulz, 1995).

The rocks on which the first eruptions took place are unknown. The upper contact with the metabasalt is sharp, if not tectonic. The basal parts of the metabasalt sections can eventually contain peridotitic or pyroxenitic flows. The sharp contact between both sections suggests that the ultramafic volcanism underwent sudden interruption, with sporadic recurrence during the initial phases of basaltic volcanism.

Metabasalt makes up the Rio Vermelho (Crixás Group), Serra Azul (Guarinos Group), Cedrolina (Pilar of Goiás Group) formations and the lower member of the Digo-Digo Formation of the Serra de Santa Rita and Faina belts (Fig. 3). The typical metabasalt is amphibole schist and phyllite with fine-grained granoblastic and nematoblastic texture. The main mineral assemblages are, in general, represented by ferro-actinolite or ferro-tschermakite followed by albite or oligoclase, and minor chlorite, clinozoisite, quartz, pyrite or magnetite. Garnet may occur locally, in particular within the contact aureoles of tonalite and trondhjemite intrusions or along shear zones. Proportions of amphibole and plagioclase are compatible with metamorphism of basalt,

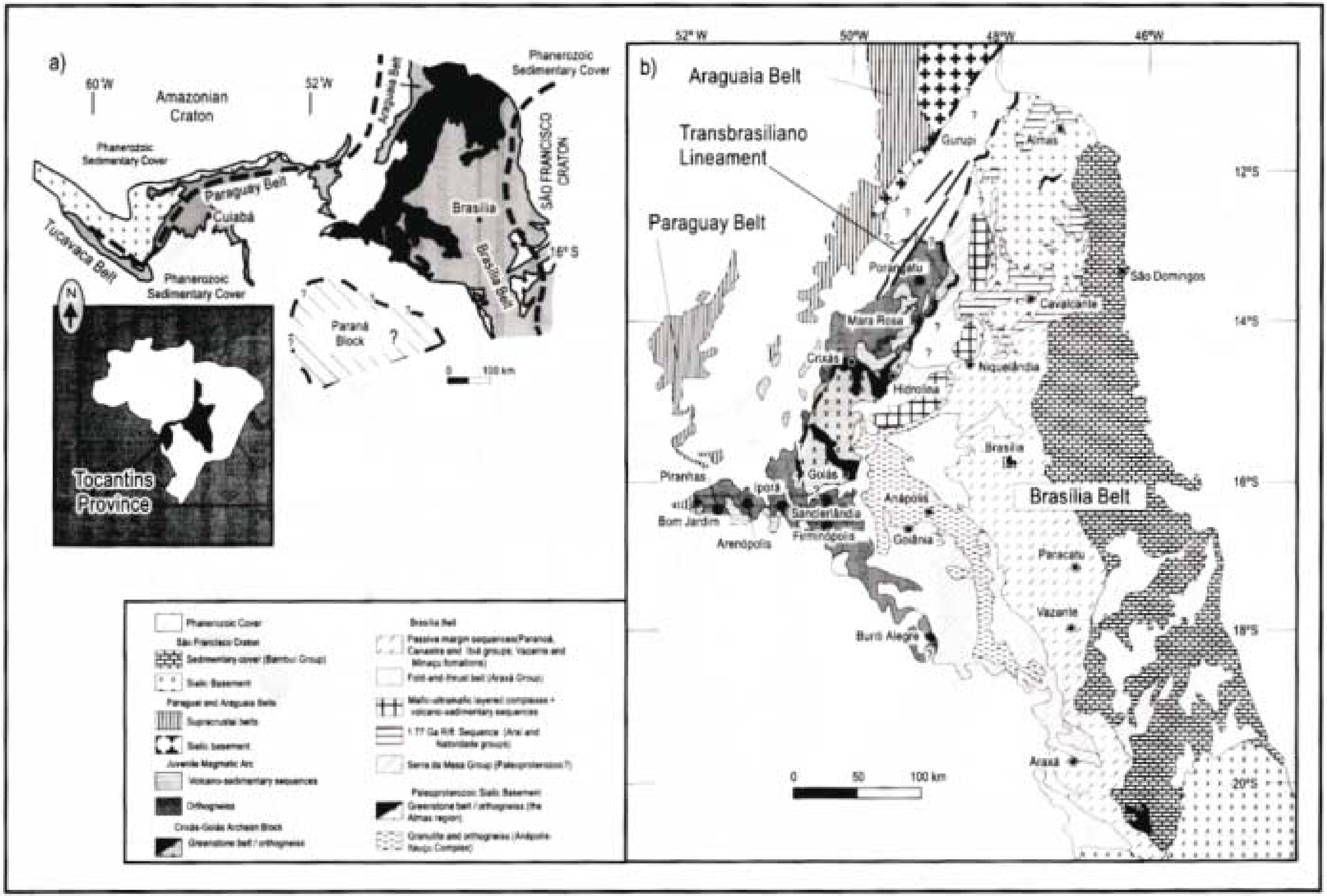


FIGURE 1 - Geological sketch maps of the southern-central part of the Tocantins Province (a) and of the Brasília Belt (b).

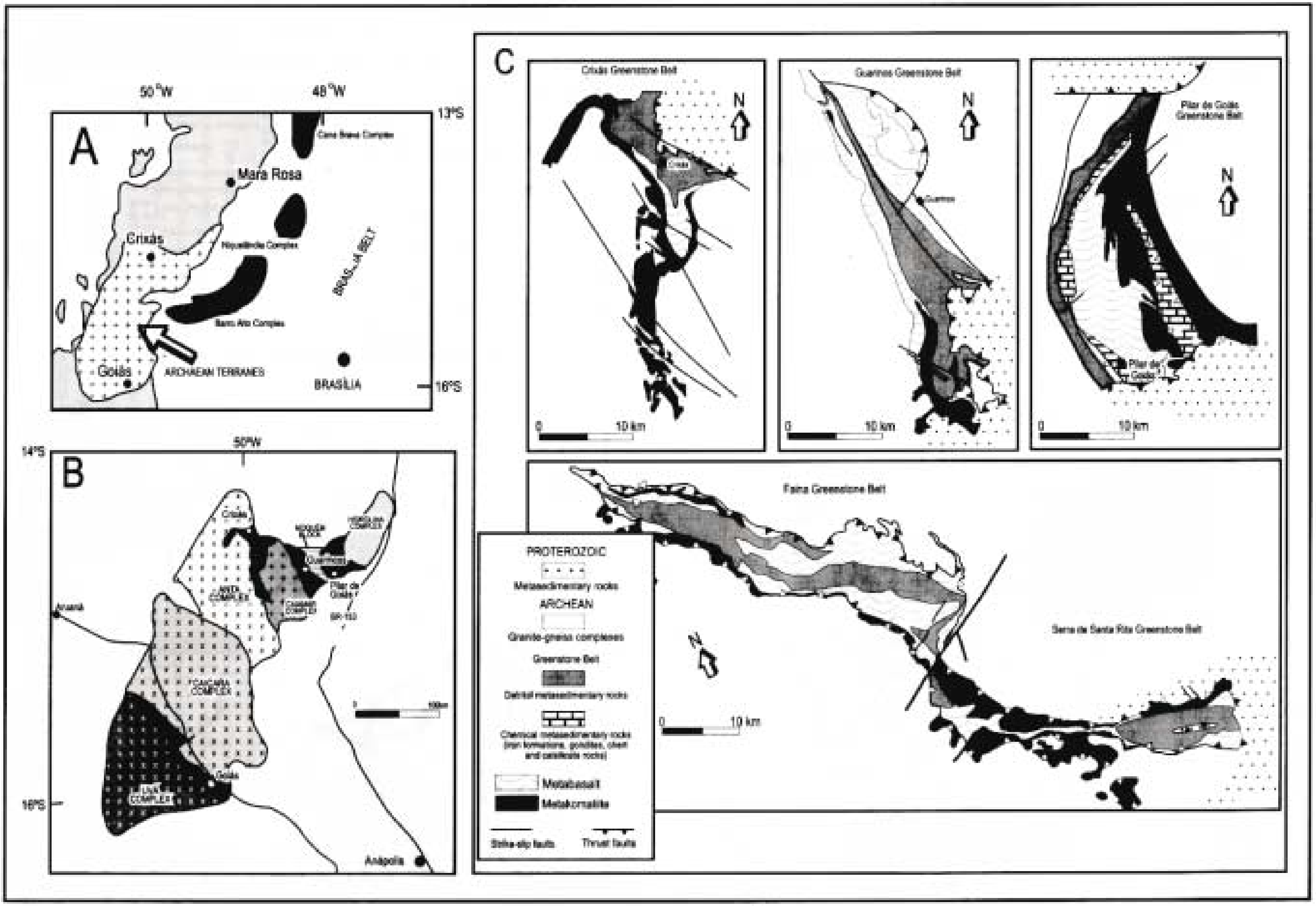


FIGURE 2 - Geological sketch map of the Crixás-Goiás Archean Block. (A) Location map of the Archean Block; (B) Regional map showing the different granitoid complexes and the supracrustal belts; (C) Simplified geological maps of the individual greenstone belts.

locally andesite. More rarely the samples are made up of similar proportions of actinolite and chlorite and derived from komatiitic basalt.

Primary features such as pillows, varioles, vesicles, and orbicules are rare. Felsic metavolcanic rocks (dacite and rhyolite metatuff) comprise the upper member of the Digo-Digo Formation of the Serra de Santa Rita Belt. Metasedimentary intercalations in the metavolcanic packages are represented, in decreasing order of abundance, by metachert, banded iron formation units, manganese formation units, and carbonaceous schist, isolated or in association. The iron formation units occur in the oxide and silicate facies. The oxide facies contain variable proportions of recrystallized quartz, magnetite and hematite, and the silicate facies quartz and grunerite. Manganese formations are rare and in general deeply weathered to a mass of manganese hydroxides, but some regoliths show that the original rocks contain rhodochrosite or spessartite. The chemical deposits can, locally, be associated with carbonaceous schist, as in the Serra de Santa Rita, Crixás and Guarinos belts, indicating exhalative activity simultaneous with pelitic income and proliferation of microrganisms.

Metasedimentary Sequences

The lower volcanic sections of the five greenstone belts give place, to the top, to metasedimentary sequences. The sedimentary history of the northern belts is observed in the Ribeirão das Antas Formation (Crixás), São Patricinho, Aimbé and Cabaçal formations (Guarinos), and Boqueirão and Serra do Moinho formations (Pilar de Goiás) (Fig. 3). In the southern part of the Serra de Santa Rita and Faina belts, metasedimentary rocks form the Fazenda Paraíso and Furna Rica groups, respectively.

Crixás Greenstone Belt

The metasedimentary rocks of the Ribeirão das Antas Formation crop out in the northern part of the belt and host the Mina III gold deposit, one of the most important in Brazil (Castro and Magalhães, 1984; Yamaoka and Araújo, 1988; Magalhães, 1991; Fortes, 1991, 1996; Theodoro, 1995, Jost *et al.*, 1996).

The Ribeirão das Antas Formation consists of a lower sequence of carbonaceous schist with marble layers followed by metarhythmite beds. The gradation between the metabasalt flows of the lower Rio Vermelho Formation and the carbonaceous schist takes place along a section of tens of metres, where both rocks alternate, indicating gradual decline of the volcanic activity and development of an anoxic environment. A few tens of metres above the last basalt flow, carbonaceous schist presents millimetric to centimetric, disseminated, filamentous, and porous fragments interpreted as fragments of felsic pumice.

Towards the top the carbonaceous schist units there occur lenses and layers of white to greysh, massive, laminated, brecciated, or oolitic marble. Laminated marble beds are interpreted as algal mats, and the brecciated facies as talus or karst breccia (Theodoro, 1995) indicating deposition in shallow water. The top of the Ribeirão das Antas sequence consists of siliciclastic metarhythmite beds in sharp contact with the underlying carbonaceous schist. Near the contact, the metarhythmite beds may contain centimetric fragments of carbonaceous schist, suggesting an unconformity between the two.

Individual layers of the siliciclastic rocks are centimetres to decimetres thick with graded bedding, parallel and cross lamination, cut-and-fill and flame structures. Their protoliths were impure sandstone, siltstone and shale forming sequences similar to distal turbidite beds. Metarhythmite sections are eventually interrupted by metric layers of sedimentary breccia with clasts of carbonaceous schist with cut-and-fill structures, suggesting that the distal environment was, from time to time, interrupted by debris flows.

Guarinos Greenstone Belt

The Guarinos Belt contains two stratigraphic successions, of contrasting sedimentary evolution, juxtaposed by a sinistral transcurrent fault parallel to its longer axis. The western half is an antiform with an overturned stratigraphic sequence (Jost *et al.*, 1995). From base to top, the metasedimentary units are the São Patricinho, Aimbé and Cabaçal formations (Fig. 3).

The São Patricinho Formation consists mainly of chlorite-rich schist, quartzite and slate, and occurs only in the southern part of the belt, where it is laterally interfingered with metabasalt flows (Jost and Oliveira, 1991; Jost *et al.*, 1995). The metabasalt flows and the São Patricinho Formation are unconformably overlain by the Aimbé Formation. The diagnostic rock of the Aimbé Formation is a peculiar banded iron formation unit in which magnetite or hematite layers alternate with muscovite (Resende, 1994; Resende and Jost, 1994, 1995). At the base, the iron formation unit is associated with at least four fossil hydrothermal exhalative centres, consisting of concentric alteration haloes, spaced at an average of 7 km along strike. These haloes are made of varied proportions of chloritoid, magnetite, chlorite, muscovite, epidote, tourmaline, and quartz. To the top, the iron formation gradually gives place to the Cabaçal Formation, subdivided into a lower member, consisting of carbonaceous schist with occasional layers of metachert, iron and manganese formation and metabasalt; and a siliciclastic upper member similar to the upper sequence of Crixás (Jost *et al.*, 1995).

The eastern half of the Guarinos Belt is a westward-dipping homocline starting with metabasalt with thin layers of banded iron or manganese formation, and carbonaceous schist. To the top, the metabasalt alternates with and give place to carbonaceous schist.

Pilar de Goiás Greenstone Belt

The Pilar de Goiás Belt is a westward-dipping homocline in which the Boqueirão and Serra do Moinho formations represent the metasedimentary sequence. Both units occur in a narrow zone along the western border of the belt. In the W, the Boqueirão Formation is autocthonous and overlies metabasalt. The succession starts with phyllite and pyrite and tremolite-bearing metachert which grades into calc-silicate rock that alternates with metachert and marble. The lithological succession, as well as major, minor, and trace element geochemistry of the formation indicate that the original chemical sediments were deposited in a deep-sea environment.

The Serra do Moinho Formation consists of metapelite and carbonaceous schist with minor iron or manganese formation units, metachert, tourmaline and chlorite schist; the latter representing mafic metavolcanic rocks (Pulz, 1995). The original lithological arrangement of this unit is difficult to unravel, due to thrusting. It is likely that the metapelite beds of the Serra do Moinho Formation are allochtonous, and that probably only the chemical metasedimentary rocks of the Boqueirão Formation represent the sedimentary history of this belt.

Serra de Santa Rita Greenstone Belt

The Serra de Santa Rita Belt is a narrow syncline with a normal northern limb and an overturned southern one. The metasedimentary rocks occur within the core of the syncline and are represented by the Fazenda Paraíso Group, subdivided, from base to top, into the Limeira and Fazenda Cruzeiro formations (Resende *et al.*, 1998). Contacts are mostly thrust faults, but original relationships are preserved in several outcrops. The Limeira Formation overlies metabasalt flows and felsic metapyroclastic beds of the Digo-Digo Formation and is formed by carbonaceous schist, metachert, banded iron formation units, marble and carbonate-bearing metapelite. The Fazenda Cruzeiro Formation consists of metamorphosed turbidite beds, followed to the top by mica-bearing quartzite and minor metapelite.

Faina Greenstone Belt

The structure and distribution of the metasedimentary rocks of the Faina Belt are similar to those of the Goiás Belt. The Fazenda Tanque, Serra São José and Córrego do Tatu formations of the Furna Rica Group (Resende *et al.*, 1998), encompass the metasedimentary units of this belt. The three units represent a typical shelf environment with two marine regressive cycles separated by a transgression.

An erosional unconformity separates the Fazenda Tanque Formation from the underlying metabasalt. It comprises orthoquartzite with lenses of conglomeratic quartzite and metaconglomerate with abundant clasts of the underlying ultramafic and mafic metavolcanic rocks, metapelite, banded iron formation units and metachert. The Serra São José Formation comprises cross-bedded orthoquartzite with lenses of dolomite marble and carbonate-bearing schist, followed by laminated metapelite with layers of fine-grained orthoquartzite and mica or feldspar-bearing quartzite.

The Córrego do Tatu Formation, at the top of the Furna Rica Group, is in tectonic contact with metapelite beds of the Serra São José Formation, and locally with mafic and ultramafic metavolcanic rocks of the lower stratigraphic sections. The formation contains about 100 m of pure, pink to grey dolomite marble laminated with metachert and hematite-bearing banded iron formation.

The lowest zones of the siliciclastic rocks of the Serra de Santa Rita and Faina belts have Sm/Nd T_{DM} model ages of 3.2 and 3.17 Ga, respectively (Resende, 1998). Trace element data show that more than 90% of the composition of the carbonaceous schist of Santa Rita and the first shelf cycle of Faina can be explained by a mafic and ultramafic source, as clearly indicated by the lower conglomerate beds of the Faina Belt. In both belts, model age values decrease towards the top of the sequences. This is interpreted as the result of mixture of compositionally distinct sources, with increasing felsic contribution.

Granite-Gneiss Terranes

Anta Complex

The extension, structure and lithological variety of the Anta Complex are still not fully known. The geological limits of the complex are shown in Figure 2. It consists of an association of leucocratic, medium to coarse, locally porphyritic granodiorite, tonalite and granite intrusions, in decreasing order of abundance (Vargas, 1992).

Supracrustal rocks of the Crixás Belt locally have apophyses of the adjacent granitoid and contact metamorphism is evidenced by a faint recrystallization of the metakomatiite and metabasalt. The contact has also been described as a narrow, sub-vertical, N20°E to N30°W mylonite zone (Vargas, 1992). The intrusions of the Anta Complex may be isotropic or foliated with the foliation decreasing in intensity towards the centre of the individual bodies. Far from the contact, within the domain of isotropic rocks, the effects of deformation are limited to narrow zones of sub-vertical, N-S shear zones.

Caiamar Complex

The Caiamar Complex, divided into Crixás-Açu Gneiss, Tocambira Tonalite, and Águas Claras Gneiss (Jost *et al.*, 1994a), is situated between the Crixás and Guarinos belts (Fig. 2). The Crixás-Açu Gneiss, in the N, displays well developed compositional banding. Radial dip of the banding and other structural evidences indicate that the gneiss is part of the core of a dome, subsequently truncated by the intrusion of the Tocambira Tonalite. The compositional banding is transected by a discrete N-S vertical foliation given by the orientation of biotite and muscovite.

The Tocambira Tonalite is the central body of the complex (Fig. 2) and intrudes the Crixás-Açu and Águas Claras gneiss, as well as the Crixás and Guarinos belts. The intrusion has an irregular elliptical shape and is made of a fine-grained grey tonalite. The more prominent contact effects of the intrusion comprise narrow drag synforms with silicification and recrystallization of metabasalt and metakomatiite under albite-epidote hornfels facies (Jost *et al.*, 1995), as well as migmatite resulting from partial melting of the Crixás-Açu Gneiss.

The Águas Claras Gneiss occurs in the southern part of the complex (Fig. 2). It can be distinguished from the rocks of the Anta Complex by means of their contrasting radiometric signatures. Metakomatiite, metabasalt and banded iron formation units cover part of the contact of the gneiss with the Tocambira Tonalite. The supracrustal rocks rest on the granitoid pluton on a thin, sub-horizontal mylonite zone with E-W stretching lineation, indicating that they are klippen of probable Neoproterozoic age. In the rare fresh exposures, the gneiss is light grey to pink, medium-grained granoblastic, locally with subhedral K-feldspar phenocrysts up to 1 cm long.

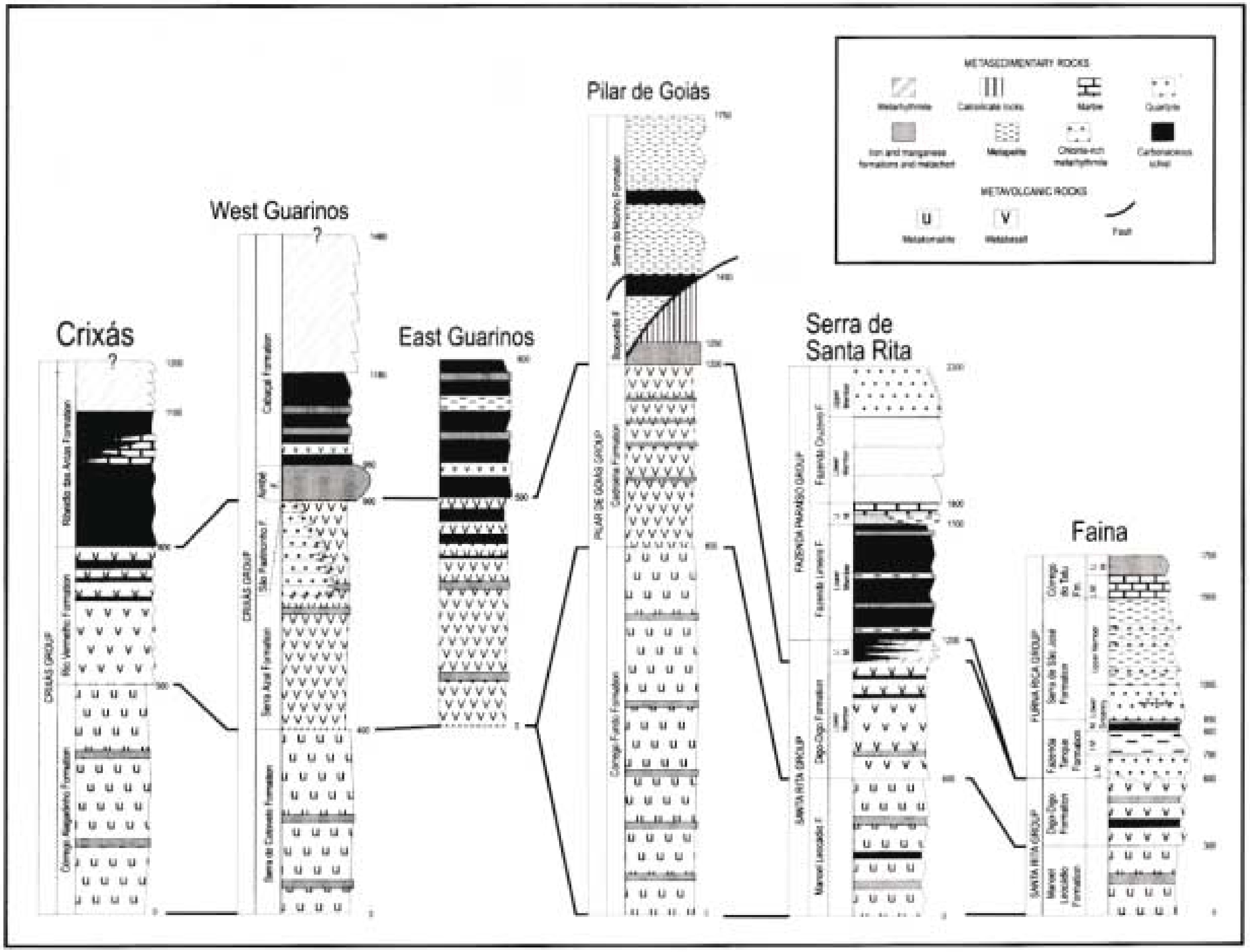


FIGURE 3 - Stratigraphic columns of the greenstone belts of Goiás (LM, IM and UM are Lower Member, Intermediate Member and Upper Member).

All three units of the Caiamar Complex are cut by diabase and microgabbro dykes and less frequently by pegmatite veins. The mafic dykes underwent the same deformational phases as the gneiss and contain upper amphibolite facies parageneses, locally showing retrograde metamorphism to the greenschist facies. This contrasts with the single greenschist facies shown by the dykes within the Tocambira Tonalite, and indicates that the gneiss and the tonalite underwent different deformation histories. Contact relationships of the tonalite with its country rocks indicate that the intrusion took place after supracrustal deformation and regional metamorphism.

Moquém Block

The Moquém Block (Danni and Ribeiro, 1978; Jost *et al.*, 1994b) is limited by the Guarinos and Pilar de Goiás greenstone belts (Fig. 2), and by Proterozoic metasedimentary rocks to the S and N. The limits are either thrust faults, such as in the northern and southern contacts with the Proterozoic units, and the northwestern contact with the Guarinos Belt; or strike slip faults, as ocurs at the southwestern contact with the Guarinos Belt, and the eastern contact with the Pilar de Goiás Belt.

In decreasing order of abundance, the Moquém Block is composed of tonalite, granite, and granodiorite-gneiss, intruded by a diabase dyke swarm (Danni and Ribeiro, 1978; Vargas 1992; Jost *et al.*, 1994b). Pegmatite bodies are rare but occur parallel to the local foliation.

The tonalite-gneiss is banded, light to dark grey, medium to coarse-grained, and is similar to the Crixás-Açu Gneiss of the Caiamar Complex. The granodiorite-gneiss is pink, medium to coarse grained, with prominent foliation and mineral stretching lineation. K-feldspar phenoclasts appear in a fine to medium-grained recrystallized matrix of plagioclase, quartz, and biotite, with minor hornblende, titanite, and magnetite. The granite-gneiss is white to pink, very coarse, with a prominent fabric given by the orientation of biotite and muscovite, stretched K-feldspar phenoclasts and quartz ribbons. The composition of the gneiss varies between that of granite and alaskite.

Hidrolina Complex

The Hidrolina Complex (Fig.2) has a pear-shaped outline oriented NE-SW and is limited by the Pilar de Goiás Belt in the SW, and by Proterozoic metassedimentary and metavolcanic units in the SE, E, and W. The complex consists of granodiorite to granite-gneiss, with subordinate tonalite-gneiss. The granodiorite to granite-gneiss is leucocratic and medium to coarse-grained. A prominent foliation appears near the contact with the Pilar de Goiás Belt. Towards the NE, the intensity of this foliation decreases and its dip changes from an average of 30°W to almost horizontal. Banded tonalite gneiss occurs as a narrow zone at the southwestern margin of the complex.

The prominent border foliation of the complex is parallel to its contact with the Pilar de Goiás Belt supracrustals, dipping to the W. This indicates that the Hidrolina Complex is the core of a dome (Danni *et al.*, 1986; Vargas, 1992). The rim zone commonly contains lenses of ultramafic rocks, frequently with layers of iron formation. The narrow lens shape of these bodies and their direction and dip parallel to the border foliation of the complex suggest that these can be either tectonic imbrications or infoldings of supracrustal rocks within the contact zone.

Caiçara and Uvá Complexes

The Caiçara and Uvá complexes (Jost *et al.*, 1999; Figure 2) are poorly known due to the lack of detailed mapping. The northern limit of the Caiçara Complex is difficult to establish, due to deep weathering. However, the contact between the Anta and Caiamar complexes is neatly defined by radiometric data. Granodiorite, tonalite, and quartz-diorite are the main rock types, cut by a mafic dyke swarm and small mafic and ultramafic intrusions (Danni *et al.*, 1981). The Uvá Complex is limited by the Goiás and Faina greenstone belts in the N, and by Proterozoic units in the W, S and E. The complex consists of tonalite gneiss and granite-granodiorite intrusions.

Near the contact with the Serra de Santa Rita and Faina belts, the intrusions of both the Caiçara and Uvá complexes commonly contain xenoliths of mafic and ultramafic supracrustal rocks (Resende *et al.*, 1998). In spite of this, the contact between the intrusions and the supracrustal sequences has vergence to the NE. Shallow angle thrust faults show that the supracrustal sequences are allochtonous.

Geochronology

Geochronological data for the Archean granite-gneiss terranes are still scarce to support an evolution model. The first radiometric ages of Hasui and Almeida (1970) in amphibolite from these terranes gave K/Ar ages of 3.067 ± 0.216 Ga and 954 ± 48 Ma. The geochronological studies that followed during the 70s and 80s were based mainly on the Rb/Sr isochron method and, only more recently, Sm/Nd and U/Pb SHRIMP data became available for the southern and northern parts of the block, respectively. Rb/Sr ages for several of the granite-gneiss units vary from *c.* 2.9 to 2.5 Ga (Table 1).

A Pb/Pb whole-rock isochron age of 2.481 ± 0.188 Ga is also reported for the granodiorite of the Caiamar Complex (Tassinari and Montalvão, 1980).

The large range of Rb/Sr isochron ages may suggest that these terranes evolved according to several events of granitogenesis during most of the late Archean. However, it is also possible that the younger ages of *c.* 2.6 - 2.5 Ga represent only minimum age estimates. Recent zircon and titanite U/Pb SHRIMP data for several rock units of the northern part of these terranes supports this assumption (Queiroz *et al.*, 1999).

In the northern complexes, the oldest units belong to the Caiamar Complex. The Águas Claras and the Crixás-Açu gneiss have zircon ages of 2.844 ± 0.007 Ga and 2.817 ± 0.009 Ga, respectively, while the Tocambira Tonalite intrusion has a U/Pb age of 2.842 ± 0.006 Ga. Zircon from these three units has older cores with ages ranging between 3.076 ± 0.006 Ga and 2.931 ± 0.007 Ga, indicating that they are contaminated with older crust. The Crixás-Açu Gneiss has two groups of metamorphic titanite. The older has an age of 2.711 ± 0.034 Ga and the younger yield the age of 2.011 ± 0.0015 Ga. The latter is consistent with other structural and geochronological data of the region (Queiroz *et al.*, 1999) related to a prominent Paleoproterozoic thrusting event that affected the Archean rocks.

Tonalite and granodiorite of the Anta Complex were emplaced at 2.820 ± 0.006 Ga, with inheritance ages lying between 3.174 ± 0.004 Ga and 2.948 ± 0.006 Ga. This indicates that the age of *c.* 2.5 Ga obtained by the Rb/Sr method represents minimum ages due to superimposed metamorphic events. The granodiorite core of the Hidrolina Complex gave an age of 2.785 ± 0.005 Ga. Granodioritic gneiss from Moquém Block show a magmatic age of 2.707 ± 0.004 Ga and the granitic gneiss is 2.711 ± 0.003 Ga old.

The granitogenesis in the northern section of the Archean terranes, therefore, occurred between *c.* 2.85 Ga and 2.7 Ga, suggests that the emplacement of these granitoid plutons resulted from several episodes of intrusion that took place after the deposition, deformation and tectonic transport of the Crixás, Guarinos and Pilar de Goiás greenstone belts. Metamorphic zircon ages of 590 ± 10 Ma for rocks of the Moquem Block also reveal the imprint of the Brasiliano Orogeny (Queiroz *et al.*, 1999).

U/Pb data are still not available for rock units of the Uvá and Caiçara complexes. More recently, Pimentel *et al.* (1996) and Potrel *et al.* (1998) demonstrated, in regional Sm/Nd studies of the gneissic terrains to the S and N of the Serra de Santa Rita greenstone belt, that T_{DM} model ages are between *c.* 3.2 and 3.0 Ga. These were interpreted as maximum ages for the protoliths of these gneissic and granitoid rocks. One Sm/Nd whole-rock isochron for the Uvá granite gave the age of 2.851 ± 0.18 Ma and $\varepsilon_{Nd(T)}$ of + 0.3 (Pimentel *et al.*, 1996).

The sialic basement of the Brasília Belt

Two main areas of the central-eastern part of the Tocantins Province have been traditionally interpreted as autochtonous exposures of the sialic basement of the Brasília Belt: (i) the Almas-Dianópolis Terrane in southeastern Tocantins and northeastern Goiás, and (ii) the Anápolis-Itauçu high-grade complex in central Goiás (Fig. 1). In the Almas-Dianópolis area the Paleoproterozoic basement orthogneiss, granitoid and supracrustal sequences were only slightly affected by the Brasiliano deformation and metamorphism. In this region, therefore, a clear unconformable relationship is preserved between the sialic basement and younger Proterozoic cover sequences. This is not the case in the Anápolis-Itauçu area, where older Paleoproterozoic units have been intensely reworked by the Brasiliano Orogeny in such a way that original stratigraphic relationships were completely obliterated.

The Almas-Dianópolis Terrane

This comprises a large area in the northeastern part of the Tocantins Province and is formed mainly by orthogneiss, deformed granitoid and some supracrustal units. These basement units appear below the uplifted Neoproterozoic low grade sedimentary formations of the Paranoá and Bambuí groups, the late Paleoproterozoic Araí Group and its possible northern equivalent, the Natividade Group (Figs. 1, 4 and 5). Farther to the N, they are covered by Phanerozoic strata of the Parnaíba and Sanfranciscana basins.

Geological knowledge of this area is still poor. First descriptions resulted largely from small-scale surveys (Barbosa *et al.*, 1969; Costa *et al.*, 1976; Araújo and Alves 1979; Correia Filho and Sá 1980; Fernandes *et al.*, 1982). More recently, several small areas were studied in greater detail, such as the Almas-Dianópolis (Costa, 1984; Cruz 1993; Cruz and Kuyumjian, 1998), Almas-Conceição do Tocantins (Padilha, 1984), São Domingos (Teixeira *et al.*, 1982; Faria *et al.*, 1986); and Cavalcante-Teresina de Goiás-Nova Roma (Botelho, 1992; Botelho *et al.*, 1993, and unpublished results). The main rock units recognized are granite-gneiss, volcano-sedimentary sequences (Riachão do Ouro Group and São Domingos Sequence), the Ticunzal Formation, mafic-ultramafic plutons, and tin granite intrusives.

Riachão do Ouro Group

Several narrow, approximately N-S belts of volcanic and sedimentary rocks are exposed in southeastern Tocantins, mainly in the Almas, Dianópolis and Natividade areas (Costa *et al.*, 1976; Correia Filho and Sá, 1980), from where they extend southwards to Conceição do Tocantins (Padilha, 1984) (Fig. 5). Large granitoid bodies, their contacts being intrusive or faulted, separate them. Since the initial suggestion by Costa *et al.* (1976), these supracrustal rocks are believed to be Archean greenstone belts. However, the few age determinations available for the associated granitoid bodies indicate that they are probably Paleoproterozoic (Hasui *et al.*, 1980; Costa, 1984).

Costa (1984) coined the name Riachão do Ouro Group to designate these supracrustal rocks. It comprises the Córrego do Paiol and Morro do Carneiro formations (Cruz and Kuyumjian, 1998). The lower Córrego do Paiol Formation is composed of metabasalt, locally with pillow structures, with rare occurrences of tremolite-chlorite schist. The metabasalt is mainly massive, dark green fine-grained amphibole-plagioclase-chlorite-epidote schist with a high-Fe tholeiite signature; whereas the tremolite-chlorite rocks are high-Mg tholeiites with komatiitic affinity (Cruz and Kuyumjian, 1996). The upper Morro do Carneiro Formation is a monotonous sequence of phengite phyllite, with carbon rich layers and less common chlorite-bearing layers. It also includes abundant beds of hematite-magnetite banded iron formation units, especially in the Conceição do Tocantins area (Thomsen and Kuyumjian, 1994), metachert and tourmaline-rich quartzite. Felsic tuff beds are exposed S of Almas, close to the contact with the Córrego do Paiol Formation.

São Domingos Sequence

The São Domingos Sequence (Teixeira *et al.*, 1982; Faria *et al.*, 1986) crops out in a narrow N-S strip around São Domingos, Goiás, between carbonate rocks of the Sete Lagoas Formation of the Bambuí Group to the W, and sandstone of the Cretaceous Urucuia Formation and Holocene deposits to the E. To the N and S the supracrustal rocks are in faulted contact with basement granite-gneiss rocks.

The sequence comprises mafic metavolcanic rocks (epidote-actinolite schist, chlorite schist, amphibolite and amphibole schist), felsic metavolcanics and tuff (feldspar phyllite and biotite-muscovite schist), metapelite (phyllite), ferruginous chert and metagreywacke (garnet micaschist, garnet-feldspar schist). Metamorphism is typically of the garnet zone of the greenschist facies. Tonalite and granite intrusions developed contact aureoles with andalusite hornfels. Aplite and pegmatite with cassiterite are related to the granite intrusions. Whole-rock chemistry indicates that the mafic volcanics have low-K tholeiite affinity, whereas the felsic metavolcanic rocks display calc-alkaline trends (Teixeira *et al.*, 1982). So far adequate age constraints are not available. Amphibole of a metabasic rock has a K/Ar age of 2.042 ± 0.143 Ga (Hasui and Almeida, 1970).

Ticunzal Formation

The Ticunzal Formation was first described in the Preto River area, W of Cavalcante, Campos Belos and Monte Alegre de Goiás (Marini *et al.*, 1978; Danni and Fuck, 1981), but it has also been recognized in Colinas do Sul and Teresina de Goiás (Fuck *et al.*, 1988) (Fig. 5). The most typical rock type is graphite schist, usually associated with biotite-phengite-quartz schist, garnet-mica schist and biotite gneiss. The lower part of the formation consists of fine to medium-grained biotite gneiss, mica schist, graphite schist with occasional muscovite schist and leptinite. The upper part comprises quartz schist, graphite-mica schist, tourmaline schist and minor quartzite. The best outcrops are found below the vertical erosion scarps sustained by the basal quartzite of the Araí Group. Restricted and euxinic marine environments with strong biologic activity have been suggested for the deposition of the Ticunzal Formation. The presence of well-formed graphite crystals points to high temperature amphibolite facies metamorphism, although superimposed retrograde lower greenschist facies paragenesis is always observed. Andalusite-bearing thermal aureoles developed around intrusive biotite-muscovite peraluminous granite and anorogenic tin granite. The age of the Ticunzal Formation has not been determined yet, but it is certainly older than *c.* 1.77 Ga, which is the age of tin granite intrusives (Pimentel *et al.*, 1991a).

Granite-Gneiss Terrane

Granite-gneiss rocks are by far the most extensive of these units, underlying most of the region, from Almas-Dianópolis in the N to Cavalcante-Teresina de Goiás in the S.

In the Almas-Dianópolis area two suites of foliated granitoid were recognized (Cruz, 1993; Cruz and Kuyumjian, 1996, 1998) (Fig. 4). The older (suite 1) is intruded into the Riachão do Ouro supracrustal rocks, and comprises mainly amphibole-bearing tonalite bodies, with minor trondhjemite, granodiorite, quartz monzodiorite and quartz diorite. Foliation is weak in the E, but becomes better developed towards the W, where banded structures are common. The younger suite (suite 2) comprises several oval-shaped plutons intruded into amphibole-bearing granitoid and supracrustals. They are coarse to medium grained tonalite and granodiorite bodies, with lesser monzogranite and trondhjemite. Biotite is the typical mafic mineral. Whole-rock geochemistry of both suites indicates calc-alkaline affinities. Suites 1 and 2 were classified as low- and high-Al TTG associations, respectively (Cruz and Kuyumjian, 1996). Major and trace elements suggest parental magmas derived from partial melting of metabasalt in the case of the younger suite, whereas the older rocks were more likely derived from the mantle (Cruz, 1993; Cruz and Kuyumjian, 1996).

Similar granitoid bodies were recognized to the W of Almas and also to the S, around Conceição do Tocantins where amphibole-rich tonalite and granodiorite dominate in the western part, while the eastern areas are underlain mainly by biotite tonalite. Chemistry and mineralogy of these rocks are similar to those described further to the N (Thomsen and Kuyumjian, 1994). Additionally to the two TTG suites, several undeformed intrusive granite and diorite-gabbro bodies of unknown age have also been recognized in the region (Padilha, 1984).

Farther to the S, in the Cavalcante-Teresina de Goiás area (Fig. 5), hornblende-biotite tonalite is the main rock type, quartz diorite and granodiorite being less frequent. Preliminary major and trace element geochemical data indicate that the protoliths were typically calc-alkaline rocks. These calc-alkaline rocks are commonly intruded by several peraluminous granite bodies, such as the Aurumina Granite, a small, foliated, two-mica granite intrusion with graphite-rich enclaves. Mineralogy and chemistry are typical of peraluminous syn-collisional granites. The Aurumina Granite hosts a small gold deposit, related to quartz veins with Cu-Pb-Zn sulphides in a graphite-rich mylonite. Similar intrusions host greisen and pegmatite with cassiterite and tantalite (*e.g.*, small bodies in the Porto Real and Monte Alegre de Goiás areas). In the latter muscovite of a pegmatite has a K/Ar age of 2.129 ± 0.026 Ga and cassiterite displays U/Pb ages between 2.27 and 2.02 Ga (Sparrenberger and Tassinari, 1998). Paleoproterozoic K/Ar ages suggest small thermal influence of the Brasiliano Orogeny in the area.

Despite the lack of more precise age constraints, it is thought that the granite-gneiss terrane of northeastern Goiás and southeastern Tocantins is the westerly extension of the basement rocks that underlie the nearby São Francisco Craton, as seen in São Domingos, Goiás or Correntina, Bahia.

The gneissic and supracrustal rock units in the southern part of this basement exposure are intruded by a number of Sn and In-bearing A-type granites of the Goiás Tin Province. These are 1.77 - 1.58 Ga rift-related granite intrusions, the oldest of which are coeval with the bimodal volcanic rocks of the Araí Group (Pimentel *et al.*, 1991a, 1999a). They are alkali-rich rocks, with high F, Sn, Rb, Th, Y, Nb, Ga and REE contents and form two groups of intrusions (Botelho, 1992): G1 granites are older (*c.* 1.77 Ga), more potassic, with very clear alkalic characteristics; G2 granites are younger (1.6 - 1.58 Ga), have lower K/Na ratios and higher Li, Rb, Sn and Ta, displaying a mild peraluminous character. Sn and In mineralization are always associated with the G2 family. Nd isotopic analysis did not reveal any difference between the two groups, regarding the sources of the original magmas (Pimentel and Botelho, 1998). Both groups display large intervals of $\varepsilon_{Nd(T)}$ values, varying from - 7.9 to - 1.4 for G2 granites and between - 6.1 and 0 for G1 granites. For both groups, Sm/Nd T_{DM} model ages are between 2.6 and 2.0 Ga suggesting remelting of the Paleoproterozoic basement

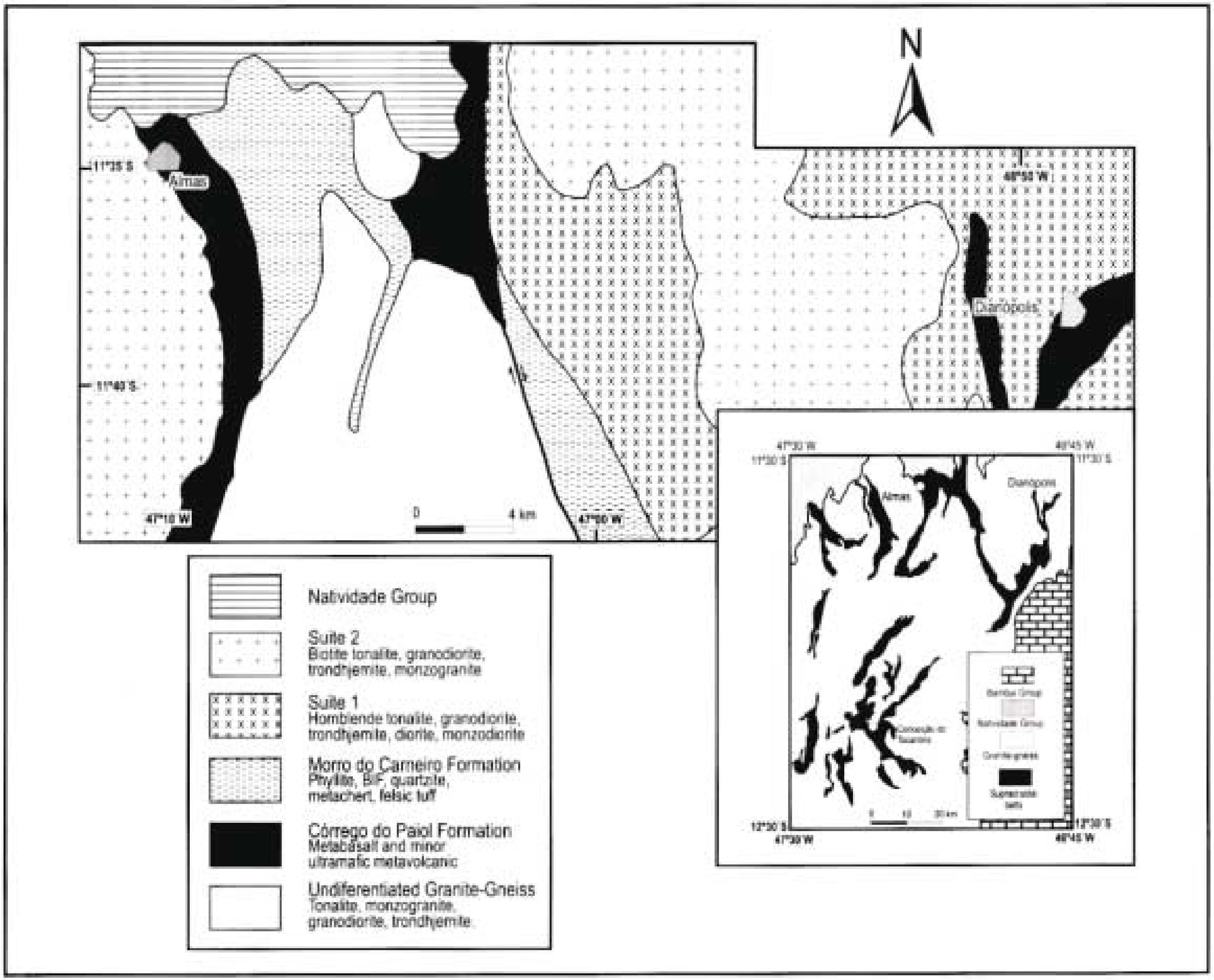


FIGURE 4 - Geological sketch map of the Paleoproterozoic terranes of Almas-Dianópolis, Tocantins. Modified after Padilha (1984); Cruz and Kuyumjian (1998)

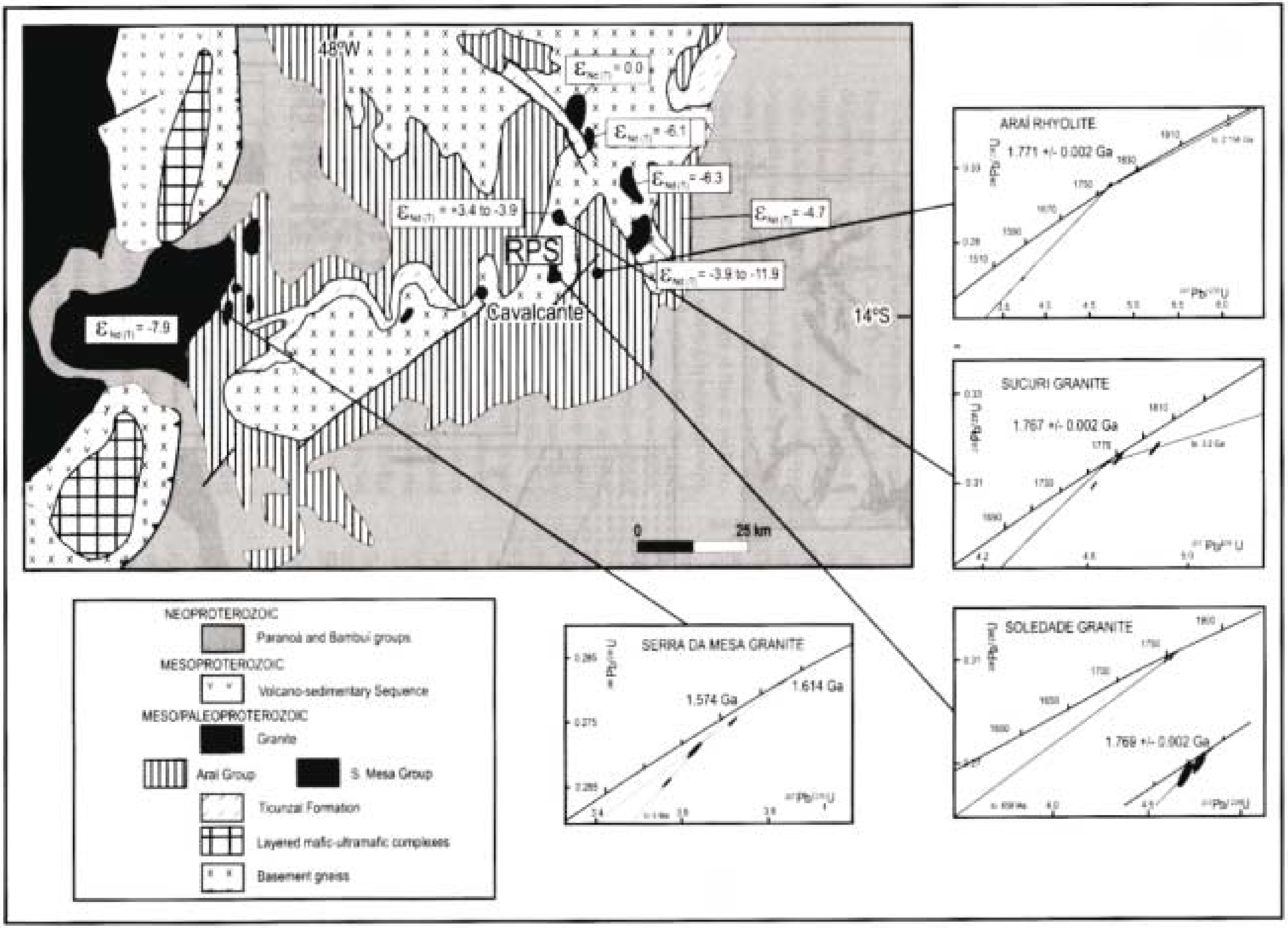


FIGURE 5 - Geological map of the Cavalcante-Teresina de Goiás area, northern Goiás.

rocks. Coeval metabasalt of the Araí Group also displays contamination with older continental crust ($\varepsilon_{Nd(T)}$ of - 5.5 and - 6.0 and T_{DM} values of 2.6 - 2.5 Ga. The preliminary isotope data suggest that the granite intrusives are predominantly derived from remelting of Paleoproterozoic continental crust, with some mixing with mantle-derived mafic magma. This bimodal suite and associated Araí Group sedimentary successions are thought to represent an aborted Paleo-Mesoproterozoic rift system.

Other areas of Paleoproterozoic Basement

Some small areas in the northern section of the Brasília Belt seem to represent Paleoproterozoic basement granite-gneiss and supracrustal sequences, which have been strongly reworked during the Brasiliano Orogeny. For example, strongly sheared Paleoproterozoic gneiss form fault-bounded slivers along the eastern borders of the Cana Brava, Niquelândia and Barro Alto layered intrusions, as well as along the eastern edge of the Goiás Magmatic Arc in the Mara Rosa-Campinorte area. In the latter, a granite-gneiss underlying supracrustal sequences of the Serra da Mesa yielded a U/Pb zircon age of 2.176 ± 0.012 Ga (Pimentel *et al.*, 1997) and granitoid and felsic metavolcanic rocks exposed immediately to the N of the Archean Hidrolina Dome display Paleoproterozoic T_{DM} model ages (2.4 - 2.1 Ga; Oliveira and Pimentel, 1998). Paleoproterozoic T_{DM} model ages between 2.4 and 2.0 Ga were also observed for dioritic/granodioritic rocks exposed E of the Barro Alto Complex (Del'Rey Silva and Pimentel, unpublished results). This agrees with the recent U/Pb age of 2.128 ± 0.015 Ga for gneiss exposed close to the eastern margin of the Barro Alto Complex (Correia *et al.*, 1997b).

The data available at this stage, therefore, seem to indicate that most of the sialic basement of the northern segment of the Brasília Belt is Paleoproterozoic. Late Archean T_{DM} model ages are observed only locally.

The Anápolis-Itauçu Complex

This high-grade complex underlies a large area in the central-southern part of Goiás, along the internal zone of the Brasília Belt. It forms a NW-SE elongated zone (about 260 x 70 km) extending from the vicinities of Itauçu, in the NW, to the Ipameri area in southern Goiás (Figs. 1, 6 and 7). The Anápolis-Itauçu Complex is exposed between metasedimentary successions of the Araxá Group, with mylonite zones separating the high-grade rocks of the complex from the Araxá metasedimentary rocks. Although geochronological data are virtually inexistent for these rock units, the Anápois-Itauçu Complex has been traditionally interpreted as an exposure of the high-grade Archean basement of the Brasília Belt (Marini *et al.*, 1984; Wolff, 1991; Winge, 1995).

Rock-types of different nature (and possibly age) constitute the complex, including: (i) orthogranulite represented by mafic-ultramafic layered bodies, as well as granulite of tonalitic and granodioritic composition, (ii) aluminous granulite, leptinite and garnet gneiss associated with supracrustal rocks such as marble, calcsilicate rocks, quartzite and fine-grained mafic granulite, (iii) narrow strips of volcano-sedimentary sequences consisting of amphibolite, micaschist, felsic metavolcanics, metachert and iron formation units, and (iv) a large number of granitoid intrusions forming NW-SE elongated bodies. The temporal and stratigraphic relationships between the rock units above are poorly known due to the strong and complex deformation patterns, resulting normally in faulted contacts.

Layered Mafic-Ultramafic Complexes

Mafic-ultramafic intrusions constitute an important part of the Anápolis-Itauçu Complex. They form NW-SE elongated bodies, parallel to the regional deformational fabric of the granulitic country-rocks. Examples of mafic-ultramafic bodies are the Goianira-Trindade (Nilson and Motta, 1969), Águas Claras (Nilson, 1992), Santa Bárbara (Silva, 1991; Silva and Nilson, 1990), Taquaral (Silva, 1997) and Serra do Gongomé complexes (Winge, 1995). Except for the Serra do Gongomé intrusion, which is made exclusively of gabbroic and dioritic rocks, pyroxenite, peridotite, gabbro and gabbro-anorthosite and their amphibolite and granulite-facies metamorphic equivalents form these complexes. They frequently display well-developed foliation and mylonitic fabrics. However, relict features such as cumulate textures and igneous layering are found in most of the intrusions, attesting to their original stratiform character (Silva and Nilson, 1990; Silva, 1991, 1997).

Relict igneous pyroxene in the Taquaral and Águas Claras mafic-ultramafic bodies have high Al_2O_3 content which is indicative of crystallization in deeper levels of the continental crust (Nilson, 1992; Silva, 1997). Major and trace element data for these two intrusions indicate that parental magmas had the composition of tholeiitic basalt, whereas the Santa Bárbara gabbro-anorthosite formed from high-Al_2O_3 tholeiitic magma. Their intrusion age is still much debated. Winge (1995) presents a Rb/Sr whole-rock isochron for gabbro, diorite and tonalite of the Serra do Gongomé Complex, indicating the age of 637 ± 19 Ma and initial $^{87}Sr/^{86}Sr$ ratios of *c.* 0.7153. According to Winge (1995), this intrusion is very similar to and can be correlated with the Americano do Brasil intrusion, farther to the W, outside the limits of the Anápolis-Itauçu Complex. The Americano do Brasil mafic-ultramafic complex has a Sm/Nd isochron age of 612 ± 66 Ma with initial ε_{Nd} of + 3.1 (Nilson *et al.*, 1997). The high initial Sr ratio for the Serra do Gongomé intrusion probably means that the original magma was emplaced into and contaminated with older continental crust.

Supracrustal Granulites

Granulitic rocks of supracrustal origin are also abundant in the Anápolis-Itauçu Complex. They are represented by garnet gneiss, sillimanite gneiss, leptinite, kinzigite, calc-silicate rocks, diopside marble, gondite, and garnet quartzite. The fine-grained leptinite and kinzigite have garnet, sillimanite and less frequently cordierite and aluminous spinel. They are normally interpreted as metamorphosed sedimentary rocks, although it has been demonstrated that they can also be the product of high-grade metamorphism of granitoid rocks, as suggested by Winge (1995), or of felsic metavolcanics as pointed out by Silva and Nilson (1990).

Layers of fine-grained mafic granulite have also been interpreted as metabasaltic rocks (Winge, 1995).

This supracrustal rock association has been interpreted as remnants of Archean greenstone belts, metamorphosed in granulite facies (Wolff, 1991; Winge, 1995). However, recent regional geochronological surveys using the Sm/Nd method do not support this interpretation. The studies by Fischel *et al.* (1998) and Pimentel *et al.* (1999b) together with data in Sato (1998) demonstrate that the protoliths of these rocks are younger. T_{DM} model ages are in the range 1.6 - 1.3 Ga and this is interpreted as the approximate upper age limit for the protoliths of these supracrustal granulites. Furthermore, the similarity of Nd isotope composition between the felsic granulite and metapelite of the Araxá Group led Pimentel *et al.* (1999b) to suggest that at least some of the aluminous granulite could be the high-grade equivalents of the Araxá metasedimentary rocks.

Fischel *et al.* (1998) have also shown that the high-grade metamorphic event happened at the end of the Neoproterozoic during the Brasiliano Orogeny. Garnet-whole rock Sm/Nd ages are in the range 630 - 610 Ma, as for example for the small granite body shown in Figure 7 (Fischel *et al.*, 1998, and unpublished results)

Volcano-Sedimentary Sequences

Narrow and discontinuous strips of volcano-sedimentary sequences also occur within the Anápolis-Itauçu Complex (Fig. 7). They received the local names of Silvânia, Rio do Peixe and Rio Veríssimo sequences. Their lithological content is very uniform among all the individual sequences. Metabasalt, amphibolite, amphibole schist, metachert, feldspar-quartz schist, ultramafic rock, quartzite, metapelite, carbonate schist and felsic metavolcanic are described in most of the sequences (Lacerda Filho and Oliveira, 1995).

The Silvânia Sequence is the most extensive of them forming a belt 6 km wide striking NW-SE for approximately 45 km along the eastern margin of the Anápolis-Itauçu Complex. It consists of amphibole schist, amphibolite, and metabasalt with relict igneous textures, quartzite, micaschist, graphite schist and kyanite schist. The metabasalt is tholeiitic to calc-alkaline (Freitas and Kuyumjian, 1995) and it has been suggested that the protoliths were formed in an island arc setting (Lacerda Filho *et al.*, 1991, 1995). Sm/Nd isotope determinations for several samples of micaschist of this sequence indicate T_{DM} model ages ranging from 1.96 to 1.2 Ga (Fischel, unpublished results)

These volcano-sedimentary sequences have been correlated by many authors with the Anicuns-Itaberaí Sequence, which forms a large N-S belt (Fig. 6) to the W of the Anápolis-Itauçu Complex (Barbosa, 1987). This rock unit has been more intensely investigated in its northern part where it includes a lower formation consisting predominantly of metachert, iron formation units, metarythmite, basic and ultrabasic rocks and marble; and an upper formation of chlorite schist and micaschist. Barbosa (1987) suggested a Paleoproterozoic age for the sequence, whereas Nunes (1993) and Winge (1995) compared it to Archean greenstone belts, based on the geochemistry characteristics of some of the ultrabasic rocks.

However, fine-grained felsic gneiss interlayered with amphibolite, both interpreted as metavolcanic rock found in a small supracrustal belt close to but not in physical continuity with the southern end of the Anicuns-Itaberaí sequence, gave a Sm/Nd whole-rock isochron age of 830 ± 99 Ma, with initial ε_{Nd} value of + 3.1 (Pimentel and Fuck, unpublished results). The data suggest that, either this small belt represents part of the Goiás Magmatic Arc and is not correlated with the Anicuns-Itaberaí Sequence or, alternatively, the latter has a Neoproterozoic age.

Granite Intrusions

Several granite intrusions occur within the Anápolis-Itauçu Complex (Lacerda Filho and Oliveira, 1995). They form NW-SE elongate bodies of varied sizes. Some seem to be intrusive into felsic/mafic granulite and also display granulitic mineral assemblages, such as the small metagranite bodies exposed between Anápolis and Goiânia (Fig.7). Others form large intrusions, such as those along the eastern margin of the Anápolis-Itauçu Complex, in tectonic contact with the metasedimentary rocks of the Araxá Group (Fig. 7). The latter have not been metamorphosed in high-grade conditions. The Maratá Granite in the Pires do Rio region and smaller plutons within the supracrustal rocks of the Araxá and Ibiá groups, farther to the S, in the Ipameri area, have a peraluminous character and contain enclaves of metapelitic rocks and quartzite. These granite intrusions have been interpreted as syn-tectonic, possibly collision-related intrusions (Lacerda Filho *et al.*, 1995; Ribeiro and Pimentel, 1998). The Maratá Granite has a U/Pb zircon age of 794 ± 10 Ma (Pimentel *et al.*, 1992) and the Sesmaria Granite, farther to the SE has a Rb/Sr isochron age of *c.* 720 Ma and initial $^{87}Sr/^{86}Sr$ ratio of 0.705. This indicates that at least some of the granite intrusions in the Anápolis-Itauçu Complex are Neoproterozoic.

Preliminary Sm/Nd isotope results for granite bodies within the Anápolis-Itauçu Complex are in the interval between 1.8 and 1.3 Ga, displaying a pattern which is similar to the Araxá Group metasedimentary rocks and felsic granulite of the complex. In their regional Sm/Nd isotope survey, Pimentel *et al.* (1999b) suggest that: (i) the peraluminous granite in the Anápolis-Itauçu Complex is derived from the remelting during the Neoproterozoic of crustal material isotopically similar to either the metasedimentary rocks of the Araxá Group or aluminous granulitic rocks of the complex, and (ii) the supracrustal granulite might represent high grade equivalents of the metasedimentary rocks of the internal zone of the Brasília Belt (Araxá Group). This implies that a large part of the Anápolis-Itauçu Complex could effectively be part of the Brasília Belt itself (only metamorphosed at a higher grade) rather than only representing continental basement. It is very likely, nevertheless, that exposures of Paleoproterozoic rocks truly representing parts of that sialic basement are present within the Anápolis-Itauçu Complex as suggested by some isolated Sm/Nd T_{DM} values between 2.4 - 1.9 Ga for gneissic rock units exposed within the complex. However, the present level of geological and geochronological knowledge of the area does not allow us to easily identify them.

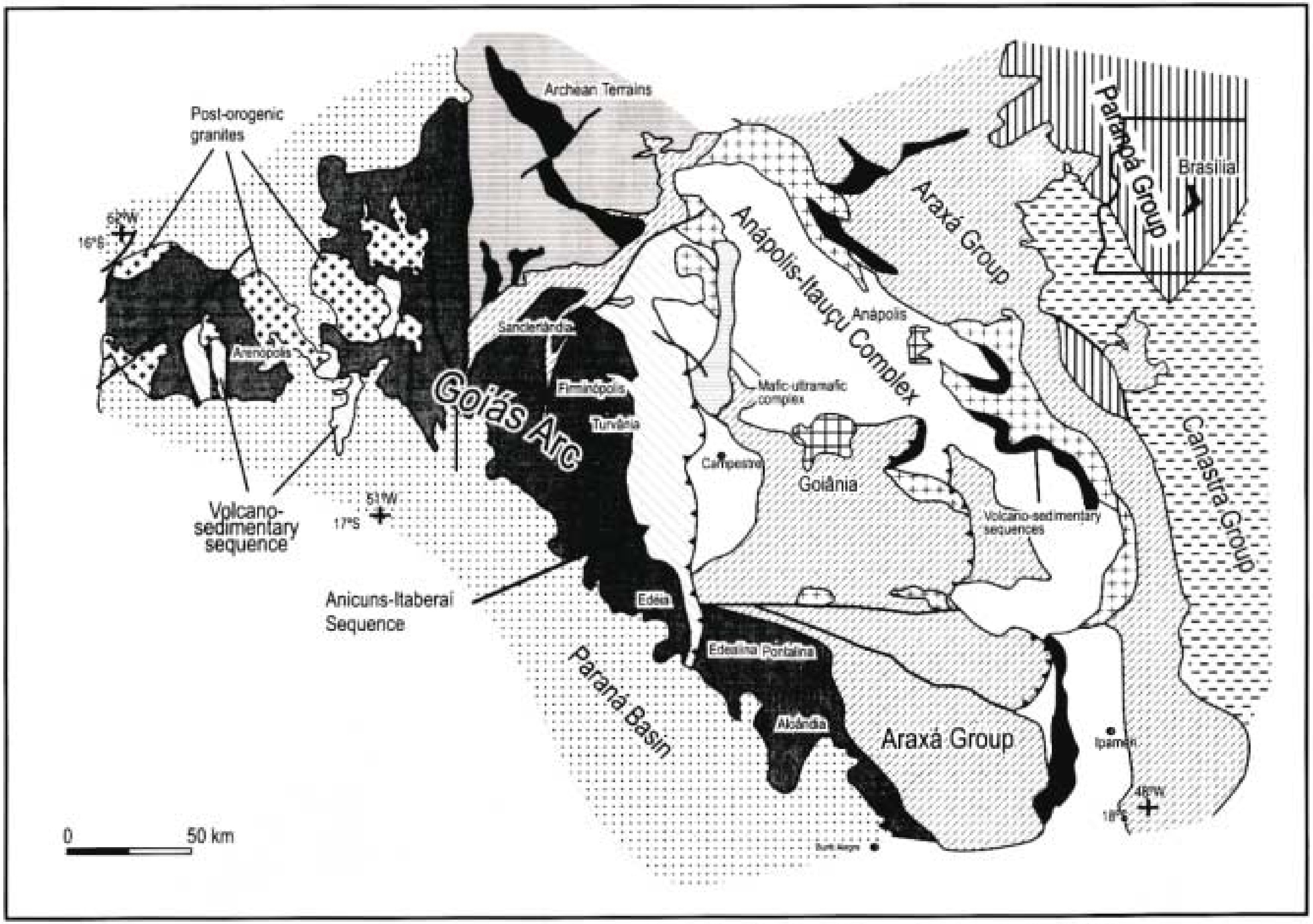


FIGURE 6 - Location map of the high-grade Anápolis-Itauçu Complex, showing also the Goiás Magmatic Arc and the Anicuns-Itaberaí Sequence (after Lacerda Filho et al., 1991).

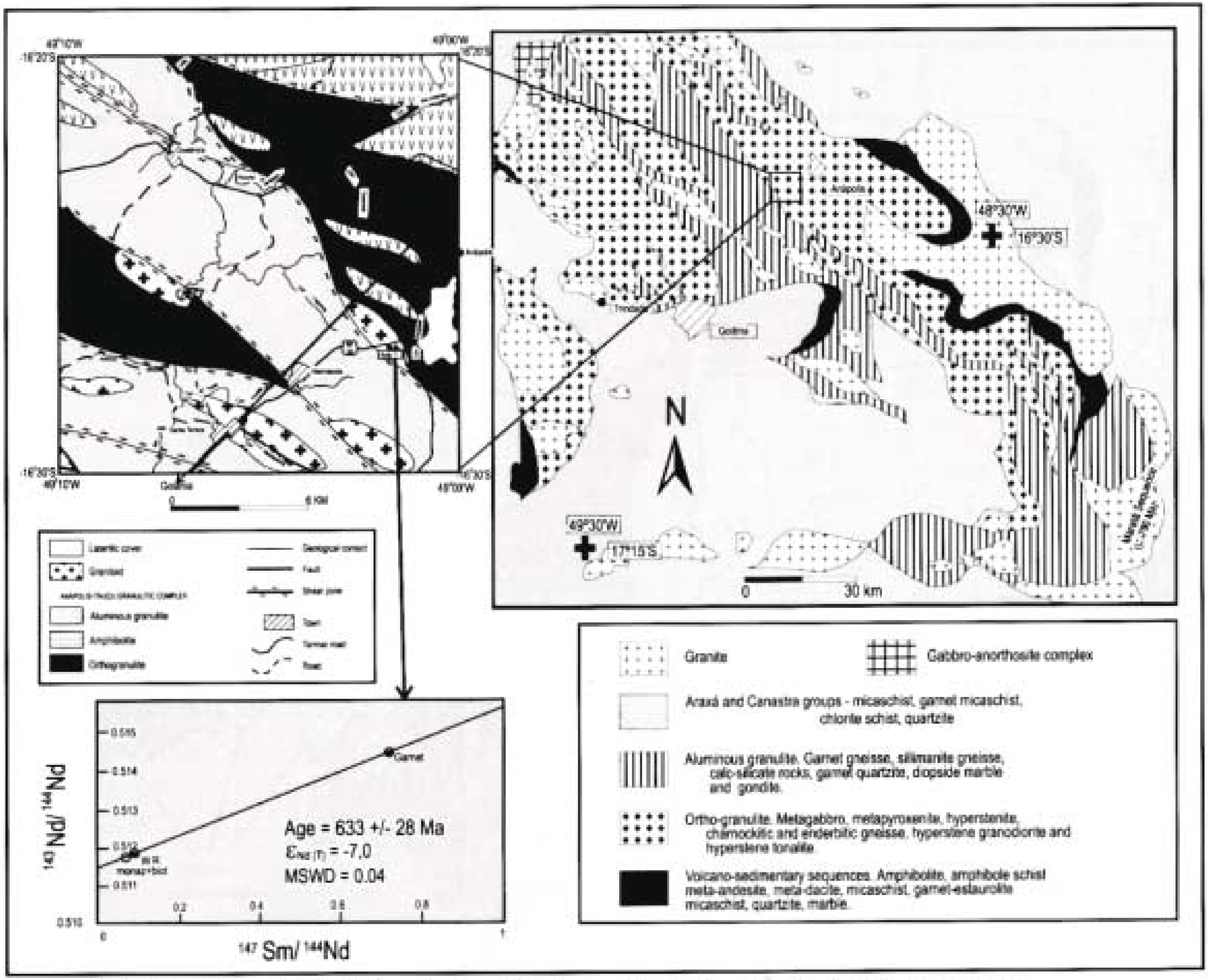


FIGURE 7 - Simplified geological map of the Anápolis-Itauçu Complex and Sm/Nd garnet-whole rock age for a sillimanite- and garnet-bearing granitoid.

The large mafic-ultramafic complexes and associated volcano-sedimentary sequences

The Barro Alto, Niquelândia and Cana Brava complexes are major Paleoproterozoic layered intrusions in central Brazil (Fig. 8). The three separate bodies are part of a 350 km long linear array of layered intrusions interpreted as part of a major continental rift (Ferreira Filho and Naldrett, 1993). To the W, these complexes are in contact with volcano-sedimentary sequences known, respectively, as the Juscelândia, Indaianópolis and Palmeirópolis sequences. The layered complexes were affected by high-grade metamorphism, progressive from amphibolite to granulite facies, at *c.* 780 Ma. It has long been recognized that the original igneous stratigraphy is partially disrupted by tectonism and high-grade metamorphism. On the other hand, several remarkable geological similarities between the three separate bodies have also been used to suggest an originally continuous magmatic structure.

Recent studies have emphasized some aspects related to the metamorphic, petrological and geochronological evolution of these complexes:

i) The layered complexes were affected by a Neoproterozoic high-grade metamorphic event, progressive from amphibolite to granulite facies. This fact was first recognized by Ferreira Filho *et al.* (1994) and has been confirmed by recent geochronological data (Suita, 1996; Correia *et al.*, 1996; 1997a; 1997b; Ferreira Filho and Pimentel, 1999).

ii) The complexes comprise two petrologically distinct magmatic systems (a younger upper layered series and an older lower layered series) (Ferreira Filho, 1994; Ferreira Filho *et al.*, 1995, 1998a).

iii) Geochronological data (Correia *et al.*, 1996, 1997a; 1997b; Ferreira Filho and Pimentel, 1999) indicate that the tectonic contact between the two distinct magmatic systems represents a major crustal discontinuity. The eastern, granulite facies layered system, present in the three intrusions, was most likely emplaced at *c.* 2.0 Ga based on U/Pb SHRIMP, Sm/Nd and Re/Os data (Fugi, 1989; Correia *et al.*, 1997a, 1997b), whereas the younger layered series of the Niquelândia and Barro Alto intrusions and also the volcano-sedimentary sequences normally show ages of *c.* 1.3 Ga (Fuck *et al.*, 1989; Suita *et al.*, 1994; Ferreira Filho and Pimentel, 1999), although this has also been interpreted as indicative of metamorphic re-equilibrium (Fuck *et al.*, 1989, Correia *et al.*, 1997a, 1997b). This will be discussed below in further detail.

A brief description of the major layered complexes is presented following the stratigraphic correlations proposed by Ferreira Filho (1998). The description of the Niquelândia Complex will be more detailed whereas the description of the Barro Alto and Cana Brava complexes will focus on similarities and differences with Niquelândia.

The Niquelândia Complex

The geology of the Niquelândia Complex was described by Ferreira Filho *et al.* (1992). To the N, S and E it is in contact with mylonitic gneiss interpreted to be faulted country rocks. To the W, the complex is in fault contact with the Indaianópolis volcano-sedimentary sequence (Danni *et al.*, 1982).

The Niquelândia Complex has the most complete and best-exposed stratigraphic sequence of the three intrusions. A recent turning point was the recognition that it comprises two petrologically distinct magmatic systems (Ferreira Filho, 1994; Ferreira Filho *et al.*, 1995, 1998a). The older system, the lower layered series (LS) occupies the eastern part of the complex, whereas the younger one, the upper layered series (US), forms its western half (Fig. 8).

The LS is divided into a lower mafic zone (LMZ) dominated by gabbronorite with minor pyroxenite, an ultramafic zone (UZ) dominated by interlayered dunite and pyroxenite, and an upper mafic zone (UMZ) dominated by gabbronorite (Fig. 8). Significant geological-petrological features of the LS are: i) The LMZ lies on the eastern border of the intrusion and is interpreted to be a border group; ii) In the UZ dunite (± harzburgite) predominates at the base and layers of websterite become more abundant toward the top. The composition of the most magnesian olivine (Fo 92-93) indicates the primitive nature of the primary magma (Rivalenti *et al.*, 1982; Ferreira Filho *et al.*, 1996); iii) The UMZ shows a progressive trend of Fe enrichment in pyroxene composition compatible with fractionation under low oxygen fugacity conditions. The pyroxene trend for these units is matched by the progressive increase in incompatible trace elements. This compositional variation indicates an open system model involving both crystal fractionation and crustal contamination; iv) Quartz dioritic and felsic intrusions are restricted to and characteristic of the upper part of the UMZ.

The age of crystallization of the LS rocks has been controversial and is still poorly constrained, due mainly to strong Neoproterozoic overprinting. Strongly discordant U/Pb zircon analyses have initially suggested that crystallization age was between *c.* 1.6 and 1.57 Ga (Ferreira Filho *et al.*, 1994). However, more recent U/Pb SHRIMP data based on one concordant zircon analysis suggest a considerably older age of *c.* 2.0 Ga (Correia *et al.*, 1996), which agrees with a poorly defined Re/Os isochron (Correia *et al.*, 1997b).

The US is formed by interlayered leuco-troctolite, anorthosite, gabbro and minor pyroxenite. Pyroxene composition lacks a strong Fe enrichment trend that, together with the presence of interlayered Fe-Ti oxide layers, support the crystallization of the US under higher oxygen fugacity conditions as compared with the LS. Trace element data indicate a distinct parental magma for the US (Ferreira Filho *et al.*, 1995, 1998a). Sm/Nd whole-rock isochrons for gabbro-anorthosites of the US yielded the ages of 1.347 ± 0.069 Ga and 1.352 ± 0.099 Ga (Ferreira Filho and Pimentel, 1999). One of the isochrons was constructed with samples that have their original igneous textures and mineralogy perfectly preserved and, therefore, this date was interpreted as the age of crystallization of the basic protolith. The positive $\varepsilon_{Nd(T)}$ value of *c.* + 4.0 for these rocks is compatible with a depleted mantle source and, therefore, an oceanic character for the original magma, without contamination with continental crust, which contrasts with the original magma of the LS. The different nature of the original magma of the LS and US of the complex, combined with the difference in age that is now apparent from the isotope data, indicates that the Niquelândia Complex and probably also the Barro Alto Complex (see below), represents two distinct intrusions that have been tectonically juxtaposed, perhaps during the Brasiliano Orogeny.

The most important tectonic feature across the Niquelândia Complex is the pervasive presence of high-grade N-N10°E ductile shear zones (Ferreira Filho *et al.*, 1992) with well developed tectonic foliation. The deformation is heterogeneous, as indicated by highly deformed and metamorphosed zones adjacent to areas of poorly developed foliation or completely preserved primary texture. It is characterized by steep (> 45°) westward-dipping, N-N10°E trending foliation with sub-horizontal mineral stretching lineation. Metamorphic banding and isoclinal to tight asymmetric folds are usually associated with the tectonic foliation.

A progressive amphibolite to granulite facies metamorphic event was described by Ferreira Filho *et al.*, (1992, 1998b). The rocks of the Niquelândia Complex show changes in texture and mineral assemblages in response to changing metamorphic grade. Based on mineral assemblages of the metamorphosed mafic rocks, three metamorphic zones are distinguished: amphibolite zone (pl + hbl ± cpx ± grt), amphibolite-granulite transition zone (pl + opx + cpx + hbl), and granulite zone (pl + opx + cpx). Geothermobarometry and mineral stability data indicate P-T conditions of peak metamorphism at about 700 °C and 6-8 kbar in the amphibolite zone and temperatures higher than 800 °C in the granulite zone. A nearly isobar cooling (IBC) path followed peak metamorphism suggesting cooling in the absence of tectonic unroofing. A second and chronologically distinct tectonic event is then required to uplift and expose the high-grade terranes. U/Pb and Sm/Nd geochronological data of Ferreira Filho *et al.* (1994) and Ferreira Filho and Pimentel (1999) seem to support this interpretation. Metamorphic zircon from quartz diorites and mylonites of the LS has U/Pb ages of c. 794 and 778 Ma, whereas Sm/Nd isotope data for garnet indicate a much younger age of 610 ± 32 Ma.

The Barro Alto Complex

The largest of the three complexes has the shape of a boomerang with a northern segment oriented N-S, and a southern part showing E-W structural trends.

In contrast with the Niquelândia Complex, no systematic petrological-geological studies have been carried out in the N-S segment of the Barro Alto Complex. Significant new developments (Ferreira Filho *et al.*, in prep.) are: i) The poorly exposed ultramafic rocks (serpentinized dunite and harzburgite) form an almost continuous ultramafic zone (UZ) some 60 km long. The overall composition of the UZ is similar to the lower part of UZ in Niquelândia. The composition of the most magnesian olivine (Fo 91-92) indicates the primitive nature of the primary magma; ii) The extensive gabbronoritic unit located in the eastern part of the complex follows a reverse fractionation path becoming progressively more primitive toward the top. Reverse fractionation path characterizes the lower border group of several layered intrusions, suggesting that the gabbronoritic unit correlates to the LMZ in Niquelândia; iii) The gabbro-anorthositic unit located in the western part of the N-S segment of the Barro Alto Complex (Fig. 8) is similar to and stratigraphically correlated with the US of Niquelândia; iv) Quartz dioritic and felsic intrusions, characteristic of the upper part of Niquelândia UMZ, are not present in the N-S segment of the Barro Alto Complex but very common in the EW segment.

Systematic petrological study of the primary igneous rocks of the E-W segment of the Barro Alto Complex is restricted to that of Oliveira (1993). The E-W segment consists mainly of gabbronorite with minor interlayered websterite and closely resembles the Niquelândia UMZ. Two relevant features of the primary igneous stratigraphy of the E-W segment of the Barro Alto Complex are: i) The gabbronoritic rocks become more primitive from S to N. Both pyroxene composition and trace element content are similar to the Niquelândia UMZ; ii) Quartz dioritic and felsic intrusions are widespread in the E-W segment of the Barro Alto Complex; the small granite intrusions form oval-shaped bodies elongated E-W and are represented by opx- garnet-sillimanite- kyanite and cordierite-bearing orthogneiss locally with enclaves of mafic rocks; (iii) the gabbro-anorthositic sequence is not as voluminous as in the E-W segment. A common feature is also the presence of supracrustal granulitic rocks associated with the mafic-ultramafic granulite. They are represented by garnet quartzite, calcsilicate rocks and fine-grained mafic granulite (Fuck *et al.*, 1981; Danni *et al.*, 1984).

Thermobarometric studies indicate temperature of 900°C at pressure of 8.5 kbar for peak metamorphic conditions (Moraes, 1997). The emplacement age of the Barro Alto Complex is still poorly constrained and very debated. The eastern gabbronorite bodies are believed to have crystallized at *c.* 1.73 - 1.72 Ga based on discordant U/Pb zircon analyses (Suita *et al.*, 1994). U/Pb data for pegmatitic gabbro-anorthosite of the upper sequence (equivalent to the Niquelândia US) gave ages between 1.35 - 1.29 Ga, which were interpreted as crystallization ages representing, therefore, an important and previously undocumented Mesoproterozoic event of mafic magmatism (Suita *et al.*, 1994). Therefore, it seems that the same age pattern observed for Niquelândia is seen in Barro Alto: the granulitic sequence shows Paleoproterozoic primary age and the amphibolite facies gabbro-anorthosite in the W, crystallized at *c.* 1.3 Ga. Metamorphic ages are also identical to those observed for the Niquelândia rocks. Metamorphic zircon showing ages between 790 and 770 Ma is found not only in the mafic granulite (Suita *et al.*, 1994) but also in granulitic supracrustal rocks.

Felsic granulite, representing one of the small granitic intrusions emplaced into mafic granulite in the E-W section of the complex, have a Rb/Sr whole-rock age of 1.266 ± 0.017 Ga (initial $^{87}Sr/^{86}Sr$ of *c.* 0.7347; Fuck *et al.*, 1989). A similar intrusive garnet and sillimanite-bearing metagranite with mafic enclaves has been dated by Correia *et al.* (1997a, b).

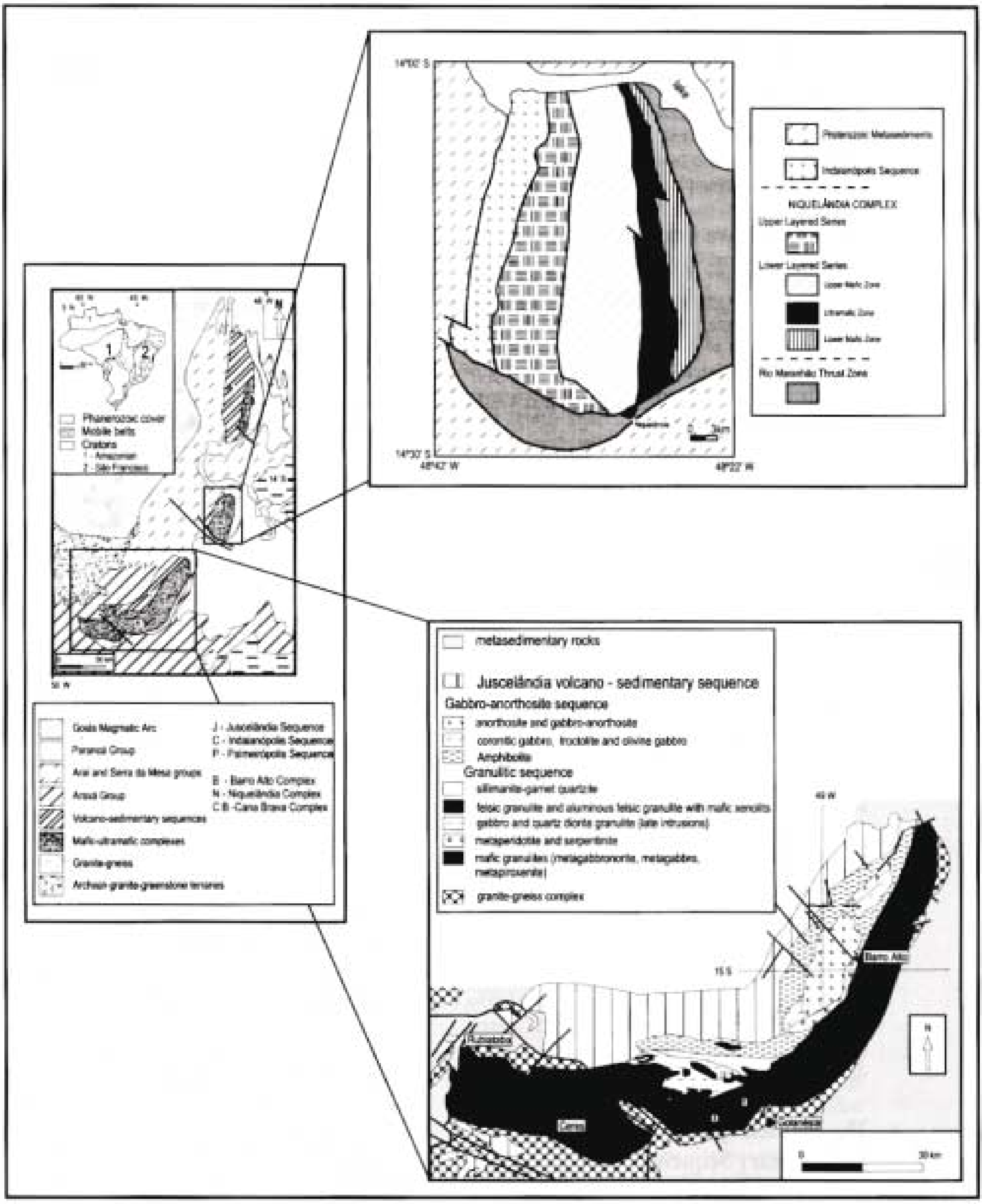


FIGURE 8 - Simplified geological maps of the Niquelândia and Barro Alto complexes.

The data resulted in two discordia diagrams revealing the same complex pattern with upper intercept ages of 1.286 ± 0.013 Ga and 1.302 ± 0.032 Ga and upper intercept ages of *c.* 780 Ma. Contrary to the interpretation of Suita *et al.* (1994) these ages have been assigned to a metamorphic event of the so-called Uruaçuano Cycle. Our contention is that these ages represent igneous crystallization around 1.3 - 1.2 Ga and high-grade metamorphism at *c.* 780 - 770 Ma.

The Cana Brava Complex

Systematic studies of the magmatic evolution of the Cana Brava Complex were carried out by Correia (1994) and Lima (1997). The Cana Brava Complex consists of: a lower mafic zone (LMZ) of gabbronorite with interlayered pyroxenite, an ultramafic zone (UZ) consisting of serpentinized dunite (± harzburgite) with interlayered pyroxenite; and an upper mafic zone (UMZ) with gabbronorite and interlayered pyroxenite. Some relevant features in the Cana Brava Complex stratigraphy are: i) stratigraphic relations of the LMZ are harder to decipher because of intense metamorphic recrystallization; nevertheless, the LMZ lies on the intrusion's eastern border and it is tentatively interpreted to be a border group; ii) the poorly exposed UZ has composition similar to the Niquelândia UZ. The composition of the most magnesian olivine (Fo 89) indicates the primitive nature of the primary magma; iii) the UMZ shows pyroxene compositions, trace element contents and a progressive trend of Fe enrichment in pyroxene composition similar to the Niquelândia UMZ; iv) quartz dioritic and felsic intrusions are restricted to and characteristic of the upper part of the UMZ. This indicates a close similarity with the Niquelândia UMZ; v) there does not seem to be an equivalent of the Niquelândia US in the Cana Brava Complex.

A whole-rock Sm/Nd isochron age of 1.97 ± 0.069 Ga ($\varepsilon_{Nd(T)}$ = - 1.5; Fuji, 1989) represents the best estimate for the crystallization age of the complex. The negative $\varepsilon_{Nd(T)}$ suggests contamination with older continental crust, similarly to that suggested for the Niquelândia LS based on petrological data (Ferreira Filho *et al.*, 1994). Rb/Sr whole-rock and Sm/Nd mineral isochron ages of 1.35 ± 0.035 Ga and 770 ± 43 Ma, respectively, for gabbro of the Cana Brava Complex have been interpreted as metamorphic ages (Correia *et al.*, 1997a)

The Palmeirópolis Volcano-Sedimentary Sequence

The Palmeirópolis Volcano-Sedimentary Sequence (PVSS), occurring at the western border of the Cana Brava Complex, covers an area of about 2500 km^2. Systematic studies of the PVSS include those concerned with geology (Figueiredo *et al.*, 1981; Leão Neto and Olivatti, 1983), petrology and tectonic environment of amphibolite protoliths (Araujo and Nilson, 1987; Araujo *et al.*, 1996a), with metamorphism (Araujo *et al.*, 1995), and with the Zn-Cu-(Pb) volcanogenic massive sulfide deposit associated with the PVSS basal unit (Araujo, 1986, 1996; Araujo and Nilson, 1988; Araujo *et al.*, 1993, 1996b, 1996c). The stratigraphic relationships between the PVSS and adjacent units are not clear due to faulted contacts and lack of reliable geochronological data.

The PVSS is a bimodal tholeiitic sequence divided into three units (Figueiredo *et al.*, 1981). The lower unit, in the eastern part of the sequence, is mainly composed of mafic metavolcanic rocks (amphibolite) with interlayered metamorphosed banded iron formation units and metachert. Mafic and ultramafic dykes and granite bodies are intrusive into the lower unit. Zn-Cu-(Pb) volcanogenic massive sulfide deposits are associated with hydrothermally-altered metamafic rocks of the PVSS lower unit (Araujo, 1986). Petrochemical data for the amphibolite shows that the protoliths of these rocks was mid-ocean ridge tholeiitic basalt (Araujo, 1986, 1996; Araujo and Nilson, 1987).

The intermediate unit is mainly composed of metamorphosed felsic to intermediate volcanic to subvolcanic bodies represented by plagioclase-muscovite-biotite-quartz schist, plagioclase-biotite-quartz schist and muscovite-quartz schist (Figueiredo *et al.*, 1981; Araujo *et al.*, 1996a).

The upper unit, in the western part of the sequence, contains interbedded chemical and pelitic metasedimentary rocks. Mineral assemblages are representative of regional metamorphism within the amphibolite facies. Penetrative deformation of the rocks is strong, and primary structures and textures are rarely preserved. Nevertheless, relict volcanic structures (such as pillows) and textures (phenocrysts and amygdales) are occasionally observed in the lower and intermediate units.

Geochronological data based on Rb/Sr systematics indicate an age of 1.157 ± 0.15 Ga (initial $^{87}Sr/^{86}Sr$ of 0.704) for the PVSS, with some reworking during the Brasiliano Cycle (Girardi *et al.*, 1978). Pb/Pb isotope data for galena from the Palmeirópolis deposit indicates model ages in the range of 1.27 to 1.17 Ga. These data are not conclusive and do not establish the precise age of deposition of the PVSS, but they do suggest that the sequence is not older than *c.* 1.3 Ga.

The Indaianópolis Volcano-Sedimentary Sequence

The Indaianópolis Volcano-Sedimentary Sequence (IVSS) occurs at the western border of the Niquelândia Complex. Its main geological features were described by Danni and Leonardos (1978); Ribeiro Filho and Teixeira (1981); Nascimento *et al.* (1981); Brod (1988); and Brod and Jost (1991, 1994).

The IVSS is divided into three units. The lower unit comprises a large volume of fine-grained amphibolite intercalated with bands of metachert, biotite-muscovite gneiss and hornblende gneiss. Relict volcanic igneous textures are occasionally preserved in the amphibolite. The intermediate unit is made of felsic to intermediate metatuff with bands of amphibolite, amphibole schist and feldspatic schist. The upper unit is characterized by pelitic and chemical metasedimentary rocks.

Two groups of volcanic rocks were recognized based on chemical composition. The first group, that comprises the amphibolite, is characterized by tholeiitic rocks similar to mid ocean ridge basalt. The second group, including amphibolite and the intermediate to felsic metatuff, has alkaline affinity and was related to Hawaii-type intraplate volcanic rocks (Brod, 1988).

Thermobarometric data are not available to the IVSS, but mineral associations suggest amphibolite facies metamorphic conditions. Recent U/Pb SHRIMP data for meta-rhyolite with preserved igneous textures indicate crystallization age of 1.299 ± 0.039 Ga representing the best estimate so far for the age of these volcano-sedimentary sequences.

The Juscelândia Volcano-sedimentary Sequence

The Juscelândia Volcano-Sedimentary Sequence (JVSS) occurs at the western/northern border of the Barro Alto Complex (Fuck *et al.*, 1981). Studies of the JVSS include analyses of the metamorphism and deformation that affected the sequence (Moraes, 1992), of the chemical and tectonic characterization of amphibolite (Danni and Kuyumjian, 1984; Moraes and Fuck, 1994), the relations between the Barro Alto Complex and the JVSS (Winge and Danni, 1994) and a brief description of the sulfide occurrence associated with the basal unit of the JVSS (Sousa and Frizzo, 1995).

The JVSS is divided into four units (Fuck *et al.*, 1981; Moraes, 1992; Moraes and Fuck, 1992a, 1994). The lower unit consists of fine-grained amphibolite with bands of metachert. The chemical composition of the amphibolite suggests that the protolith was ocean floor basalt (Danni and Kuyumjian, 1984; Kuyumjian and Danni, 1991; Moraes and Fuck, 1992b). Based on trace element geochemistry of the amphibolite, Moraes (1997) suggested that the volcanism started in a continental rift environment that evolved to the formation of an ocean floor.

The lower part of the intermediate unit consists of biotite gneiss with minor metachert, which are replaced by muscovite-biotite gneiss. The latter is a microcline and oligoclase-rich rock with relict phenocrysts of feldspar and blue quartz and interpreted to be originated from granitoid and felsic volcanic rocks. The upper unit of the JVSS consists of pelitic and chemical metasedimentary rocks, with intercalated layers of amphibolite.

Thermobarometric data indicate amphibolite facies metamorphic conditions for the JVSS (Moraes 1997). Ferreira Filho *et al.* (1999) recognized granulitic facies mineral assemblages in mafic rocks from the JVSS. These mafic rocks show primary volcanic pillow structures, still preserved in low-strain zones. The granulitic paragenesis indicates that, at least at the eastern part of the sequence, the metamorphic peak conditions are higher than what was previously suggested (Moraes, 1997). The only geochronological data available so far for rocks of the JVSS is a whole-rock Rb/Sr isochron with the age of 1.33 ± 0.067 Ga (initial $^{87}Sr/^{86}Sr$ of *c.* 0.7082; Fuck *et al.*, 1989) for a small granite body within the supracrustal rock units.

The Goiás Magmatic Arc

Two main areas of Neoproterozoic juvenile crust have been identified in western and northern Goiás. These are here referred to as the Arenópolis and Mara Rosa arcs, respectively. Most of the arc magmatism is represented, at the present level of erosion, by metaplutonic rocks ranging in composition from tonalite to granodiorite. They are exposed between narrow NNW to NNE belts of Neoproterozoic arc volcano-sedimentary sequences. In the S, the litho-stratigraphic units comprising the Goiás Magmatic Arc are particularly well known in the Bom Jardim de Goiás (Seer, 1985), Arenópolis-Piranhas (Pimentel, 1985; Pimentel *et al.*, 1985, 1991b; Pimentel and Fuck, 1986, 1987) and Jaupaci-Iporá (Amaro, 1989; Rodrigues, 1996; Gioia, 1997; Rodrigues *et al.*, 1999). In northwestern Goiás the arc units have been investigated mainly in the Chapada-Mara Rosa area by Arantes *et al.* (1991); Kuyumjian (1989, 1994); Viana *et al.* (1995); Junges (1998); Pimentel *et al.* (1997).

The Arenópolis Arc

The Arenópolis Arc underlies large areas in southwestern and southern Goiás, extending from the vicinities of Bom Jardim de Goiás in the W to Buriti Alegre in the SE (Figs. 1, 6 and 9).

The supracrustal and orthogneissic units that constitute this section of juvenile Neoproterozoic continental crust were juxtaposed to each other along important NNE to NNW stike-slip faults, which are part of the continental-scale Transbrasiliano Lineament, running across the region up to the northeastern coast of Brazil. Therefore, the establishment of the original stratigraphic and temporal relationships between the different rock units is not always easy. Large late to post-orogenic K-rich granite bodies were emplaced into the arc rocks at the end of the Neoproterozoic and beginning of the Paleozoic.

Calcic to Calc-alkaline Orthogneiss

These are hornblende and biotite-bearing metatonalite and metagranodiorite (*e.g.*, the Arenópolis, Sanclerlândia, Matrinxã , Firminópolis and Turvânia gneiss, and the Choupana Granitoid). They show mineral assemblages indicative of metamorphism under epidote amphibolite facies and commonly display relict igneous textures and structures, such as mafic enclaves, porphyritic textures and magma mixing features.

Major and trace element data suggest that the igneous protoliths were metaluminous calcic to calc-alkaline with high CaO, MgO, P_2O_5 and Al_2O_3 (Pimentel, 1991). They show clear LILE enrichment in relation to Nb, Y and Yb, which is normally interpreted as diagnostic features of igneous rocks formed in magmatic arcs. Low Rb/Sr, Zr, Nb, Y, LREE and Ga of the Arenópolis, Sanclerlândia and Firminópolis gneiss are comparable to values found in very primitive M-type granitoid of intraoceanic island arcs. The Matrinxã rocks show characteristics, which are transitional between M and I-type granitoid plutons (Pimentel, 1991). Some of their geochemical features, such as low K_2O/Na_2O ratios, high Ni, Cr and Sr contents, and strongly fractionated REE patterns with high La/Yb and Sr/Yb ratios are similar to modern adakites, which form in subduction environments where the subducted slab is young (< 20 Ma) and hot and therefore capable of undergoing partial melting. The adakitic composition of the Neoproterozoic arc granitoid intrusives of Goiás could be explained, therefore,

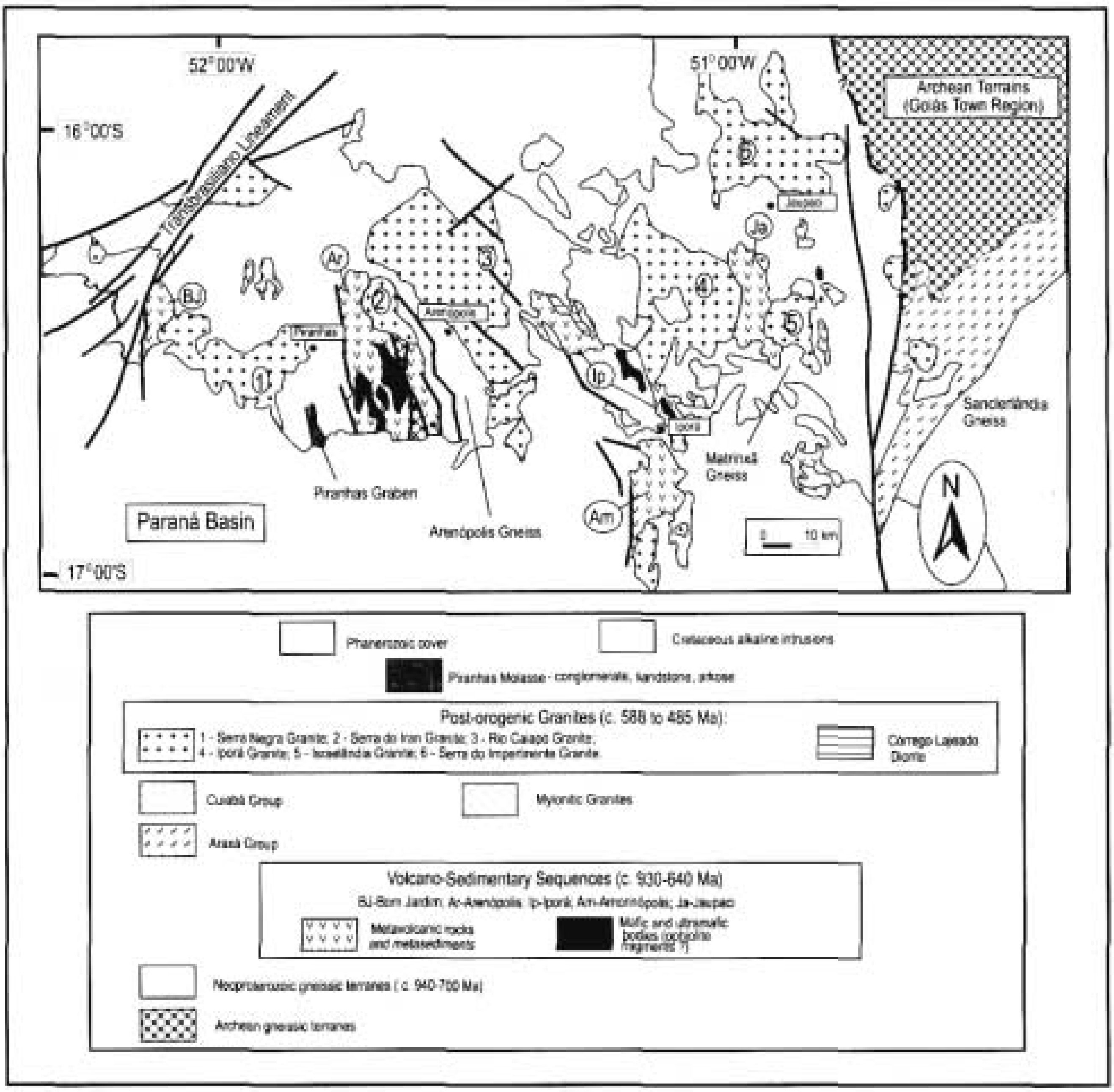


FIGURE 9 - Geological sketch map of the Goiás Magmatic Arc in the Arenópolis region, western Goiás.

using a model in which the original magmas were formed in response to subduction of young and hot oceanic lithosphere constituting small oceanic basins such as back-arc settings, in between island arc systems.

U/Pb, Sm/Nd and Rb/Sr isotope determinations indicate ages between c. 940 and 630 Ma. The Arenópolis Gneiss has a U/Pb zircon age of 899 ± 7 Ma and a whole-rock Rb/Sr age of 818 ± 57 Ma with initial $^{87}Sr/^{86}Sr$ of 0.7042. The positive $\varepsilon_{Nd(T)}$ values between + 1.9 and + 3.2 are indicative of the very primitive nature of the igneous protoliths. Similar geochronological pattern is observed for the other gneissic units (Table 2). The Matrinxã and Sanclerlândia gneiss have poorly defined Rb/Sr whole-rock ages of c. 890 and 940 Ma, respectively, with very low initial Sr ratios (c. 0.7025). U/Pb sphene age of c. 637 Ma for the Arenópolis Gneiss is interpreted as indicative of the Brasiliano metamorphic event.

Mylonitic Granite

Mylonitic granitic and granodioritic rocks forming narrow and long bodies elongated parallel to the NNW-NNE strike-slip faults are common in southwestern Goiás (e.g., Macacos and Serra do Tatu granites). The mylonitic foliation is vertical or sub-vertical. In the vicinities of Iporá, mylonitic granite is metaluminous or slightly peraluminous and shows geochemical characteristics similar to the high-K calc-alkaline series (Rodrigues, 1996), being distinctively more evolved geochemically than the orthogneiss discussed above. Rb/Sr isotope data for these rocks indicate isochron ages varying from c. 470 to 690 Ma and initial $^{87}Sr/^{86}Sr$ ratios between 0.7032 and 0.7062. Sm/Nd T_{DM} model ages are between 1.1 Ga and 950 Ma and initial ε_{Nd} values are positive (+ 0.4 to + 3.5; Table 2).

Volcano-Sedimentary Sequences

Several approximately N-S strips of volcano-sedimentary rocks are known in western/southwestern Goiás (Fig. 9). From W to E these are: Bom Jardim de Goiás, Arenópolis, Iporá/Amorinópolis and Jaupaci sequences. Farther to the E there is also the Anicuns-Itaberaí Sequence of unknown age. Recently obtained, unpublished Nd isotope data (as discussed in a previous section) indicate that at least part of it could belong to the arc association. The internal organization of the Bom Jardim de Goiás, Arenópolis, Iporá/Amorinópolis and Jaupaci sequences are better known.

The Bom Jardim de Goiás Sequence consists of greenschist facies metabasalt, meta-andesite and meta-rhyolite displaying preserved primary features such as pillows and pyroclastic textures (Seer, 1985). The upper unit of the supracrustal sequence is made mainly of detrital metasedimentary rocks (metaconglomerate, meta-arkose, siltstone and phyllite)

The Arenópolis Sequence (Pimentel and Fuck, 1986) is formed by two main units, separated from each other by a very narrow strip of strongly deformed banded gneiss displaying Archean T_{DM} model ages (the Ribeirão Gneiss; Pimentel, 1992). The Córrego do Santo Antônio Unit occupies the western part of the sequence and is made mainly of metapelite with garnet, staurolite, kyanite and sillimanite, marble, calc-silicate rocks, metachert, gondite and mafic-ultramafic bodies. This was interpreted as an accretionary prism and the mafic-ultramafic complexes as fragments of ophiolites (Pimentel and Fuck, 1986). The Córrego da Onça Unit, E of the Ribeirão Gneiss, is dominated by calc-alkaline metavolcanic rocks including metabasalt, meta-andesite, meta-dacite and metarhyolite, with lesser amounts of low-K metatholeiite, metagreywacke, fine-grained quartzite and metachert. Geochemical data for these rocks (low K, K/Na, Rb and REE) suggest a very primitive nature for the original magma, which is confirmed by Sm/Nd isotope data which show $\varepsilon_{Nd(T)}$ between + 2.5 and + 2.8 (Pimentel and Fuck, 1992). U/Pb zircon dating of metarhyolite of this sequence indicates the age of 929 ± 8 Ma for the cristallyzation of the protolith. Titanite from the same sample indicates the recristallyzation age of c. 594 Ma.

The volcano-sedimentary sequences in the Iporá area (Iporá, Jaupaci and Amorinópolis sequences, Figure 9) are apparently poor in intermediate volcanic rocks, comprising bimodal suites. They consist of metabasalt with relict prophyritic textures and vesicles, and meta-rhyolite, normally showing the pyroclastic nature of the protolith. Metasedimentary rocks are almost absent in the Jaupaci Sequence (Amaro, 1989) but are abundant in the Iporá and Amorinópolis sequences, being represented mainly by micaschist and metaconglomerate.

U/Pb zircon data for metarhyolite of the Jaupaci and Iporá sequences indicate crystallization ages of 764 ± 14 Ma and 636 ± 6 Ma, distinctively younger than the Arenópolis volcanics (Pimentel *et al.*, 1991a). Rb/Sr isochron age for mylonitic metarhyolites in the Fazenda Nova and Jaupaci areas indicate ages of 600 ± 31 Ma and 594 ± 37 Ma with initial Sr isotope ratios of c. 0.7036 and 0.7052, respectively. These are interpreted as recrystallization ages (Pimentel and Fuck, 1994).

The main structural feature observed in the Arenópolis section of the Goiás Magmatic Arc is a sub-vertical NNW to NNE mylonitic foliation, which obliterates previous deformational structures. Kinematic indicators and stretching lineation associated with this foliation, normally indicate dextral strike slip movements (Amaro, 1989). The geochronological data above suggest that these important strike slip shear zones formed during the Brasiliano event at c. 600 ± 30 Ma.

Late- to Post-Orogenic Granitic Magmatism and Associated Mafic and Ultramafic bodies

The last tectono-metamorphic event in the Brasília Belt happened between 630 - 590 Ma. This was responsible for the development of: (i) W-NW dipping foliation showing kynematic indicators revealing tectonic transport to the E (e.g., in the Mara Rosa Arc area), and (ii) wide and long subvertical, dominantly NE-trending mylonite zones, such as the Transbrasiliano Lineament (Fig. 1). During the waning stages or immediately after this last deformation, a number of small mafic-ultramafic layered complexes, gabbro-dioritic intrusions, and large granite plutons were emplaced into the western part of the Brasília Belt.

The gabbro-dioritic intrusions and mafic-ultramafic layered complexes (e.g., Lajeado intrusion and Americano do Brasil Complex, in western Goiás) are either only slightly deformed or completely free of any pervasive deformational fabric. The gabbro-diorite bodies commonly display magma mixing features with gabbro globules enclosed in dioritic rocks. These rocks display typical arc geochemical signatures. The Americano do Brasil Complex is a small layered intrusion differentiated from peridotite to gabbro and the original tholeiitic parental magma also had arc characteristics (Nilson, 1981). Gabbro and diorite also occur as small bodies inside the larger granite intrusions. Some of these gabbro-dioritic intrusions have been dated by the Rb/Sr isochron method (Table 2), giving poorly constrained ages of *c.* 630 - 610 Ma, with initial $^{87}Sr/^{86}Sr$ ratios of *c.* 0.703 - 0.704, suggesting little contamination with much older sialic crust. The Americano do Brasil Complex was recently dated by the Sm/Nd method at 612 ± 66 Ma, with an initial ε_{Nd} value of + 3.1 also indicating little or no contamination with ancient crust (Nilson *et al.*, 1997).

The granite intrusions form large bodies of K-rich calc-alkaline granite (*e.g.*, Serra Negra, Serra do Iran, Caiapó, Iporá, Sanclerlândia and Serra do Impertinente granites, shown in Fig. 9). Most are completely undeformed and made exclusively of equigranular biotite granite, although some have hornblende-bearing and porphyritic facies (*e.g.*, the Caiapó Granite; Pimentel *et al.*, 1996). K-rich, high crustal level, microgranite dykes also occur in small swarms in some localities in the Arenópolis area. In a broad regional scale, this igneous event is typically bimodal and, except for some localities where magma mixing between mafic and felsic end members took place, intermediate rocks as such are not found.

In the Arenópolis region, Rb/Sr isotope data suggest that the post-Brasiliano granite magmatism took place in two distinct episodes (Table 2): (i) an older event between *c.* 590 and 560 Ma, and (ii) a young event dated between *c.* 508 and 485 Ma (Pimentel *et al.*, 1996). Both groups are predominantly metaluminous, have $K_2O/Na_2O > 1$ and display general characteristics that are transitional between the high-K calc-alkaline and the shoshonitic series. The younger group has a more alkalic character, being similar to A-type granite (Serra Negra and Iporá granites, late facies of the Serra do Impertinente intrusion), whereas the older group has characteristics similar to the Caledonian I-type granite (Caiapó, Serra do Iran, and Israelândia granites, and early facies of the Serra do Impertinente Complex). This relationship is clearly displayed by the Serra do Impertinente intrusion. The early (575 Ma) porphyritic facies of this intrusion has Caledonian I-type characteristics with lower Ga, Zr+Nb+Ce+Yb, and higher La/Yb ratios, whereas the late (*c.* 485 Ma) equigranular facies has a typical A-type signature. Both groups have Sr and Nd isotope characteristics which are compatible with remelting of the Neoproterozoic arc granitoid and metavolcanic rocks (initial Sr ratios between 0.703 - 0.710 and $\varepsilon_{Nd(T)}$ between - 4.6 and + 3.0). Involvement of important amounts of much older continental crust can be ruled out.

The Mara Rosa Arc

In northern Goiás and southern Tocantins, the Goiás Magmatic Arc consists largely of tonalitic/dioritic orthogneiss, which underlie large areas in between narrow N20°-30°E volcano-sedimentary sequences. The rock units comprising the Goiás Magmatic Arc in this region are better known in the area between Porangatu, in the N, and Chapada, in the S (Fig. 10). The extension of these juvenile terranes towards the N is still unkown due to the lack of geochronological data.

The Supracrustal Belts

In the region between Chapada and Porangatu, regional mapping projects have identified a number of individual volcano-sedimentary sequences (Machado *et al.*, 1981; Ribeiro Filho, 1981; Lacerda, 1986; Arantes *et al.*, 1991; Kuyumjian, 1989). In the Mara Rosa area, these supracrustals form three individual NNE belts, known as the eastern, central and western belts, separated from each other by metatonalite/metadiorite (Fig. 10). According to Arantes *et al.* (1991) these belts are made of metabasalt, intermediate and felsic metatuff, fine-grained greywacke, garnet micaschist, chert, iron formation units, quartzite and ultramafic rocks, metamorphosed in the greenschist to amphibolite facies. Within the supracrustal belts, small elongated bodies of mylonitic granite have also been recognized (Palermo, 1996), some of which were previously misinterpreted as metavolcanic rocks. Small gold deposits are associated with these supracrustal belts (the Posse and Zacarias deposits; Arantes *et al.*, 1991).

Amphibolite of the Mara Rosa Sequence is either tholeiitic, rich in Mg, Ni and Cr and similar to boninites, or calc-alkaline. According to Palermo (1996) the former could represent fragments of the oceanic crust and the latter are related to the arc magmatism. In the Chapada area, the garnet and epidote amphibolite is chemically similar to modern arc tholeiite and were interpreted as being originated in a back-arc setting (Kuyumjian, 1994).

Geochronological and isotope data for the supracrustal belts are shown on Table 2. U/Pb zircon data for a felsite in the Posse gold mine yielded the age of 862 ± 8 Ma. This was previously interpreted as a metavolcanic rock, but the work of Palermo (1996) demonstrated that the protolith was a small granitic pluton that has been extremely deformed. Titanite from the same sample yielded the age of 632 ± 4 Ma for the metamorphism. Rb/Sr isochrons for metavolcanic and metasedimentary rocks in the Chapada and Mara Rosa areas give ages in the interval between 524 and 603 Ma with low initial $^{87}Sr/^{86}Sr$ ratios (0.7035 to 0.7045; Richardson *et al.*, 1988; Viana 1995). Sm/Nd T_{DM} values are between 1.1 Ga and 900 Ma (Pimentel *et al.*, 1997).

Detrital metasedimentary rocks represented by feldspathic garnet micaschist and fine-grained biotite gneiss are abundant in the supracrustal belts, especiallly in the western belt. Sm/Nd isotopes for these metasedimentary rocks indicate T_{DM} values dominantly in the range 1.2 Ga - 900 Ma, with only one analysis yielding a significantly older model age of 1.6 Ga. This indicates that they are the products of erosion of the arc rocks, with little contribution from older sources. The depositon of the original sediments must have taken place far from any old continental source area and probably happened in an intraoceanic setting. Sm/Nd garnet whole-rock isochrons for these metasedimentary rocks indicate ages of *c.* 765, 733, 610, and, 604 Ma. These

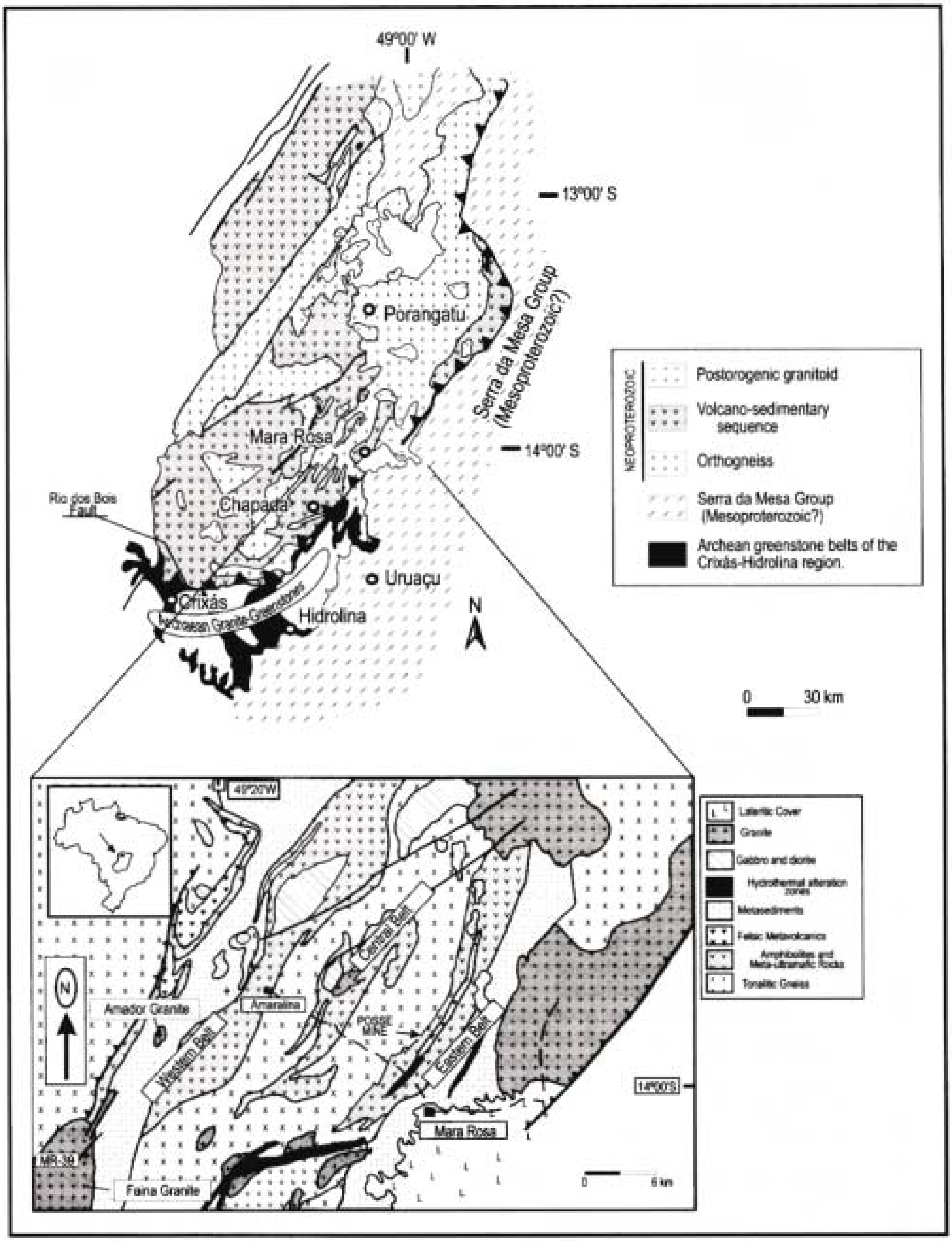


FIGURE 10 - Geological map of the Goiás Magmatic Arc in the Mara Rosa region, northern Goiás (after Arantes et al., 1991)

were interpreted as resulting from two metamorphic episodes: an early one at *c.* 760 Ma and a later metamorphism with typical Brasiliano ages (Junges 1998; Pimentel and Junges, 1997).

Orthogneiss

In the Mara Rosa area, these are medium to coarse grained grey coloured metaplutonic rocks of dioritic to tonalitic composition. They locally show well-preserved plutonic textures such as enclaves and porphyritic textures. They are geochemically very primitive with SiO_2 content less than 60%, display calcic to calc-alkaline character with low Rb, Nb, Y, Zr and REE, being similar to M-type granitoid of immature island arcs (Viana *et al.*, 1995). Tonalite bodies exposed in the vicinity of Chapada, about 40 km to the southwest of Mara Rosa are very similar petrographically and geochemically to the Mara Rosa rocks (Kuyumjian, 1989).

The stratigraphic relationships between this orthogneiss and the supracrustal belts are not clear due to intense deformation along the contacts between them. Arantes *et al.* (1991) however, describe the presence of metavolcanic enclaves in the tonalite.

Geochronological and isotope data for these rocks are in Table 2. U/Pb zircon data indicate crystallization of the protolith at 856 ± 13 Ma. Rb/Sr and K/Ar data indicate younger ages (< 600 Ma) reflcting the closure of these systems after the last metamorphism. Nd isotope data, similarly with the orthogneiss from the Arenópolis area, indicate the very primitive nature of the original magma, with T_{DM} values of *c.* 1.0 Ga - 900 Ma and $\varepsilon_{Nd(T)}$ of + 4.6.

Late- to Post-Orogenic Granite and Gabbro-Dioritic Intrusions

Similarly to the Arenópolis Arc, the last deformational event affecting the Mara Rosa rocks at *c.* 600 Ma was immediately followed by the intrusion of several granitic intrusions (*e.g.*, the Faina, Angelim, Estrela and Amador granites, Figure 10) as wells as gabbro-dioritic bodies. The granite bodies include mainly biotite granite and two-mica leucogranite, with local granodioritic facies (Viana, 1995). The mafic intrusions consist of diorite and, to a lesser extent, gabbro, and very commonly display magma mixing structures.

The granite *s.s.* is, in general, more peraluminous than the post-orogenic intrusions in the Arenópolis area (Junges, 1998) indicating larger contribution of metasedimentary material in the origin of the parental magma. Their Sm/Nd isotope composition indicates that they are the products of re-melting of metasedimentary rocks of the Mara Rosa Arc itself.

As in the Arenópolis area, therefore, the Precambrian geological evolution of the Mara Rosa Arc ended with an important bimodal magmatic event. In both cases this is interpreted as associated with final uplift and collapse of the Brasiliano Orogen.

The bimodal nature of this magmatism suggests that the input of heat required to promote large scale melting of the crust was most probably provided by the emplacement and/or underplating of mafic magma into the continental crust. This abundant post-Brasiliano granitic magmatism is spatially confined to the regions underlain by the juvenile arc rocks. This may reflect that these were preferable sites of uplift and melting of the mantle and continental crust or, alternatively, that the juvenile crust behaved as a more "fertile" material for granite production. Field and geochronological evidence suggest that the emplacement of these bodies occurred under extension, which accompanied rapid uplift and unroofing of the region. Amphibolite facies metavolcanic rocks with U/Pb sphene ages of *c.* 590 Ma are intruded by *c.* 560 Ma sub-volcanic microgranite dykes. Therefore, uplift rate of 0.4 to 0.6 mm/year is reported for the Arenópolis region in the period between 590 - 560 Ma, in the same order of magnitude, therefore, as rates estimated for modern mountain belts.

Geological Evolution

Field, isotope and geochronological data gathered over the last ten years for rock units in the central and eastern parts of the Tocantins Province have helped to clarify many aspects of the geological evolution of the province, more especifically of the Brasília Belt. Although many questions still remain unclear, the new data have demonstrated that large areas which were previously interpreted as sialic basement to the supracrustal sequences of the Brasília Belt are, in fact, underlain by Neoproterozoic granitoid plutons and metavolcanic/metasedimentary sequences and, therefore, ensimatic evolution models are more appropriate.

Regional Sm/Nd isotope investigations have indicated that most of the sialic basement to the Brasília Belt is Paleoproterozoic. It is now apparent that the only truly continuous and extensive Archean area in the central Tocantins Province corresponds to the granite-greenstone terranes of the Crixás-Goiás area. Similar units are not found anywhere else to the E, where the basement of the belt is exposed. Therefore, it is suggested that this Archean block is allocthonous in respect to the sialic basement of the belt, and it might have been accreted to the eastern limb of the São Francisco Continent during the Neoproterozoic.

The main stages of the tectonic evolution of the Brasília Belt, from an early rift stage at *c.* 1.77 - 1.60 Ga, through arc magmatism, peraluminous syn-collisional intrusions, to a post-orogenic bimodal event, are registered by the granitoid suites recently investigated in greater detail (Pimentel *et al.*, 1999a, b). The 1.77 - 1.6 Ga rift event was important regionally, well documented in the northeastern part of the province, but the subsequent history, between 1.6 and 1.2 Ga is still obscure. Most of the Neoproterozoic magmatism documented in the Brasília Belt is related to convergence and closure of an oceanic basin or to post-collisional uplift.

One of the most controversial aspects of the geological evolution of the Tocantins Province relies on the significance and age of the three large mafic-ultramafic complexes and associated volcano-sedimentary sequences, and more specifically, on the significance of the *c.* 1.3 Ga ages commonly found for mafic and felsic rocks of these units. Although still insufficient, the geochronological data now available describe a pattern that seems to apply to the three complexes. First, high-grade metamorphism happened between *c.* 790 and 770 Ma. Second, the eastern, granulite-facies layered series, commonly display Paleoproterozoic

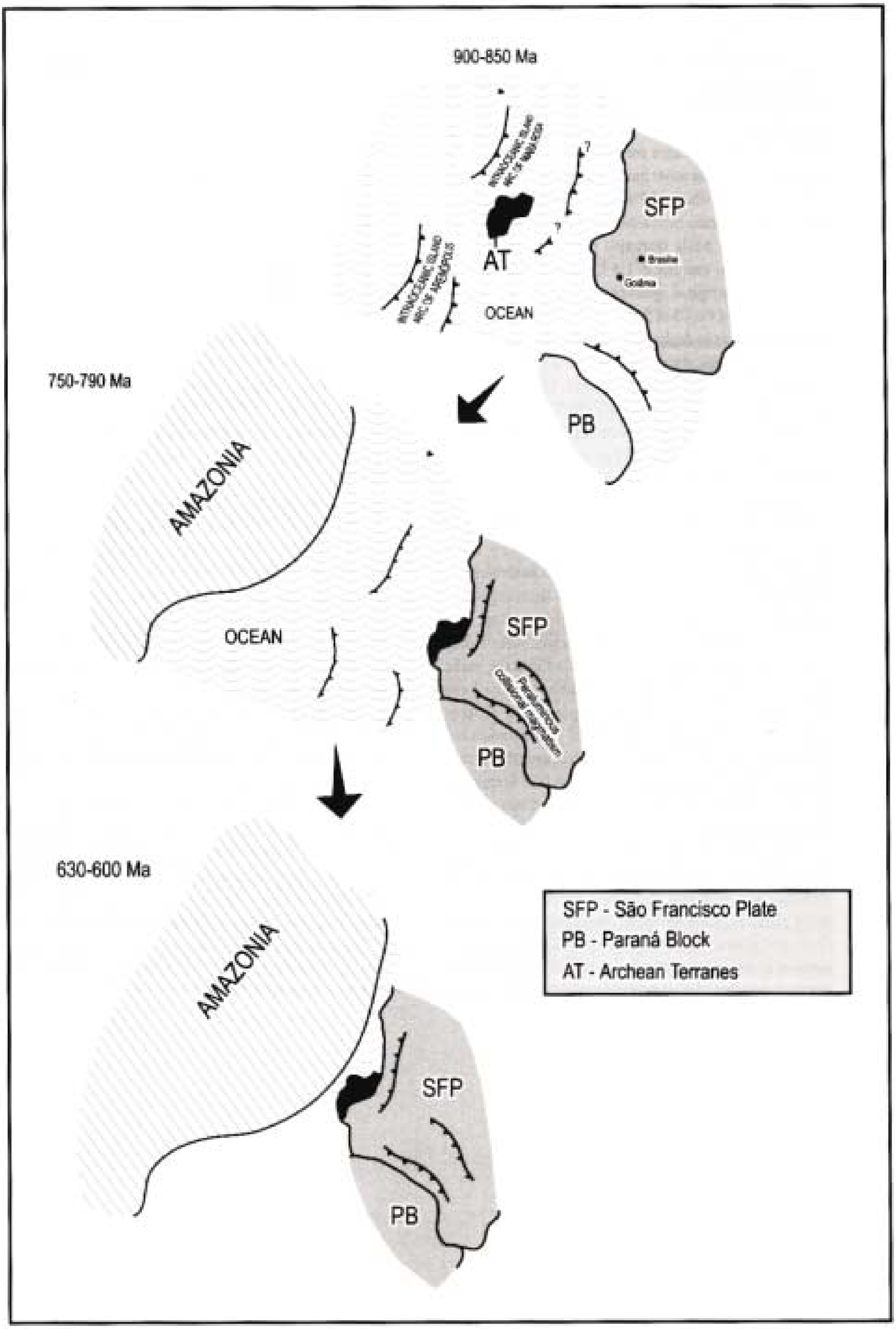


FIGURE 11 - Reconstruction model for the Neoproterozoic evolution of the Tocantins Province.

primary ages, whereas, rocks belonging to the western amphibolite-facies gabbro-anorthosites and volcanic sequences normally show ages around 1.3 Ga. These latter have been interpreted by some authors as metamorphic ages and were associated with the so-called Uruaçuano Orogenic Cycle (Fuck *et al.*, 1989; Correia *et al.*, 1997a, b), although similar metamorphic ages are unknown in the rest of the Brasília Belt. On the other hand, the studies of Suita *et al.* (1994); Ferreira Filho and Pimentel (1999), on the Barro Alto and Niquelândia intrusions, respectively, were carried out on selected basic intrusive rocks preserved from metamorphism, and the *c.* 1.3 Ga ages observed were interpreted as original igneous ages. Initial Nd isotope characteristics of the US of the Niquelândia Complex is very primitive, similar to depleted oceanic basaltic rocks.

Therefore, the data could be also interpreted as indicating rift-related magmatism with the formation of oceanic crust at 1.3 Ga. The oceanic character of the basalt comprising the volcano-sedimentary sequences of Juscelândia, Indaianópolis and Palmeirópolis suggest that they represent volcanic counterparts of the gabbro of the upper series of the Barro Alto and Niquelândia complexes. These rocks might, therefore, represent large fragments of Mesoproterozoic oceanic crust preserved within the Neoproterozoic Brasília Belt. The bimodal nature of this 1.3 Ga magmatic event could indicate that these sequences correspond to the initial stages of rifting starting in a continental setting and evolving to an oceanic basin. In this model, the small granite bodies intruded into the mafic granulites in Barro Alto and Niquelândia that have been repeatedly dated at *c.* 1.3 Ga, might represent the plutonic equivalents of the felsic volcanics in the volcano-sedimentary sequences. The isotope data available at this stage, however, do not allow a definitive conclusion regarding this controversial aspect of the geological history of the Tocantins Province.

The *c.* 900 - 850 Ma primitive meta-tonalite of Arenópolis and Mara Rosa and the *c.* 920 Ma Arenópolis rhyolitic rocks represent the earliest arc products formed during convergence between the Amazonian and São Francisco plates. The long history of arc magmatism, between *c.* 900 and 630 Ma, suggests that during most of this period of time, the western margin of the São Francisco Continent faced a large oceanic basin where intraoceanic island arcs formed and were accreted to each other or to the continental margin (Fig. 11). At some stage during the Neoproterozoic evolution of the belt, the Archean block of Crixás-Goiás was accreted to the margin of the belt. This is compatible with a peripheral position of the Brasília Belt in relation to the Rodinia Supercontinent. Final closure between the Amazonian and São Francisco continents ocurred at *c.* 600 Ma, with the development of extensive N-NE trending shear zones as well as E-verging thrust faults.

The *c.* 790 Ma old collisional magmatism associated with the accretionary wedge of the Brasilia Belt is described only in the southern part of the belt, S of the City of Brasília. This points to distinct tectonic evolution models for the northern and southern part of the belt. Although there is no supporting tectonic evidence at the present level of knowledge, this "early Brasiliano" collisional event is interpreted as representing the collision between the southwestern part of the São Francisco Continent and the Paraná Block, which is now covered by the Paleozoic and Mesozoic rocks of the Paraná Basin.

The typically Brasiliano (630 Ma) metamorphic ages recorded in several areas of the province, including the granulite of the Anápolis-Itauçu Complex are interpreted as resulting from the collisional event representing final closure of the ocean. The Brasiliano granulite of the Anápolis-Itauçu Complex probably indicate the site where crustal thickening reached its maximum during the late Neoproterozoic.

Following the closure of the ocean and crustal thickening, some areas of the orogen were invaded by mafic magmas which promoted dehydration melting of the crustal rocks, generating large volumes of post-orogenic K-rich calc-alkaline granite. Their Nd-Sr isotope composition indicates that the original magmas were mostly derived from juvenile Neoproterozoic arc rocks. This early phase of the *c.* 590 - 560 Ma extension-related bimodal magmatism is roughly coeval with the deposition of the Paraguai Belt rocks and its correlatives of the Tucavaca Belt in Bolívia (Fig. 1). The latter is interpreted in recent reconstruction models as being formed during rifting of Laurentia from the western margin of Gondwana.

Table 1 - Rb-Sr geochronological data for gneissic units of the Archean Crixás- Goiás Block.

Rock Unit	Rb/Sr age (Ga)	References
Tocambira Tonalite	2.965 ± 0.065 2.924 ± 0.150	1,2,3
Tonalitic gneiss, Itapuranga Complex	2.840 ± 0.025 2.651 ± 0.027	4
Granodiorite, Hidrolina Complex	2.653 ± 0.040	2
Tonalitic gneiss, Uvá Complex	2.564 ± 0.140	5
Granodiorite, Anta Complex	2.475 ± 0.020 2.530 ± 0.098	1, 3

1. Tassinari and Montalvão (1980), 2. Montalvão (1986), 3. Vargas (1992), 4. Tassinari et al. (1981), 5. Pimentel et al. (1985)

Table 2 - Geochronological data for rock units of the Goiás Magmatic Arc.

Rock Unit	Age (Ma)	(87Sr/86Sr)i	TDM (Ga)	∈Nd(T)	Refer.
Orthogneisses					
Arenópolis gneiss	899±7 a 818±57b 637 c	0.7042	1.2-1.0	+1.9/+3.2	1,2
Matrinxã gneiss	c. 895b	0.7026	0.9	+ 6.0	2
Sanclerlândia gneiss	c. 940b	0.7025	1.0-0.9	+4.0/+6.0	2
Firminópolis gneiss	628±65d		1.2-1.1	-1.7	3
Choupana granite-gneiss	855±98d		1.1-0.9	+4.0	Unp
Mara Rosa tonalite gneiss	856±13a 600±136b 517±16h 533±16h	0.7032	1.0-0.9	+4.6	8,7
Porangatu granite-gneiss	589±61b	0.705			12
Mylonitic granites					
Santa Fé granite-gneiss	467±10b	0.705			4
Serra do Tatu granite	692±110b	0.7062			5
Porphyritic granite - Jaupaci	643±19b	0.7032	0.95	+3.2/+3.5	2
Mylonitic granite west of the Iporá sequence	673±75b	0.7048	1.1-1.0	+0.4/+1.9	6
Volcano-Sedimentary Sequences					
Metarhyolite–Arenópolis Sequence	929±8a 933±60b 594c	0.7035	1.2-0.9	+2.5/+5.8	1
Amphibolite – Arenópolis Sequence	859±43e				7
Mylonitic meta-rhyolite – Fazenda Nova	600±31b	0.7036	1.1-0.9	+0.2/+2.4	2
Meta-rhyolite – Jaupaci Sequence	764±14a 594±37b	0.7052	1.0-0.9	+3.8/+4.7	1,2
Meta-rhyolite – Iporá Sequence	636±6a 597±5a		1.1-1.0	+0.6/+5.0	6
Pontalina volcanics	830±99d		1.1-1.0	+3.3	Unp
Amphibolite – Mara Rosa	730±37e				7
Felsic metavolcanic - Chapada	561±9b	0.7041			10
Biotite schist - Chapada	524±2f 532±1f				10
Garnet biotite schist – Mara Rosa	c. 560b	0.7045	1.0-0.9		8
Mylonite – Mara Rosa central belt	603±135b	0.7035			9
Garnet micaschist - Mara Rosa western belt	733±75g 765±75g 604±66g 610±52g		1.0-0.9	-3.0/+2.0	11
Felsite – Posse Mine, Mara Rosa	862±8a 632±4c		1.1-1.0	+3.7	8

a – Zircon U/Pb age; b – Whole-rock Rb/Sr isochron; c - Sphene U/Pb age; d – Whole-rock Sm/Nd age; e – K/Ar amphibole date; f – Rb/Sr mineral isochron; g – Sm/Nd garnet-whole rock date; h - K/Ar biotite age.
References: 1 – Pimentel et al., 1991; 2 – Pimentel and Fuck, 1994; 3 – Pimentel and Gioia, 1997; 4 – Barbour et al., 1979; 5 – Pimentel et al., 1985; 6 – Rodrigues, 1996; 7 – Hasui and Almeida, 1970; 8 – Pimentel et al., 1997; 9 – Viana et al., 1995; 10 – Richardson et al., 1988; 11 – Pimentel et al., 1998; 12 - Tassinari et al., 1981

REFERENCES

Almeida, F.F.M., Hasui, Y., Neves, B.B.B. and Fuck, R.A. (1981). Brazilian structural provinces: an introduction. *Earth Sciences Review, 17, 1-29.*

Amaro, V.E. (1989). *Geologia e petrologia da seqüência metavulcânica de Jaupaci-GO e lineamentos associados.* MSc Dissertation. UnB, Brasília, 230 p.

Arantes, D., Osborne, G.A., Buck, P.S. and Porto, C.G. (1991). The Mara Rosa volcano-sedimentary sequence and associated gold mineralization. In: *Brazil Gold '91*, ed. Ladeira, E.A., pp. 221-229. Balkema, Rotterdam.

Araújo, S.M. (1986). *Petrologia e mineralizações sulfetadas da seqüência vulcano-sedimentar de Palmeirópolis-Goiás.* MSc Dissertation, UnB, Brasília, 196 p.

Araújo, S.M. (1996). *Geochemical and isotopic composition of alteration zones in highly metamorphosed volcanogenic massive sulfide deposits and their potential application to mineral exploration.* PhD Thesis, Univ. of Toronto, Toronto, 201 p.

Araújo, S.M., Fawcett, J.J. and Scott, S.D. (1995). Metamorphism of hydrothermally altered rocks in a volcanogenic massive sulfide deposit: the Palmeirópolis, Brazil example. *Revista Brasileira de Geociências*, 25, 173-180.

Araújo, S.M., Longstaffe, F.J. and Scott, S.D. (1996b). Effects of the high-grade metamorphism on the oxygen isotope composition of the alteration zones associated with the Palmeirópolis volcanogenic massive sulfide deposit, Tocantins State, Brazil. *Anais, XXXIX Congresso Brasileiro de Geologia, Salvador, SBG*, 7, 339-342.

Araújo, S.M. Gorton, M.P. and Scott, S.D. (1996a). Os anfibolitos enriquecidos em elementos traço da seqüência vulcano-sedimentar de Palmeirópolis, Tocantins: uma ferramenta na prospecção de metais básico na região. *Anais, XXXIX Congresso Brasileiro de Geologia, Salvador*, 3, 178-180.

Araújo, S.M., Longstaffe, F.J. and Scott, S.D. (1993). Oxygen isotopes and hydrothermal alteration associated with Geco, Canada, and Palmeirópolis, Brazil volcanogenic massive sulfide deposits: evidence for isotopic rehomogeneization during high-grade metamorphism. *Program with Abstracts, GAC MAC Meeting*, p. A3.

Araújo, S.M. and Nilson, A.A. (1987). Caracterização petroquímica petrotectônica dos anfibolitos da seqüência vulcano-sedimentar de Palmeirópolis-Goiás. *Anais, I Congresso Brasileiro de Geoquímica*, 335-348.

Araújo, S.M. and Nilson, A.A. (1988). Depósito de zinco, cobre e chumbo de Palmeirópolis, Goiás. In: Principais Depósitos Minerais do Brasil, ed. Schobbenhaus, C. and Coelho, C.E.S., *MME/DNPM/CVRD*, 3, 171-180.

Araújo, S.M. and Scott, D.S. (1991). *The Palmeirópolis massive sulfide deposit, Brazil.* Program with Abstracts, GAC-MAC meeting.

Araújo, S.M., Scott, D.S. and Longstaffe, F.J. (1996c). Oxygen isotope composition of alteration zones of highly metamorphosed volcanogenic massive sulfide deposits: Geco, Canada and Palmeirópolis, Brazil. *Economic Geology*, 91, 697-712

Araújo, V.A. and Alves, A.C. (1979). *Projeto Cana Brava-Porto Real. Relatório Final.* CPRM/DNPM Goiânia.

Arndt, N.T., Teixeira, N.A. and White, W.M. (1989). Bizarre geochemistry of komatiites from the Crixás greenstone belt, Brazil. *Contributions to Mineralogy and Petrology*, 74, 549-564.

Barbosa, O., Baptista, M.B., Dyer, R.C., Braun, O.P.G. and Cotta, J.C. (1969). *Geologia e inventário dos recursos minerais do Projeto Brasília.* DNPM/PROSPEC, Rio de Janeiro, 225 p.

Barbosa, P.A.R. (1987). Geologia e recursos minerais das seqüências supracrustais a leste de Mossâmedes-GO. *MSc Dissertation, UnB, Brasília, 226 p.*

Barbour, A.P., Girardi, V.A.V., Kawashita, K. and Souza, A.M.S. (1979). Geologia do complexo máfico-ultramáfico alcalino de Santa Fé, Goiás. *Boletim do IG*, 10, 11-18.

Botelho, N.F. (1992). *Les ensembles subalcalins a peralumineux mineralisés en Sn et In de la Sous Province Paranã, État de Goiás, Brésil.* Paris, École Nationale Superieure des Mines de Saint Etienne, Tèse de Doctorat.

Botelho, N.F., Bilal, E., Moutte, J. and Fonteilles, M. (1993). Precambrian A-type tin-bearing granites in the Goiás tin province, central Brazil: a review. *Resumos Expandidos. I Workshop sobre Magmatismo Granítico e Mineralizações Associadas, Rio de Janeiro.* Academia Brasileira de Ciências, 5-8.

Brito Neves, B.B. and Cordani, U.G. (1991). Tectonic evolution of South America during the late Proterozoic. *Precambrian Research*, 53, 23-40.

Brod, J.A. (1988). *Geologia das rochas vulcano-sedimentares da região de Indaianópolis, GO e lineamentos adjacentes.* MSc Dissertation, UnB, Brasília, 267 p.

Brod, J.A. and Jost, H. (1991). Características estruturais, litológicas e magmáticas da zona de cisalhamento dúctil do Rio Traíras, Bloco do Complexo de Niquelândia, Goiás. *Revista Brasileira de Geociências*, 21, 205-217.

Brod, J.A. and Jost, H. (1994). Revisão estratigráfica da região de Indaianópolis, Anais, *IV Simpósio de Geologia do Centro-Oeste, Brasília*, SBG, 138-141.

Bruni, M.A.L., Schobbenhaus, C., Cordani, U.G., Campos, D.A., Campos, D.R.B., Rocha, A.J.D., Bruni, E.C., Silva, H.P., Almeida, J.T., Batista, J.J., Sena, R.M., Sá, V.R., Derze, G.R. and Lagarde, A. (1976). *Carta Geológica do Brasil ao Milionésimo, Folha Brasília (SD.23).*, MME/DNPM, Brasília, 163 p.

Castro, J.H.G. and Magalhães, L.F. (1984). Revisão estratigráfica do Grupo Pilar de Goiás na Faixa Crixás. *Anais, XXXIII Congresso Brasileiro de Geologia, Rio de Janeiro*, p. 2563-2576.

Correia, C.T. (1994). *Petrologia do Complexo Máfico-Ultramáfico de Cana Brava, Goiás.* Doctorate Thesis, USP, São Paulo, 151 p.

Correia, C.T., Girardi, J.A.V., Lambert, D.D., Kinny, P.D. and Reeves, S.J. (1996). 2 Ga U-Pb SHRIMP-II and Re-Os ages for the Niquelândia basic-ultrabasic layered intrusion, central Goiás, Brasil. *Anais, XXXIX Congresso Brasileiro de Geologia, Slavador, SBG*, 187-189.

Correia, C.T., Girardi, V.A.V., Tassinari, C.C.G. and Jost, H. (1997a). Rb-Sr and Sm-Nd geochronology of the Cana Brava layered mafic-ultramafic intrusion, Brazil, and considerations regarding its tectonic evolution. *Revista Brasileira de Geociências*, 27(2), 163-168.

Correia, C.T., Tassinari, C.C.G., Lambert, D.D., Kinny, P. and Girardi, V.A.V. (1997b). U-Pb (SHRIMP), Sm-Nd and Re-Os systematics of the Cana Brava, Niquelândia and Barro Alto layered intrusions in central Brazil, and constraints on the tectonic evolution. *Extended Abstracts. I South-American Symposium on Isotope Geology – SSAGI, Campos do Jordão*, IG-USP, 88-89.

Correia Filho, F.C.L. and Sá, A.M. (1980). *Projeto Natividade. Relatório Final.* Goiânia, DNPM/CPRM, 6V.

Costa, J.B.S. (1984). *Aspectos litoestruturais e evolução crustal da região centro-norte de Goiás.* Belém, UFPA, Doctorate Thesis, 210 p.

Costa, L.A.M., Portella, A.C.P., Nilson, A.A., Vale, C.R.O., Marchetto, C.L.M., Santos, E.L., Meneguesso, G., Inda, H.A.V., Sterna, L.H., Marchetto, M., Baptista, M.B., Fratin, O., Mosmann, O., Oliveira, T.F. and Silva, W.G. (1976). *Projeto Leste Tocantins, Oeste do São Francisco. Relatório Final.* Rio de Janeiro, DNPM/CPRM/PROSPEC, 200 p.

Cruz, E.L.C.C. (1993). *Geologia e mineralizações auríferas do Terreno Granitóide-Greenstone de Almas-Dianópolis, Tocantins.* UnB, Brasília, MSc Dissertation, 152 p.

Cruz, E.L.C.C. and Kuyumjian, R.M. (1996). Chemical characterization of metabasalts and granitoids from the Almas-Dianópolis granite-greenstone terrane, central Brazil. *Extended Abstracts, Symposium on Archaean of the South American Platform. Brasília*, SBG, 53-54.

Cruz, E.L.C.C. and Kuyumjian, R.M. (1998). The geology and tectonic evolution of the Tocantins granite-greenstone terrane: Almas-

Dianópolis region, Tocantins State, central Brazil. *Revista Brasileira de Geociências*, 28(2),173-182.

Danni, J.C.M. (1988). Os greenstone belts da Provincia Tocantins no estado de Goiás, Brasil. *Revista Brasileira de Geociências*, 18, 381-390.

Danni, J.C.M., Dardenne, M.A. and Fuck, R.A. (1981). Geologia da Serra da Santa Rita e Seqüência Serra do Cantagalo. *Atas, I Simpósio de Geologia do Centro-Oeste, Goiânia*, SBG, 265-280.

Danni, J.C.M. and Fuck, R.A. (1981). Unidades tectono-estratigráficas do embasamento das faixas dobradas da margem ocidental do Cráton do São Francisco. *Anais, I Simpósio sobre o Cráton do São Francisco e suas Faixas Marginais. Salvador*, SME/CPM/SBG, 130-139.

Danni, J.C.M., Fuck, R.A., Kuyumjian, R.M., Leonardos, O.H. and Winge, M. (1984). O Complexo de Barro Alto na região de Ceres-Rubiataba, Goiás. *Revista Brasileira de Geociências*, 14(2), 128-136.

Danni, J.C.M., Fuck, R.A. and Leonardos, O.H. (1982). Archean and Lower Proterozoic units in central Brazil. *Geologisches Rundschau*, 71(1), 291-317.

Danni, J.C.M., Jost, H., Winge, M. and Andrade, G.F. (1986). Aspectos da evolução dos terrenos granito-greenstone: exemplo da região de Hidrolina. *Anais, XXXIV Congresso Brasileiro de Geologia, Goiânia*, SBG, 570-584.

Danni, J.C.M. and Kuyumjian, R.M. (1984). A origem dos anfibolitos da seqüência vulcano-sedimentar de Juscelândia, Goiás. *Anais, XXXIII Congresso Brasileiro de Geologia, Rio de Janeiro*, 9, 4126-4136.

Danni, J.C.M. and Leonardos, O.H. (1978). As seqüências granulítica, anortosítica-anfibolítica e ultramáfica da região de Niquelândia - Goiás. *Resumos, XXX Congresso Brasileiro de Geologia*, 45.

Danni, J.C.M. and Ribeiro C.C. (1978). Caracterização estratigráfica da seqüência vulcano-sedimentar de Pilar de Goiás e de Guarinos. *Anais, XXX Congresso Brasileiro de Geologia*, SBG, *2, 582-596.*

Faria, A., Marini, O.J., Winge, M., Kuyumjian, R.M. and Figueiredo, A.N. (1986). Geologia da região de São Domingos. *Anais, XXXIV Congresso Brasileiro de Geologia, Goiânia*, SBG, 2, 861-875.

Fernandes, P.E.C.A., Montes, M.L., Braz, E.R.C., Montes, A.S.L., Silva, L.L., Oliveira, F.L.L., Ghignone, J.I., Siga Jr., O. and Castro, H.E.F. (1982). *Geologia. Projeto RADAMBRASIL, Folha SD.23 Brasília.* Rio de Janeiro, MME (Levantamento de Recursos Naturais, 29), 25-204.

Ferreira Filho, C.F. (1994). *The Niquelândia mafic-ultramafic layered intrusion, north Goiás, Brazil: petrology, age and potential for PGE ore deposits*. PhD Thesis, Univ. of Toronto, Toronto, 270 p.

Ferreira Filho, C.F. (1998). Geology and petrology of the large layered intrusions of central Brazil, implications for PGE mineralization. *Platinum Symposium, Rustemburg, South Africa, Extended Abstracts*, VIII, 107-110.

Ferreira Filho, C.F., Araújo, S.M. and Cruz, H.P. (1999). Estruturas vulcânicas preservadas em granulitos da seqüência vulcano-sedimentar Juscelândia, GO. *Revista Brasileira de Geociências* (in press).

Ferreira Filho, C.F., Kamo, S., Fuck, R.A., Krogh, T.E. and Naldrett, A.J. (1994). Zircon and rutile geochronology of the Niquelândia layered mafic and ultramafic intrusion, Brazil: constraints for the htiming of magmatism and high grade metamorphism. *Precambrian Research*, 68, 241-255

Ferreira Filho, C.F., Moraes, R., Fawcett, J.J. and Naldrett, A.J. (1998b). Amphibolite to granulite progressive metamorphism in the Niquelândia Complex, central Brazil: regional tectonic implications. *Journal of South American Earth Sciences*, 11, 35-50.

Ferreira Filho, C.F. and Naldrett, A.J. (1993). The Niquelândia mafic-ultramafic complex revisited: tectonic setting and potential for PGE deposits. *Actas, I Brazilian PGE Meeting, Brasília*, 25-28.

Ferreira Filho, C.F. and Pimentel, M.M. (1999). Sm-Nd isotope systematics and REE-Hf-Ta-Th data of troctolites and their amphibolitized equivalents of the Niquelândia Complex Upper Layered Series, central Brazil: further constraints for the timing of magmatism and high grade metamorphism. *Journal of South American Earth Sciences* (in press).

Ferreira Filho, C.F., Naldrett, A.J. and Gorton, M.P. (1995). Petrology of the Mesoproterozoic Niquelândia layered mafic-ultramafic complex, central Brazil: evidence for a composite intrusion in an intracontinental rift system. *Actas, International Field Conference and Symposium, Duluth*, p. 45.

Ferreira Filho, C.F., Naldrett, A.J. and Gorton, M.P. (1998a). REE and pyroxene compositional variation across the Niquelândia layered intrusion, Brazil: petrological and metallogenetic implications. *Transactions of the Intitution of Mining and Metallurgy*, 107 (section B), 1-22.

Ferreira Filho, C.F., Nilson, A.A. and Naldrett, A.J. (1992). The Niquelândia mafic-ultramafic complex, Goiás, Brazil: a contribution to the ophiolite x stratiform controversy based on new geological and structural data. *Precambrian Research*, 59, 125-143.

Ferreira Filho, C.F., Silva, C.B. and Lopes, R.O. (1996). Variação críptica de olivinas em dunitos da Unidade Ultramáfica do Complexo de Niquelândia, GO. *Actas, Congresso Brasileiro de Geologia, Salvador*, 6, 179-182.

Figueiredo, J.A. Leão Neto, R. and Valente, C.R. (1981). Depósitos de sulfetos maciços de Zn, Cu e Pb da região de Palmeirópolis, GO. *Atas, I Simpósio de Geologia do Centro-Oeste, Goiânia*, 422-441.

Fischel, D.P., Pimentel, M.M. and Fuck, R.A. (1998). Idade do metamorfismo de alto grau no Complexo Anápolis-Itauçu, Goiás, determinada pelo método Sm-Nd. *Revista Brasileira de Geociências*, 28(4), 607-609.

Fortes, P.T.F.O. (1991). *Geologia do depósito aurífero Mina III, Crixás, Goiás*. MSc dissert., UnB, Brasília, 194 p.

Fortes, P.T.F.O. (1996). *Metalogênese dos depósitos auríferos Mina III, Mina Nova e Mina Inglesa, Greenstone Belt de Crixás, GO.* Doctorate thesis, UnB, Brasília, 176 p.

Fortes, P.T.F.O., Cheilletz, A., Giuliani, G. and Féraud, G. (1997). A Brasiliano age (500±5 Ma) for the Mina III gold deposit, Crixás greenstone belt, central Brazil. *International Geology Review*, 39, 449-460.

Fortes, P.T.F.O., Giuliani, G., Takaki, T., Pimentel, M.M. and Teixeira, W. (1995). Aspectos geoquímicos do depósito aurífero de Mina III, greentone belt de Crixás, Goiás. *Geochimica Brasiliensis*, 9(1), 13-31.

Freitas, M.E. and Kuyumjian, R.M. (1995). Variação química das micas durante o hidrotermalismo peraluminoso de Vianópolis, GO. *Boletim de Geociências do Centro-Oeste*, 18(1/2), 44-56.

Fuck, R.A., Brito Neves, B.B., Cordani, U.G. and Kawashita, K. (1989). Geocronologia Rb-Sr no Complexo Barro Alto, Goiás: evidência de metamorfismo de alto grau e colisão continental há 1300 Ma no Brasil Central. *Geochimica Brasiliensis*, 3(2), 125-140.

Fuck, R.A., Danni, J.C.M., Winge, M., Andrade, G.F., Barreira, C.F., Leonardos, O.H. and Kuyumjian, R.M. (1981). Geologia da região de Goianésia. *Atas, I Simpósio de Geologia do Centro-Oeste, Goiânia*, SBG-NCO, 447-467.

Fuck, R.A., Marini, O.J., Dardenne, M.A. and Figueiredo, A.N. (1988). Coberturas metassedimentares do Proterozóico médio: os grupos Araí e Paranoá na região de Niquelândia-Colinas, Goiás. *Revista Brasileira de Geociências*, 18(1), 54-62.

Fugi, M.Y. (1989). *REE geochemistry and Sm/Nd geochronology of the Cana Brava Complex, Brazil.* MSc Dissertation, Kobe University, 55 p.

Gioia, S.M.C. (1997). *Preparação da Metodologia Sm-Nd para Datação de Amostras Geológicas e sua Aplicação em Rochas das áreas de Firminópolis, Fazenda Nova e Americano do Brasil.* MSc Dissertation, UnB, 100 p.

Girardi, V.A.V., Kawashita, K., Basei and M.A.S., Cordani, U.G. (1978). Algumas considerações sobre a evolução geológica da região de Cana Brava, a partir de dados geocronológicos, Anais, *XXX*

Congresso Brasileiro de Geologia, 337-348.

Hasui, Y. and Almeida, F.F.M. (1970). Geocronologia do centro-oeste brasileiro. *Boletim da Sociedade Brasileira de Geologia*, 19(1), 5-26..

Hasui, Y., Tassinari, C.C.G., Siga Jr., O., Teixeira, W., Almeida, F.F.M. and Kawashita, K. (1980). Datações Rb-Sr e K-Ar no centro-norte do Brasil e seu significado geológico-geotectônico. *Anais, XXXI Congr. Bras. Geologia, 31. Balneário Camboriú*, SBG, Anais, 5, 2669-2676.

Jost, H., Carmelo, A.C. and Menezes, P.R. (1994a). Litologias, relações de contato e estruturas do Bloco do Caiamar, Crixás, Goiás. *Resumos Expandidos, IV Simpósio de Geologia do Centro-Oeste, Brasília*. SBG, 18-20.

Jost, H., Figueiredo, A.M.G. and Ferreira, A.V. (1996). Are all detrital metasedimentary rocks of the Crixás greenstone belt Archean and of the same provenance? A discussion based on REE geochemistry. *Extended Abstracts, I Symposium on Archaean of the South American Platform, Brasília*, 44-46.

Jost, H., Kuyumjian, R.M., Freitas, A.L.S., Costa, A.L.L., Nascimento, C.T.C., Vasconcelos, F.M., Galotti, L., Martins, M.C.A., Carvalho, M.N. and Condé, V.C. 1995. Geologia da porção norte do greenstone belt de Guarinos, GO. *Revista Brasileira de Geociências*, 25, 51-60.

Jost, H. and Oliveira, A.M. (1991). Stratigraphy of the greenstone belts, Crixás region, Goiás, central Brazil. *Journal of South American Earth Sciences*, 4, 201-214.

Jost, H., Resende, M., Kuyumjian, R.M., Queiroz, C.L., Osborne, G.A., Blum, M.L.B, Pires, A.C.B. and Moraes, R.A.V. (1999). Terrenos arqueanos de Goiás. In: *Geotectônica do Brasil*, ed. Hasui, Y. and Sena Costa, R. (in press).

Jost, H., Theodoro, S.M.C.H., Figueiredo, A.M.G. and Boaventura, G.R. (1996). Propriedades geoquímicas e proveniência de rochas metassedimentares detríticas arqueanas dos greenstone belts de Crixás e Guarinos, Goiás. *Revista Brasileira de Geociências*, 26, 151-166.

Jost, H., Vargas, M.C., Fuck, R.A., Kuyumjian, R.M. and Pires, A.C.B. (1994b). Relações de contato, litologias, geologia estrutural e geofísica do Bloco Arqueano do Moquém, Crixás, Goiás. *Resumos Expandidos, IV Simpósio de Geologia do Centro-Oeste, Brasília*, 15-17.

Junges, S.L. (1998). *Características geoquímicas e isotópicas dos metassedimentos da seqüência vulcano-sedimentar de Mara Rosa - GO*. MSc Dissertation, UnB, Brasília, 120 p.

Kuyumjian, R.M. (1981). *Geologia e mineralizações auríferas do greenstone belt da Faixa Crixás, GO*. MSC dissertation, UnB, Brasília, 69 p.

Kuyumjian, R.M. (1989). Geoquímica e significado do posicionamento geotectônico de rochas plutônicas da região de Chapada, Goiás, Brasil. *Anais, XXXIII Congresso Brasileiro de Geologia, Rio de Janeiro*, SBG-CPRM, 195-201.

Kuyumjian, R.M. (1994). Geologia da seqüência Mara Rosa na região de Chapada, Goiás. *Anais, IV Simpósio de Geologia do Centro-Oeste, Brasília*, SBG, 142-144.

Kuyumjian, R.M. and Danni, J.C.M. (1991). Geoquímica de anfibolitos da seqüência de Juscelândia: implicações tectônicas. *Revista Brasileira de Geociências*, 21, 218-223.

Kuyumjian, R.M. and Teixeira, N.A. (1982). Um novo tipo de estrutura em lavas ultramáficas: greenstone belt de Crixás, GO. *Revista Brasileira de Geociências*, 12, 572-577.

Lacerda, H. (1986). As mineralizações auríferas da região de Mara Rosa (GO). *Revista Brasileira de Geociências*, 16(3), 274-284.

Lacerda, H. and Lima Júnior, E.A. (1996). *Mapa geológico do greenstone belt de Crixás – faixas Guarinos e Pilar de Goiás*. MME/DNPM, Goiânia.

Lacerda Filho, J.V. Camargo, M.A. and Wildner, W. (1995). Granitóide sintectônico de Maratá. *Anais, V Simpósio de Geologia do Centro-Oeste, Cuiabá*, SBG, 66-67.

Lacerda Filho, J.V., Marques, V., Scislewiski, G, Jorge, L., Justo, E.C. and Oliveira, C.C. (1991). *Projeto Geologia da Região Centro-Oeste, Folha Caraíba*, CPRM. Goiânia, 145 p.

Lacerda Filho, J.V. and Oliveira, C.C. (1995). Geologia da região centro-sul de Goiás. *Boletim de Geociências do Centro-Oeste*, 18(1/2), 3-19.

Leão Neto, R. and Olivatti, O. (1983). *Projeto Palmeirópolis – Etapa Preliminar*. DNPM-CPRM, Goiânia, 29 p.

Lima, T.M. (1997). *Geologia, estratigrafia e petrologia da porção sul do Complexo máfico-ultramáfico de Cana Brava, Goiás*. MSc Dissertation, UnB, Brasília, 312 p.

Machado, E.C., Souza, C.J.M., Silva, M.A. and Berbert, C.O. (1981). *Projeto Porangatu – relatório final. Etapa II*, DNPM/CPRM, 1, 246 p.

Magalhães, L.F. (1991). *Cinturão de cisalhamento de empurrão Córrego Geral/Meia Pataca: geologia, deformação, alteração hidrotermal e mineralizações auríferas associadas (Crixás, Goiás)*. MSc dissertation, UnB, Brasília, 233 p.

Marini, O.J., Fuck, R.A., Dardenne, M.A., Danni, J.C.M. (1984). Província Tocantins. Setores Central e Sudeste. In: O Pré-Cambriano do Brasil, eds. by Almeida, F.F.M. and Hasui, Y., pp. 205-264. Edgard Blucher, São Paulo.

Marini, O.J., Liberal, G.S., Reis, L.T., Trindade, C.A.H. and Souza, S.L. (1978). Nova unidade litoestratigráfica do Pré-Cambriano do Estado de Goiás. *Congresso Brasileiro Geologia, 30. Recife, SBG, Bol.* 1, 126-127.

Montalvão, R.M.G. (1985). *Evolução geotectônica dos terrenos granitóide-greenstone belts de Crixás, Guarinos, Pilar de Goiás-Hidrolina (Goiás)*. Doctorate Thesis. USP, São Paulo, 372 p.

Montalvão, R.M.G., Hildred, P.D., Bezerra, P.E.L., Prado, P. and Silva, S.J. (1982). Petrographic and chemical aspects of the mafic-ultramafic rocks of the Crixás, Guarinos, Pilar de Goiás-Hidrolina and Goiás greenstone belts, central Brazil. *Revista Brasileira de Geociências*, 12(1/3), 331-347.

Moraes, R. (1992). *Metamorfismo e deformação da seqüência vulcano-sedimentar de Juscelândia, Goiás, e geoquímica de seus anfibolitos*. MSc. Dissertation, UnB, Brasília, 171 p.

Moraes, R. (1997). *Condições e evolução do metamorfismo no Complexo Barro Alto e litogeoquímica do vulcanismo bimodal da seqüência Juscelândia, Goianésia, Goiás*. Doctorate Thesis. UnB, Brasília, 382 p.

Moraes, R. and Fuck, R.A. (1992a). As relações entre deformação e metamorfismo nas seqüências Juscelândia e Serra da Malacacheta, Complexo de Barro Alto. *Anais, XXXVII Congresso Brasileiro de Geologia*, São Paulo, SBG, 1, 413.

Moraes, R. and Fuck, R.A (1992b). A geoquímica dos anfibolitos da seqüência vulcano-sedimentar de Juscelândia, Goiás. *Anais, XXXVII Congresso Brasileiro de Geologia, São Paulo*, SBG, 2, 27-28.

Moraes, R. and Fuck, R.A. (1994). Deformação e metamorfismo das seqüências Juscelândia e Serra da Malacacheta, Complexo de Barro Alto, Goiás. *Revista Brasileira de Geociências*, 24, 189-197

Nascimento, F.S., Veloso, F.D.L.M. and Sabóia, L. (1981). Caracterização e discussão sobre a seqüência vulcano-sedimentar da borda oeste do Maciço de Niquelândia (oeste do rio Traíras). Atas, I Simpósio de Geologia do Centro-Oeste, Goiânia, SBG, 470-492

Nilson, A.A. (1981). *The nature of the Americano do Brasil mafic-ultramafic complex and associated sulfide mineralization, Goiás, Brasil*. PhD Thesis, University of Western Ontario, 460 p.

Nilson, A.A. (1992). Geologia e aspectos petrológicos do complexo máfico-ultramáfico de Águas Claras, Araçu, Goiás. *Resumos Expandidos, XXXVII Congresso Brasileiro de Geologia, São Paulo*, SBG, 457.

Nilson, A.A, Gioia, S. and Pimentel, M.M. (1997). Idade Sm-Nd do Complexo Máfico-Ultramáfico de Americano do Brasil, Goiás e características isotópicas das rochas encaixantes. *Anais, VI Congresso Brasileiro de Geoquímica, Salvador*, 643-645.

Nilson, A.A. and Motta, F. (1969). *Geologia da área de Goiania-Trindade*. DNPM/DFPM, Rio de Janeiro, Bol. 93, 95 p.

Nunes, N.S.V. (1993). Geologia do Grupo Araxá na região de Anicuns, Goiás. *Anais, IV Simpósio de Geologia do Centro-Oeste, Brasília, SBG*, 173-176.

Oliveira, A.M. (1993). *Petrografia, estratigrafia, petroquímica e potencialidade para elementos do Grupo da Platina (EGP) do Complexo de Barro Alto, na região de Goianésia, Goiás.* MSc Dissertation, UnB, Brasília, 86 p.

Oliveira, C.G. and Pimentel, M.M. (1998). Geology of the Northern Part of the Hidrolina Dome and Sm-Nd Isotopic Data For Metavolcanic and Granitoid Rocks in the Campinorte Region, Goiás, Central Brazil. In: *International Conference on Precambrian and Craton Tectonics - International Conference on Basement Tectonics, 14., Abstracts. UFOP-International Basement Tectonics Association, Ouro Preto*, 55-56.

Padilha, J.L. (1984). Prospecção de ouro na região nordeste de Goiás – Projeto Pindorama - DOCEGEO. In: *Encontro Regional do Ouro de Goiás, 1. Goiânia, SBG, Anais*, 78-95.

Palermo, N. (1996). *Le gisement aurifère précambrien de Posse (Goiás, Brésil) dans sons cadre géologique.* These de doctorat. ENSMP, Paris, 175 p.

Pimentel, M.M. (1985). *A seqüência vulcano-sedimentar de Arenópolis – GO: petrologia ígnea e metamórfica, contexto geotectônico e considerações metalogenéticas preliminares.* MSc Dissertation, UnB, Brasília, 188 p.

Pimentel, M.M. (1991). *Late Proterozoic Crustal Evolution of the Tocantins Province in central Brazil: an Isotopic and Geochemical Study.* PhD. Thesis, University of Oxford, 250 p.

Pimentel, M.M. (1992) Reajuste do sistema isotópico Sm-Nd durante o Neoproterozóico em gnaisses do oeste de Goiás. *Revista Brasileira de Geociências*, 22(4), 1-7.

Pimentel, M.M. and Botelho, N.F. (1998). Sm-Nd isotopic characteristics of the 1.77-1.6 Ga rift magmatism in central Brazil. *Abstracts, ICOG-9 (IX International Conference on Geochronology, Cosmochronology and Isotope Geology), Chinese Science Bulletin, Chinese Academy of Sciences, Beijing*, 43, 99.

Pimentel, M.M. and Fuck, R.A. (1986). Geologia da seqüência vulcano-sedimentar de Arenópolis (GO). *Revista Brasileira de Geociências*,16, 217-223.

Pimentel, M.M. and Fuck, R.A. (1987). Origem e evolução das rochas metavulcânicas e metaplutônicas da região de Arenópolis (GO). *Revista Brasileira Geociências*, 17,2-14.

Pimentel, M.M. and Fuck, R.A. (1992). Neoproterozoic crustal accretion in central Brazil. *Geology*, 20(4), 375-379.

Pimentel, M.M. and Fuck, R.A. (1994). Geocronologia Rb-Sr da porção sudoeste do Maciço Mediano de Goiás. *Revista Brasileira de Geociências*, 24(2),104-111.

Pimentel, M.M., Fuck, R.A. and Alvarenga, C.J.S. de (1996). Post-Brasiliano (Pan-African) high-K granitic magmatism in central Brazil: late precambrian/early Paleozoic extension. *Precambrian Research*, 80, 217-238.

Pimentel, M.M., Fuck, R.A. and Botelho, N.F. (1999a). Granites and the geodynamic history of the Neoproterozoic Brasília Belt, Central Brazil: a review. *Lithos*, 46(3), 463-483.

Pimentel, M.M., Fuck, R.A., Cordani, U.G. and Kawashita, K. (1985). Geocronologia de rochas graníticas e gnáissicas da região de Arenópolis-Piranhas, Goiás. *Revista Brasileira de Geociências*, 15, 3-8.

Pimentel, M.M., Fuck, R.A. and Fischel, D.P. (1999b.) Estudo isotópico Sm-Nd regional da porção central da Faixa Brasília: implicações para idade e origem dos granulitos do Complexo Anápolis-Itauçu e sedimentos do Grupo Araxá. *Revista Brasileira da Geociências*, 29(3) (in press).

Pimentel, M.M., Fuck, R.A. and Silva, J.L.H. (1996a). Dados Rb-Sr e Sm-Nd da região de Jussara-Goiás-Mossâmedes (GO), e o limite entre terrenos antigos do Maciço de Goiás e o Arco Magmático de Goiás. *Revista Brasileira de Geociências*, 26, 61-70.

Pimentel, M.M., Heaman, L. and Fuck, R.A. (1991a). U-Pb zircon and sphene geochronology of late Proterozoic volcanic arc rock units from southwestern Goiás, central Brazil. *Journal of South American Earth Sciences*, 4, 329-339.

Pimentel, M.M., Heaman, L., Fuck, R.A. and Marini, O.J. (1991b). U-Pb zircon geochronology of Precambrian tin-bearing continental-type acid magmatism in central Brazil. *Precambrian Research*, 52, 321-335.

Pimentel, M.M., Heaman, L and Fuck, R.A. (1992). Idade do meta-riolito da seqüência Maratá, Grupo Araxá, Goiás: estudo geocronológico pelos métodos U-Pb em zircão, Sm-Nd e Rb-Sr. *Anais da Academia Brasileira de Ciências*, 64(1), 19-28

Pimentel, M.M. and Junges, S. (1997). Nd Isotopic characteristics of metasediments of the Neoproterozoic Mara Rosa Arc, Goiás, Central Brazil. *Extended Abstracts, I South American Symposium On Isotope Geology, SSAGI, Campos do Jordão*, 237-239.

Pimentel, M.M., Whitehouse, M.J., Viana, M.G., Fuck, R.A. and Machado, N. (1997). The Mara Rosa arc in the Tocantins Province: further evidence for Neoproterozoic crustal accretion in central Brazil. *Precambrian Research*, 81,299-310.

Pulz, G.M. (1995). *Modelos prospectivos para ouro em greenstone belts: exemplo dos depósitos Maria Lázara e Ogó na região de Guarinos e Pilar de Goiás.* Doctorate Thesis, UnB, Brasília, 190 p.

Potrel, A., Resende, M.G. and Jost, H. (1998). Transition in acid magmatism during Archean: example of granite-gneissic basement of the Goiás Massif. *Anais, XV Congresso Brasileiro de Geologia, SBG*, 69.

Queiroz, C.L., Jost, H. and McNaughton, N. (1999). U-Pb-SHRIMP ages of Crixás granite-greenstone belt terranes: from Archean to Neoproterozoic. *Resumos Expandidos, VII Simpósio Nacional de Estudos Tectônicos, Lençois, Bahia.*

Resende, M.G. (1994). *Geologia e geoquímica da Formação Aimbé, Grupo Guarinos, Goiás.* MSc Dissertation, UnB, Brasília, 124 p.

Resende, M.G. (1998). *Estudo da evolução das supracrustais metassedimentares dos greenstone belts de Goiás e Faina, GO.* Doctorate Thesis, UnB, Brasília, 200 p.

Resende, M.G. and Jost, H. (1994). Redefinição da Formação Aimbé, greenstone belt de Guarinos, Goiás e sua interpretação paleogeográfica e paleotectônic. *Boletim de Geociências do Centro-Oeste*, 17(1/2), 49-57.

Resende, M.G. and Jost, H. (1995). Petrogênese de formações ferríferas e metahidrotermalitos da Formação Aimbé, Grupo Guarinos (Arqueano), Goiás. *Revista Brasileira de Geociências*, 25, 41-50.

Resende, M.G., Jost, H., Osborne, G.A. and Mol, A. (1998). The stratigraphy of the Goiás and Faina greenstone belts, central Brazil: a new proposal. *Revista Brasileira de Geociências*, 28(1),77-94.

Ribeiro, R.J.C and Pimentel, M.M. (1998). Origem dos granitos neoproterozóicos associados ao Grupo Araxá na região de Ipameri, Goiás: evidências isotópicas pelo método Sm-Nd. *Anais, XV Congresso Brasileiro de Geologia, Belo Horizonte*, SBG, 42.

Ribeiro Filho, W. (1981). Reavaliação da geologia de Pilar-Mara Rosa. *Atas, Simpósio de Geologia do Centro-Oeste, Goiânia*, 281-296.

Ribeiro Filho, W., Costa, J.F.G., Araújo, V.A., Valente, C.R., Machado, E.C., Matos, S.H.S., Rodrigues, R., Vivian, O., Sá, A.M. and Lima, P.F.C. (1978). *Projeto geologia da região de Pilar-Mara Rosa.* DNPM/CPRM, Goiânia, 9 vol.

Ribeiro Filho, W. and Teixeira, N.A. (1981). Seqüência vulcano-sedimentar da borda oeste dos complexos de Niquelândia e Canabrava. *Boletim Informativo da Sociedade Brasileira de Geologia – Núcleo Centro-Oeste*, 10, 157-177

Richardson, S.V., Kessler, S.E., Essene, E.J. and Jones, L.M. (1986). Origin and Geochemistry of the Chapada Cu-Au Deposit, Goias, Brazil: A Metamorphosed Wall-Rock Porphyry Copper Deposit. *Economic Geology*, 81, 1884-1898.

Strontium isotopic geochemistry of Pan-African/Brasiliano rocks, Chapada Copper deposit. *Economic Geology*, 81, 1884-1898.

Rivalenti, G., Girardi, V.A.V., Sinigoi, S., Rossi, A. and Siena, F. (1982). The Niquelândia mafic-ultramafic complex of central Goiás, Brazil: petrological considerations. *Revista Brasileira de Geociências*, 12, 380-391.

Rodrigues, J.B. (1996). *Geocronologia e geoquímica da seqüência vulcano-sedimentar de Iporá e rochas graníticas associadas*. MSc Dissertation. UnB, Brasília, 101 p.

Rodrigues, J.B., Gioia, S.M.L.C. and Pimentel, M.M. (1999). Geocronologia e geoquímica de rochas ortognáissicas da região entre Iporá e Firminópolis: implicações para a evolução do Arco Magmático de Goiás. *Revista Brasileira de Geociências*, 29(3) (in press).

Sabóia, L.A.. 1979. Os greenstone belts de Crixás e Goiás, GO. *Boletim Informativo SBG-NCO*, 9, 44-72.

Sabóia, L.A. and Teixeira, N.A. (1983). Ultramafic flows of the Crixás greenstone belt, Goiás, Brazil. *Precambrian Research*, 22, 23-40.

Sabóia, L.A., Teixeira, L.A., Castro, J.H.G. and Teixeira, H.S. (1979). Geologia do greenstone belt de Crixás (GO) e suas implicações geotectônicas. *Anais, I Simpósio do Craton do São Francisco e suas Faixas Marginais, Salvador*, SME/CPM/SBG, 39-50.

Sato, K. (1998). *Evolução crustal da Plataforma Sul Americana, com base na geoquímica isotópica Sm-Nd*. Doctorate Thesis, USP, São Paulo, 290 p.

Schobbenhaus, C., Campos, D.A., Derze, G. and Asmus, H.E. (1981). *Mapa geológico do Brasil e da área oceânica adjacente, incluindo depósitos minerais*. DNPM, Brasília.

Seer, H.J. (1985). *Geologia, deformação e mineralização de cobre no complexo vulcano-sedimentar de Bom Jardim de Goiás*. MSc Dissertation, UnB, Brasília, 190 p.

Silva, F.O. (1991). *Geologia, estrutura, petrologia e mineralizações de Fe, Ti e V associadas ao complexo gabro-anortosítico acamadado de Santa Bárbara (Goiás)*. MSc Dissertation, UnB, Brasília, 190 p.

Silva, F.O. (1997). *Geologia e petrologia do extremo noroeste do Complexo máfico-ultramáfico de Taquaral, GO*. Doctorate Thesis, UnB, Brasília, 171 p.

Silva, F.O. and Nilson, A.A. (1990). Caracterização do complexo gabro-anortosítico acamadado de Santa Bárbara, Goiás. *Anais, XXXVI Congresso Brasileiro de Geologia, Natal*, SBG, 1898-1912.

Sousa, J.M. and Frizzo, C. (1995). Sulfeto maciço de Juscelândia-GO. *Anais, V Simpósio de Geologia do Centro-Oeste, Cuiabá*, SBG, 51-55.

Sparrenberger, I. and Tassinari, C.C.G. (1998). Idade da mineralização estanífera na Subprovíncia do Rio Paranã (GO) através de datações U-Pb em cassiterita. *Congr. Bras. Geologia, 40. Belo Horizonte, SBG, Anais*, p. 143.

Suita, M.T.F. (1996). *Geoquímica e metalogenia de elementos do grupo da platina (EGP + Au) em complexos máfico-ultramáficos do Brasil: critérios e guias com ênfase no complexo máfico-ultramáfico acamadado de alto grau de Barro Alto*. Doctorate Thesis, UFRGS, Porto Alegre, 600 p.

Suita, M.T.F., Kamo, S.L., Krogh, T.E., Fyfe, W.S. and Hartmann, L.A. (1994). U-Pb ages from the high-grade Barro Alto mafic-ultramafic complex (Goiás, Central Brazil): middle Proterozoic continental mafic magmatism and upper Proterozoic continental collision. *Abstracts, VIII International Conference on Geochronology, Cosmochronology and Isotope Geology, Berkeley*, USGS, p. 309.

Tassinari, C.C.G. and Montalvão, R.M.G. (1980). Estudo geocronológico do greenstone belt de Crixás, GO. *Anais, XXXI Congresso Brasileiro de Geologia, Camboriú, SBG*, 5, 2752-2759.

Tassinari, C.C.G., Siga Jr., O. and Teixeira, W. (1981). Panorama geocronológico do centro-oeste brasileiro: solução, problemática e sugestões. *Atas, I Simpósio de Geologia do Centro-Oeste, Goiânia*, SBG, p. 120.

Teixeira, N.A., Poli, A.J.C. and Ferreira, M.C.B. (1982). Contribuição à geologia e petrologia da região de São Domingos - GO. *Revista Brasileira de Geociências*, 12(4), 562-571.

Theodoro, S.C. (1995). Ambiente de sedimentação da Formação Ribeirão das Antas, Grupo Crixás, Goiás, Brasil. *MSC dissertation, UnB, Brasília, 88 p.*

Thomsen, F.P.R. and Kuyumjian, R.M. (1994). As formações ferríferas bandadas da região de Conceição do Tocantins - TO: caracterização preliminar. *Congr. Bras. Geologia, 38. Balneário Camboriú, SBG, Anais*, 1, 72-73.

Tomazzoli, E.R. (1985). *Geologia, petrologia, deformação e potencial aurífero do greenstone belt de Goiás/GO*. MSc Dissertation, UnB, Brasília, 206 p.

Vargas, M.C. (1992). *Geologia das rochas granito-gnáissicas da região de Crixás, Guarinos, Pilar de Goiás e Hidrolina*. MSc Dissertation, UnB, Brasília, 167 p.

Viana, M.G. (1995). *Geocronologia e geoquímica das rochas ortognáissicas e metavulcânicas da região de Mara Rosa - GO*. MSc Dissertation, UnB, Brasília. 118 p.

Viana, M.G., Pimentel, M.M., Whitehouse, M.J., Fuck, R.A. and Machado, N. (1995). O Arco Magmático de Mara Rosa, Goiás: dados geoquímicos e geocronológicos e suas implicações regionais. *Revista Brasileira de Geociências*, 25, 111-123.

Winge, M. (1995). *Evolução dos terrenos granulíticos da Província Tocantins, Brasil Central*. Doctorate Thesis. UnB, Brasília, 550 p.

Winge, M. and Danni, J.C.M. (1994). A transição entre a Seqüência gabro-anortosítica Serra da Malacacheta (SSM) e os metabasaltos da base da Seqüência Juscelândia (SJ) - implicações na gênese dos Complexos Barro Alto (CBA), Niquelândia (CA) e Cana Brava (CC), Goiás, *Atas, IV Simpósio de Geologia do Centro Oeste*, Brasília, SBG, 131-134.

Wolff, F. (1991). *Caracterização petrográfica, litogeoquímica, química mineral e geotermometria de rochas de alto grau metamórfico do Complexo Anápolis-Itauçu, Goiás*. Dissertação de Mestrado, UFRJ, Rio de Janeiro, 148 p.

Yamaoka, W. and Araújo, E.M. (1988). Depósito de ouro de Mina III, Crixás, Goiás. In: *Principais Depósitos Minerais de Brasil*, ed. DNPM/CVRD, 3, 491-498, DNPM, Brasília.

THE BRASÍLIA FOLD BELT

Marcel A. Dardenne

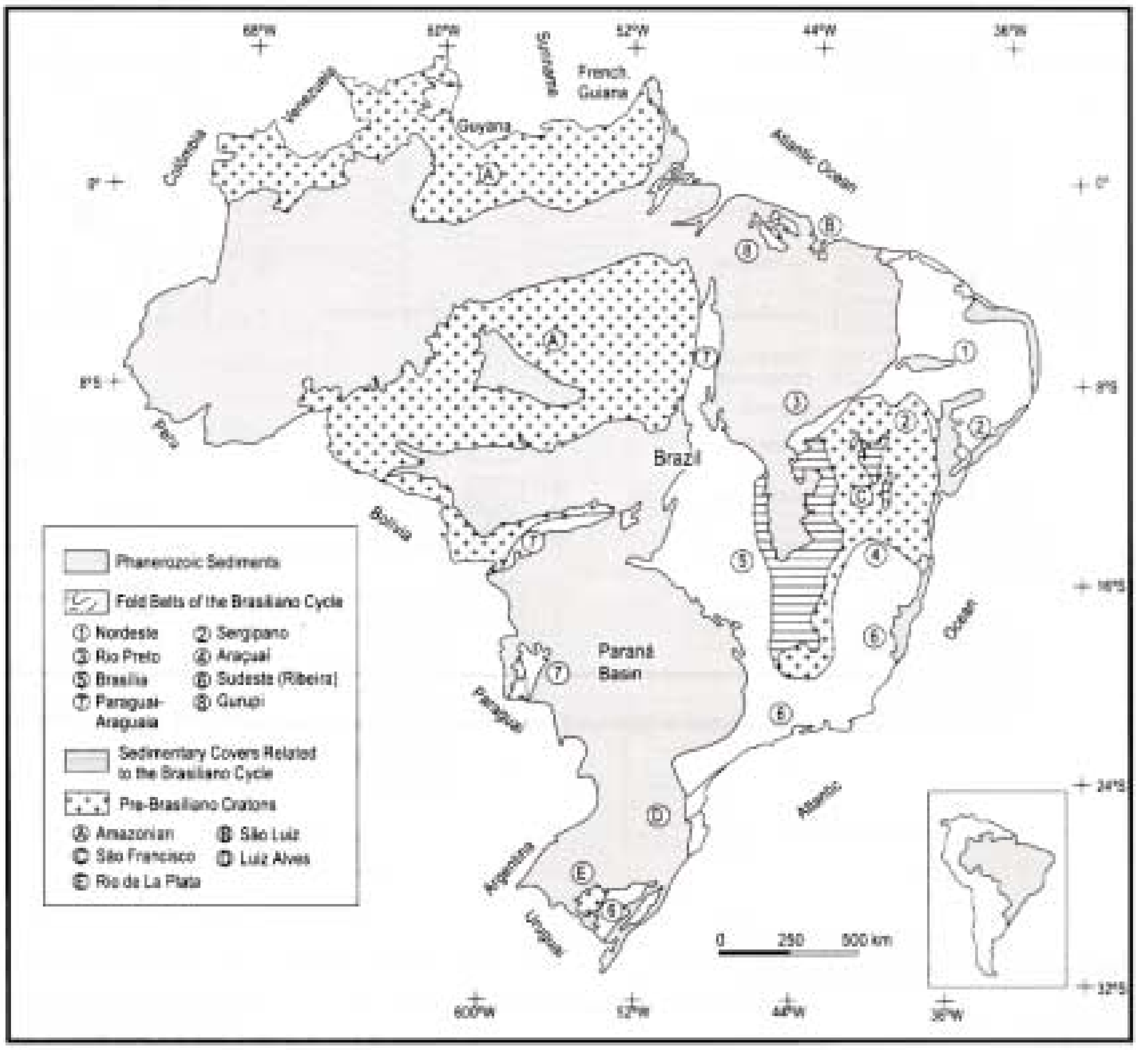
In central Brazil, the Tocantins Province represents an orogen of large dimensions that developed during the Neoproterozoic as a function of the convergence and collision of three important continental blocks. These blocks are the Amazonian Craton (AC) in the NW, the São Francisco Craton (SFC) in the E and a third supposed craton below the Paraná Basin in the SW. This orogen is formed by the Paraguai-Araguaia and Brasília fold belts. (Fig. 1)

The Brasília Fold Belt (BFB) is situated in the eastern part of the Tocantins Province, and extends for more than 1000 km in a general N-S direction along the western margin of the SFC (Fig. 1). Almeida (1977) defined the Tocantins Province as an important unit of the geological framework of Brazil that was stabilized at the end of the Transamazonian Cycle (*c.* 2.0 Ga). The Province is limited by the fold belts related to the Brasiliano Cycle (*c.* 600 Ma): the Brasília, Rio Preto, Riacho do Pontal, Sergipana, Araguaia, Alto Rio Grande and Ribeira fold belts (Fuck *et al.*, 1993).

The main units of the BFB (Fig. 2) are:

- Thick sedimentary and metasedimentary sequences: Araí and Serra da Mesa groups (Paleo/Mesoproterozoic); Paranoá and Canastra groups (Mesoproterozoic); Araxá, Ibiá and Vazante groups (Meso/Neoproterozoic); Bambuí Group (Neoproterozoic).
- Igneous intrusions and volcano-sedimentary sequences of various ages: Mafic-ultramafic complexes of Niquelândia, Cana Brava and Barro Alto (Paleo/Mesoproterozoic); anorogenic granite of the sub-provinces of Rio Paraná and Rio Tocantins (Paleo/Mesoproterozoic); volcano-sedimentary sequences of Juscelândia, Palmeirópolis and Indaianópolis (Mesoproterozoic); syn, late and post-orogenic granitic and mafic-ultramafic magmatism (Neoproterozoic).

FIGURE 1 - Fold belts of the Brasiliano Cycle and correlated cratons (after Schobbenhaus et al., 1984).



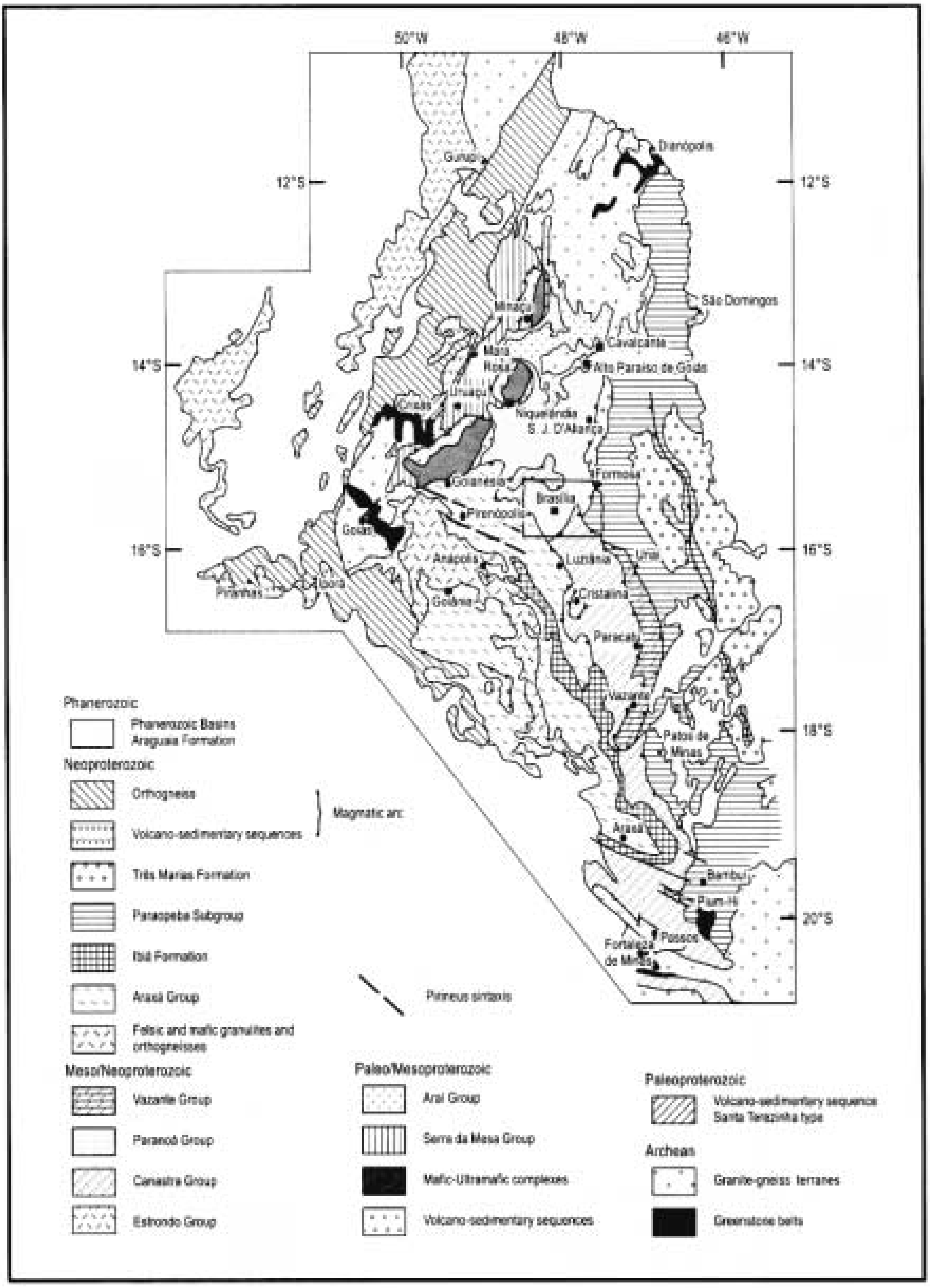


Fig. 2 - Lithostratigraphic columns and geological map of Paleo-Meso-Neoproterozoic units in the Brasília fold Belt (from Marini et al., 1974)

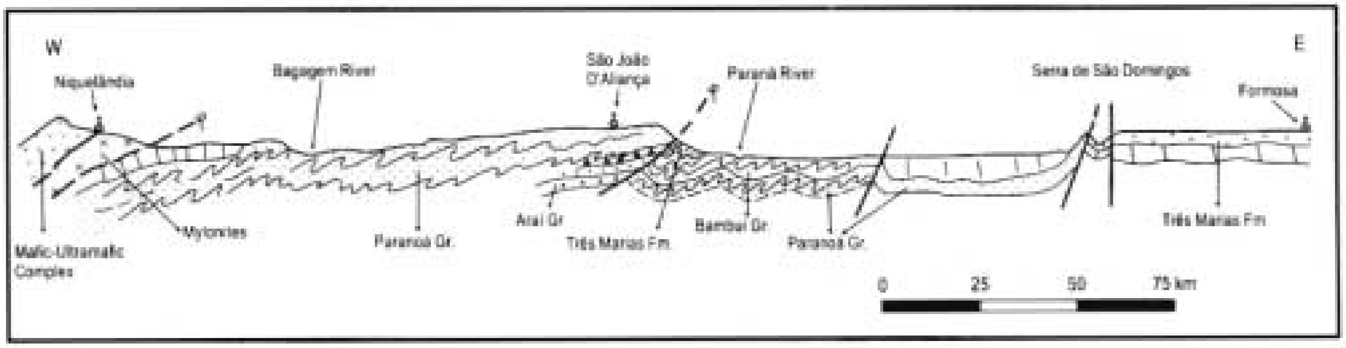


Fig. 3a - Schematic cross-section of the Brasília Belt between Niquelândia and Formosa - Goiás, Central Brazil (after Dardenne, 1978).

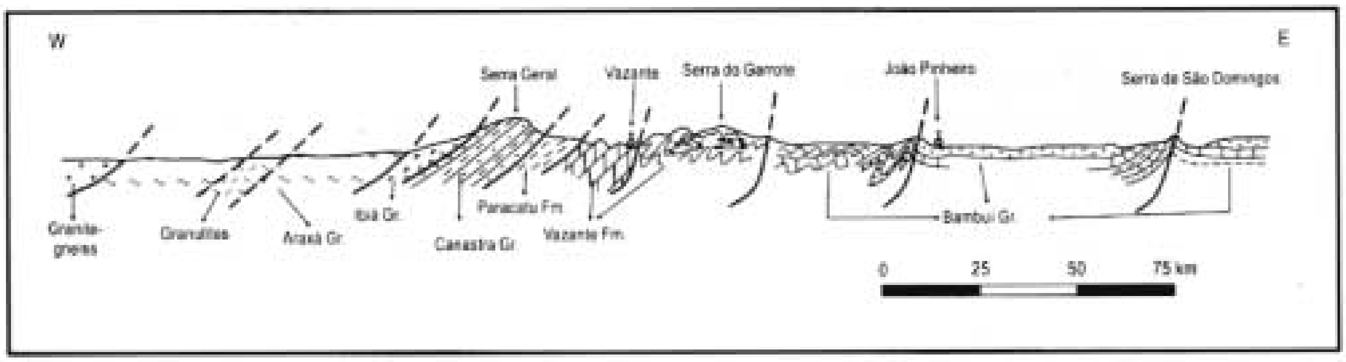


Fig. 3b - Schematic cross-section showing the relationships between the Araxá, Ibiá, Canastra, Vazante, and Bambuí groups (after Dardenne, 1978).

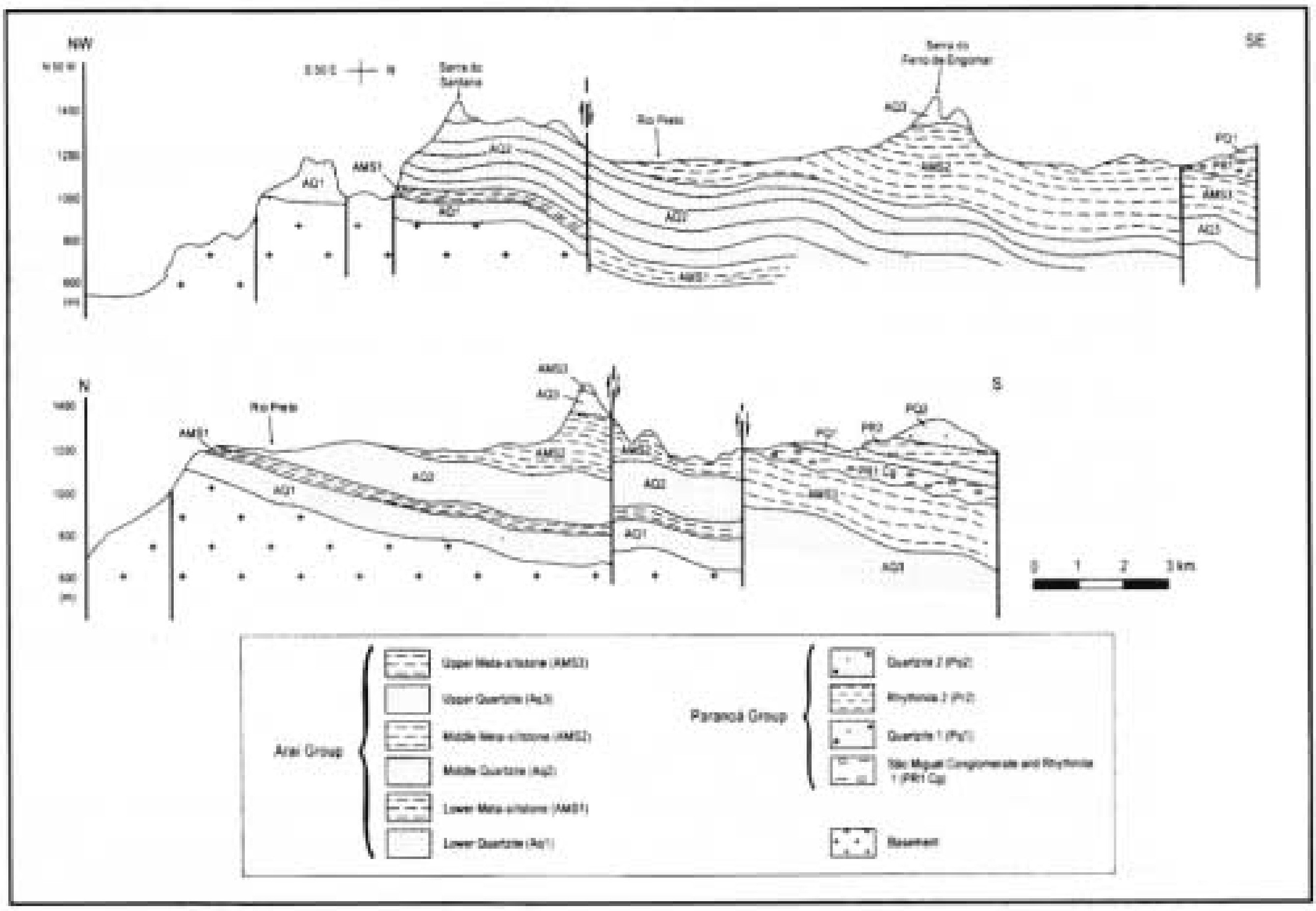


Fig. 4 - Geological cross-sections of the Araí and Paranoá groups in the Chapada dos Veadeiros National Park - Goiás, Central Brazil (after Dardenne et al., 1997).

• The Magmatic Arc of Goiás (Neoproterozoic), formed by the volcano-sedimentary sequences of Mara Rosa and Chapada and by the tonalitic/granodioritic rocks of juvenile origin that occur over large areas of the western part of the BFB.

In addition to these units, the BFB involves a more ancient unit known as the Goiás Massif, which includes the Archean granite-greenstone terranes of Goiás Velho, Crixás, Guarinos and Pilar de Goiás; the Paleoproterozoic granite-gneiss basement of Cavalcante; the Paleoproterozoic volcano-sedimentary sequences of Santa Terezinha, Almas-Dianópolis and São Domingos; and the Ticunzal meta-sedimentary sequence.

Tectonic Framework of the Brasília Fold Belt

In general, the main sedimentary and metasedimentary units of the BFB show tectonic deformation that is progressively more intense towards the W, accompanied by increasing metamorphism. The metamorphic grade varied from non metamorphosed sediments in the cratonic area to amphibolite and even granulite facies metamorphism in the western part of the BFB. This evolution of the deformation and associated metamorphism reflects the clear vergence of the BFB with respect to the SFC. This tectonic zonation, initially proposed by Costa and Angeiras (1971), was described by Dardenne (1978) and reformulated by Fuck *et al.* (1994), after the approximate definition and individualization of the Internal Zone to the W, the External Zone in the central region, and Cratonic Zone to the E.

However, since the studies of Marini *et al.* (1981, 1984a, b), the importance of the mega-inflexion or syntaxis of the Pirineus Lineament was stressed. This is a WNW-ESE lineament situated on the same latitude as the Federal District that permits the subdivision of the BFB into northern and southern segments showing a singular geotectonic evolution, but with very distinct characteristics (Fonseca, 1996; Fonseca *et al.*, 1995; Strieder, 1993; Araújo Filho, 1999).

In the northern segment of the BFB, most of the sedimentary units were not metamorphosed or only underwent low greenschist metamorphism. Its stratigraphic relationships remain well preserved, permitting detailed lithostratigraphy and the reconstruction of the paleogeography and depositional systems. This characteristic is due to the positioning in an upper crustal level of the granite-gneiss basement, which acted as a rigid block in front of the compressive trend of the Brasiliano Cycle. In the area situated immediately to the N of Brasília, there occur the sedimentary sequences of the Paranoá and Araí groups. Here the effects of compression are expressed by the development of great dextral transcurrent faults and thrusts that caused the motion of the basement but affected, only locally, the sedimentary cover (Fig. 3a). The metasedimentary sequences that display a metamorphic grade higher than the amphibolite facies occur only in areas to the W of the megasuture. These rocks include the mafic/ultramafic complexes of Niquelândia-Barro Alto-Cana Brava, metamorphosed in the granulite facies. In general, the vergence of the deformation observed in the northern segment of the BFB indicates a main compressive trend with a NW-SE strike (Fonseca *et al.*, 1995; Fonseca and Dardenne, 1995; Araújo Filho, 1999). According to Araújo Filho (1999), the northern segment of the BFB was thrusted over the southern segment at the end of the Brasiliano Cycle (Roscoe and Araújo Filho, 1994).

The southern segment of the BFB shows distinct tectonic features when compared with the northern segment. The deformation and the associated metamorphism were very intense, obliterating the stratigraphic relationships between the various units (Fig. 3b). The Araxá, Canastra, Ibiá and Vazante groups were involved in a complex imbricate system of nappes and thrusts indicating tectonic transport of great magnitude, in the order of tens to hundreds of kilometres. The contacts between the various assemblages involved correspond to low angle shear zones, frequently showing the characteristic arched form of sheath folds developing lateral sheared ramps (Strieder, 1993, 1994; Simões, 1995; Simões, 1991; Simões and Navarro, 1996; Seer *et al.*, 1998; Schmidt and Fleischer, 1978; Teixeira and Danni, 1978; Valeriano *et al.*, 1997; Valeriano, 1992; Simões and Valeriano, 1990; Seer, 1999; Araújo Filho, 1999). Towards the SFC, the deformation and the associated metamorphism decrease progressively giving a measure of justification for the division of the BFB in Internal, External and Cratonic zones, the contacts of which are marked by great regional faults oriented N-S (Dardenne, 1978). In general, the vergence of the initial deformation observed in the southern segment of the BFB indicates a main compressive trend oriented from SW to NE. This is followed by predominantly SE transport marked by regional sinistral transcurrent shear zones showing the same orientation (Seer, 1999; Valeriano *et al.*, 1997; Araújo Filho, 1999).

Description of the Brasília Fold Belt Units

With a view to pointing out the differences observed between the northern and southern segments of the BFB, the description of the units will be done according to the respective segment in which the units occur. However, this does not apply to the Bambuí Group and Goiás Magmatic Arc, the characteristics of which are the same through all the BFB.

Units of the northern segment of the BFB

The various units described in the northern segment of the BFB are the Araí-Serra da Mesa-Paranoá groups and the volcano-sedimentary sequences of Juscelândia-Indaianópolis-Palmeirópolis.

Araí Group

The Araí Group (Barbosa *et al.*, 1969) represents a thick sedimentary sequence (around 1500 m) of clastic and pelitic nature, lying on an erosional and angular unconformity over two main units. These units are the granite-gneiss basement of the Cavalcante-Dianópolis region and the Ticunzal Formation (Marini *et al.*, 1978), and the anorogenic tin-

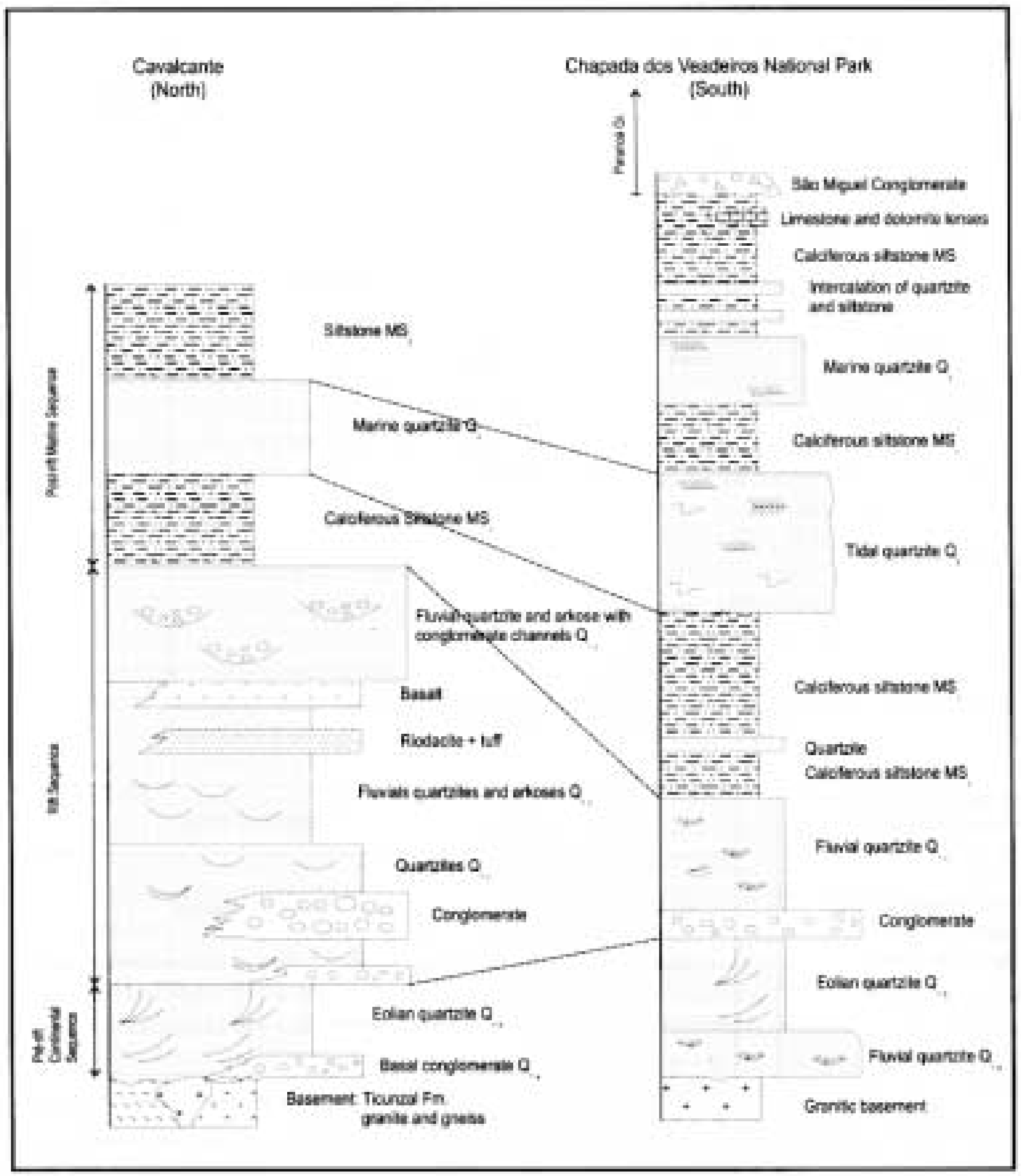
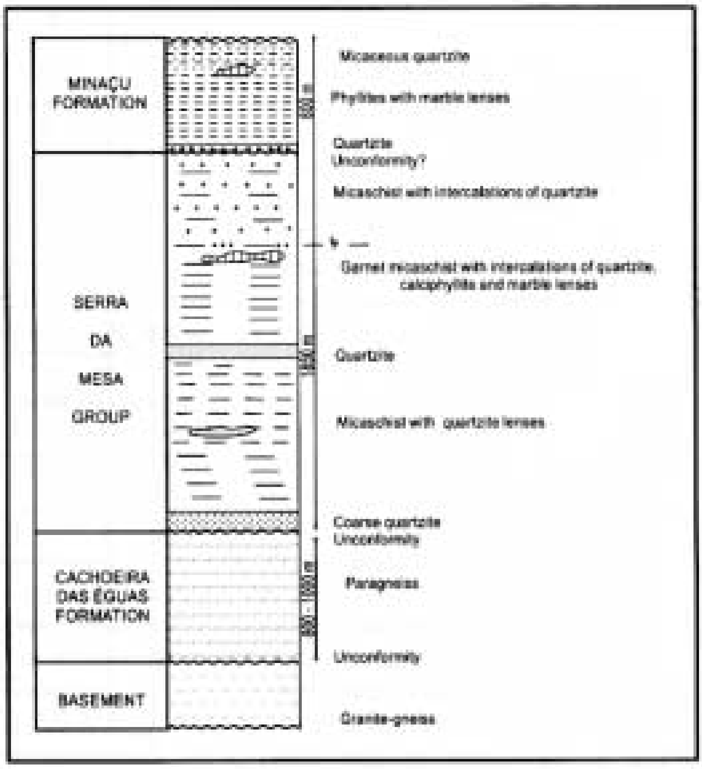


FIGURE 5 - Schematic lithostratigraphic columns of the Araí Group in the regions of the Cavalcante and Chapada dos Veadeiros National Park, Goiás (after Dardenne et al., 1997, 1999).

FIGURE 6 - Lithostratigraphic column of the Serra da Mesa Group (after Marini and Fuck, 1977).



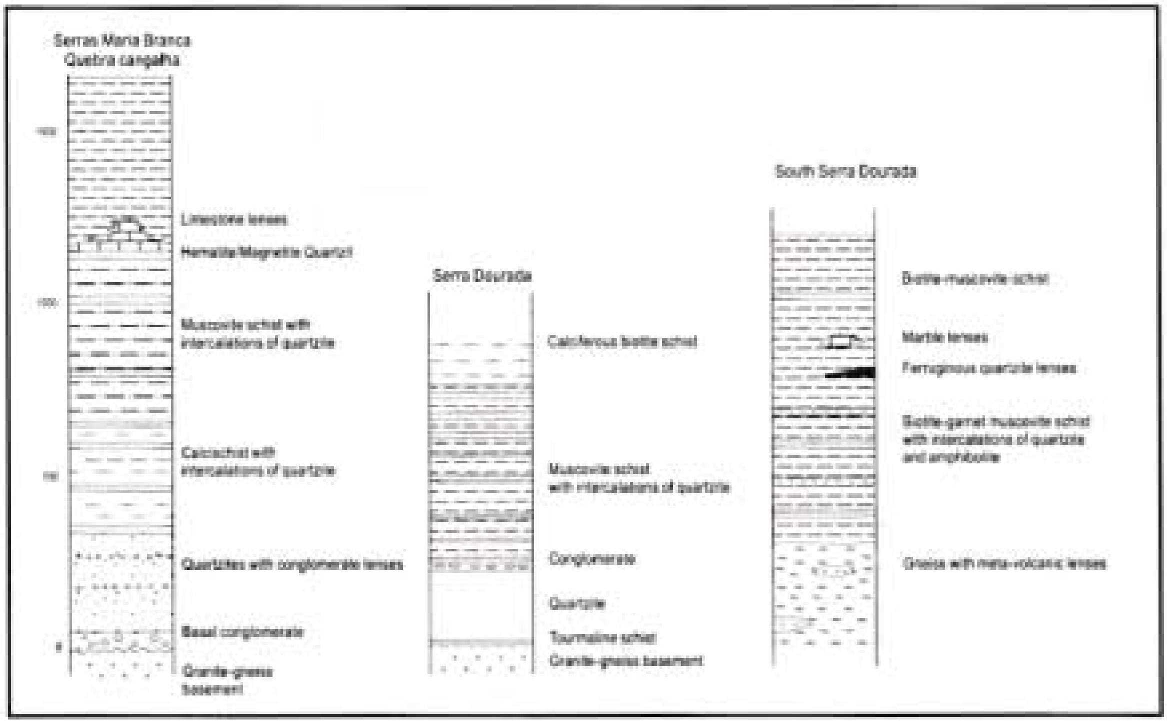
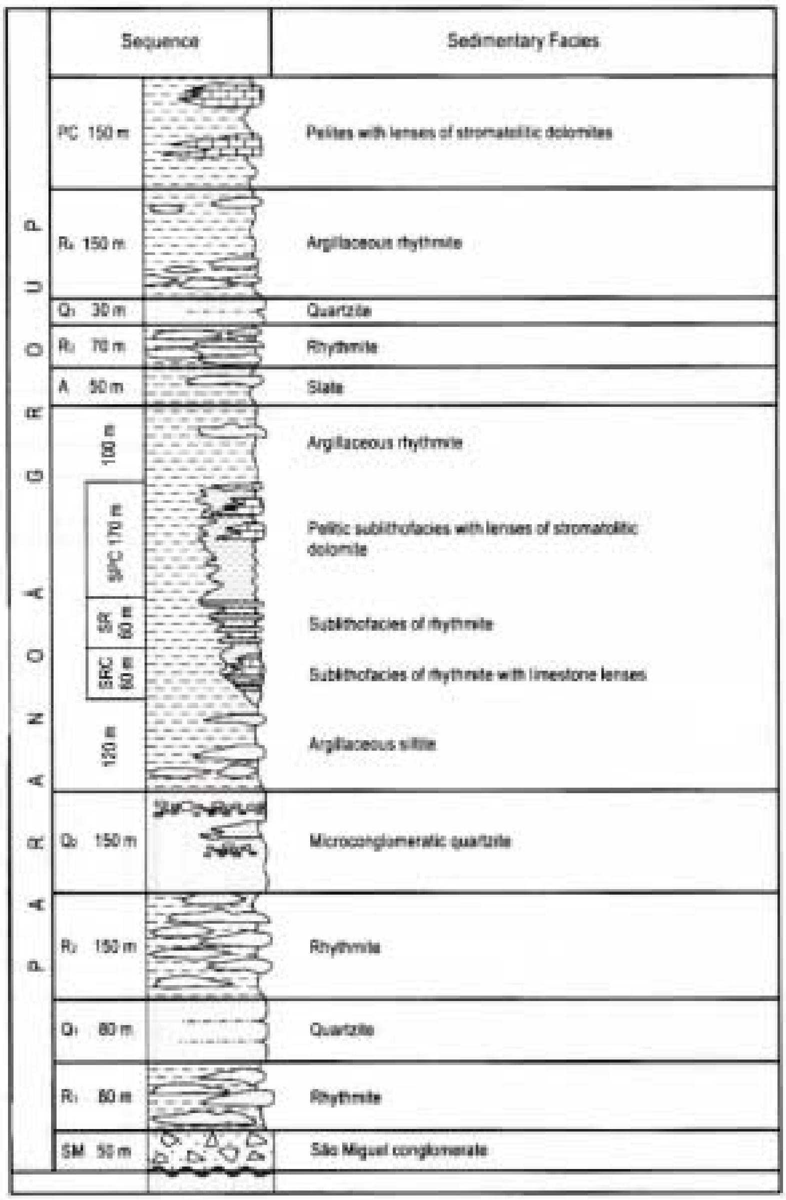


FIGURE 7 - Lithostratigraphic columns of the Serra Dourada Group (after Dardenne et al., 1981).

FIGURE 8 - Lithostratigraphic column of the Paranoá Group in the Alto Paraíso region (after Faria, 1995).



bearing granites of the Rio Paranã Sub-province, related to the Paleoproterozoic. The Natividade Group, situated at the northern edge of the BFB, shows a similar sedimentary sequence to the Araí Group, being probably coeval (Marini *et al.*, 1984a, b).

According to classical descriptions, the Araí Group is divided into two formations: The Arraias Formation at the base and the Traíras Formation at the top (Barbosa *et al.*, 1969). The Arraias Formation consists of thick quartzite beds associated with intraformational conglomerate units, intercalated with metasiltstone. Volcanic rocks are associated with the clastic sediments in the lower parts of this formation. They consist of basalt, andesite, dacite, rhyolite and volcanic agglomerate. The acid alkaline volcanism was dated at 1.771 ± 0.002 Ga by U/Pb analysis in zircon by Pimentel *et al.* (1991). Coeval with this volcanism are tin-bearing granites of the Rio Paranã Sub-province (Marini and Botelho, 1986; Pimentel *et al.*, 1991; Botelho, 1992). A younger generation of anorogenic tin granites dated at 1.56 Ga (Pimentel *et al.*, 1991) and related to a reactivation of the rift, occurs more to W in the Rio Tocantins Sub-province (Marini and Botelho, 1986). The Traíras Formation consists mainly of pelitic sediments with some carbonate and thick quartzite layers.

The typical sequence of this group occurs in the northern part of the BFB, and more specifically to the N of Alto Paraíso, in the Chapada dos Veadeiros National Park and in Cavalcante-Teresina de Goiás-Araí-Arraias region. Recent field work (Dardenne *et al.*, 1997; Dardenne *et al.*, 1999) helped to clarify the evolution of the Araí sedimentary sequence associated with the development of the rift in the areas around Cavalcante and São Jorge (Chapada dos Veadeiros National Park). The following units are defined (figs. 4 and 5):

• Continental pre-rift sequence, showing essentially continental eolian and fluvial sediments.

• Rift sequence, consisting of an abundance of alluvial conglomerate and intraformational breccia and intercalations of volcanic rocks, evolving upwards to braided fluvial sediments.

• Post-rift sequence, marine and transgressive, marked by carbonate-bearing pelite beds, intercalated with thick beds of quartzite, locally with lenses of limestone and dolomite. These beds are interpreted to have been deposited on a platform dominated by tide and storm currents.

The sedimentary provenance defined from paleocurrents indicates a preferential trend from NNE to SSW, the source of the transported material being related to the granite-gneiss basement.

Serra da Mesa Group

The Serra da Mesa Group (Fig. 6) was defined by Marini *et al.* (1977, 1984a, b), and Fuck and Marini (1981) in the northern part of the BFB around the granite-gneiss brachyanticlines of Serra Dourada, Serra do Encosto and Serra da Mesa. It consists of a thick sequence of quartzite and micaschist (around 1850 m), which has been variably correlated with the Araxá Group (Barbosa *et al.*, 1969; Fuck and Marini, 1981) and to the Araí Group (Marini *et al.*, 1984a, b). The base of the group consists of coarse and conglomeratic quartzite beds (about 80m) passing to fine-grained, laminated and muscovite rich quartzite. The micaschist, intercalated with micaceous and fine-grained quartzite beds, contain garnet, staurolite and kyanite, indicative of amphibolite facies metamorphism. In the upper part of the micaschist unit there occur thick lenses of calc-schist and marble.

The Serra da Mesa, Serra do Encosto and Serra Dourada tin granites belong to the Rio Tocantins Sub-province (Marini and Botelho, 1986) and are classified as anorogenic granites, in that they do not show evidence of being intrusive into the Serra da Mesa metasediments. Dating of these granites by U/Pb and Pb/Pb indicates ages between 1.658 and 1.574 Ga (Pimentel *et al.*, 1991), implying a younger age for the Serra da Mesa Group, which is intruded by pegmatitic granitoid of uncertain age. The Peixe metasyenite dated at 1.47 Ga by Rossi *et al.* (1996), is considered intrusive into the micaschist correlated with the Serra da Mesa Group supporting the hypothesis advanced by Marini *et al.* (1984a, b) for a correlation with the Araí Group. At the same time, the Serra da Mesa micaschist tectonically overlies the volcano-sedimentary sequences of Palmeirópolis, Juscelândia and Indaianópolis dated at 1.3 Ga by Araújo *et al.* (1995) and Correia *et al.* (1999). These data permit the consideration that the Serra da Mesa metasediments were deposited in an interval between 1.574 and 1.47 Ga, which may correspond approximately to the period of post-rift transgression for the Araí Group.

In the region of Goiás Velho, the metasedimentary sequence, denominated the Serra Dourada Group (Fig. 7) by Dardenne *et al.* (1981), shows many similarities with the Araí lithostratigraphy, to which it has been correlated. The data presented show that the problem of the time relationships between the Serra da Mesa, Araí and Paranoá groups is far from being settled, as it probably involves units of different ages not yet differentiated in the Internal Zone of the BFB.

Paranoá Group

The sandy, pelitic and carbonate rocks, assigned to the Paranoá Group (Barbosa *et al.*, 1969, 1970; Dardenne, 1978, 1979, 1981; Baeta *et al.*, 1978; Dardenne and Faria, 1985; Faria and Dardenne, 1995; Faria, 1995; Dardenne *et al.*, 1997), represent an important lithostratigraphic unit that is separated by an unconformity from the Araí Group at the base and from the Bambuí Group at the top (Dardenne, 1978, 1979, 1981). This group occupies large areas in the northern segment of the BFB, to the N of the Federal District. The Paranoá Group has been described in detail in the regions of Alto Paraíso, where the type section of this group (Fig. 8) occurs; and also in the areas of São João da Aliança, São Gabriel and the Federal District (Fig. 9).

The São Miguel paraconglomerate, some 50 m thick, and first described by Dyer (1970), constitutes the base of the Paranoá Group, lying on an erosional unconformity over the sediments of the Araí Group. It consists of quartzite, calcareous siltstone and muddy limestone clasts in a pelitic-carbonate matrix of greenish colour. It occurs in the form of thick and massive beds, sometimes with incipient local stratification with cross and convolute structures. It is overlain directly by quartz, pelitic and carbonate-bearing rhythmite beds with mud cracks, and evaporite layers replaced by carbonates and cubic salt molds. This assemblage, characteristic of a tidal to supratidal environment, is followed by marine sediments consisting of and alternation of rhythmite beds and thick zones of quartzite deposited on a platform dominated by tidal currents.

The upper part of the Paranoá Group displays a more varied depositional environments, reflecting significant fluctuations of the sea level: deep water pelite beds; tidal rhythmite units and quartzite; storm rhythmite and quartzite beds; lenticular intercalations of micritic, intraclastic and oolitic grey limestone units and stromatolitic dolomite with cyanobacteria mats and columns with convex and conical laminations. These columnar stromatolites, identified as *Conophyton metula Kirichenko* type (Dardenne *et al.*, 1976; Melo Filho, 1996), suggest sedimentation between 1.2 Ga and 900 Ma, corresponding to the interval between the deposition of the Araí and Bambuí groups.

In the Alto Paraíso region, paleocurrent indicators show a main N-S source for the clastic material, similar to that observed for the sediments of the Araí Group. In the São João da Aliança and São Gabriel regions, situated to the S, the measured paleocurrent directions indicate a transport from E to W (Faria, 1995). In the region of Formosa, Cabeceiras and Bezerra (Guimarães and Dardenne, 1989), only the upper part of the Paranoá Group is exposed in the centre of large anticlines of the External Zone, where beds of arkose, locally conglomeratic, and intercalated with beds of stromatolitic dolomite and glauconitic rhythmite have been observed. Provenance studies by Guimarães (1997) suggest sedimentation on a passive margin (Fig. 10).

Towards the Internal Zone of the BFB, the Paranoá Group shows the increasing P-T conditions of the metamorphism (Fuck *et al.*, 1988). Concomitantly, the thickness of the sandy zones decrease. This assemblage was assigned to the Minaçu Formation (Marini and Fuck, 1981), in which there predominate intercalations of metarhythmite and beds of limestone and dolomite with columnar stromatolites, showing convex and conical laminations, overlain by beds of arkose (Fig. 11). In the region of Castelã, the dolomitic lenses intercalated with micaschist contain stratiform Pb-Zn sulfide mineralization, which are similar to the sulfide occurrence found in the Uruaçu area. The ages obtained from lead isotope determinations in galena approximate 1.2 Ga (Freitas-Silva and Dardenne, 1997).

Palmeirópolis, Juscelândia and Indaianópolis Volcano-Sedimentary Sequences

These volcano-sedimentary sequences (Fig. 12), defined by Danni and Leonardos (1980); Fuck *et al.*, (1981); Nascimento *et al.* (1981); Ribeiro Filho and Teixeira (1980); lie along the western margin of the mafic-ultramafic complexes of Cana Brava, Niquelândia and Barro Alto, from which they are separated by shear zones (Brod and Jost, 1994; Ferreira Filho *et al.*, 1992). They consist of fine-grained amphibolite, generally associated with gneiss, metachert and micaschist with biotite-muscovite-staurolite. In this association, metabasalt, metandesite, metarhyolite and acid to intermediate meta-volcaniclastics rocks are observed. Characteristic pillow lavas occur in the metabasalt units of the Palmeirópolis and Juscelândia sequences. Petrochemical studies of the amphibolite (metabasalt) have demonstrated that they are similar to ocean floor tholeiite (Araújo and Nilson, 1987, 1988; Araújo *et al.*, 1995; Araújo, 1998).

Pb/Pb and U/Pb analyses indicate an age *c.* 1.25 Ga for the Palmeirópolis sequence suggesting that the Mesoproterozoic volcano-sedimentary sequences constitute the oceanic basin equivalent of the Paranoá Group. This ocean floor volcanism could have been contemporaneous with the gabbro-anorthosite intrusions that occur in the western part of the Barro Alto and Niquelândia complexes, which have been dated at *c.* 1.3 Ga by U/Pb (Suita, 1998) and Sm/Nd.

In the Palmeirópolis and Juscelândia sequences, massive sulphide deposits are associated with the basaltic volcanism (Figueiredo *et al.*, 1981). The sulphides include pyrrhotite, pyrite, sphalerite, chalcopyrite and galena. This VMS-type mineralization is accompanied by intense hydrothermal alteration, which is manifest as felsic schist with antophyllite, biotite, plagioclase, sillimanite and quartz (Araújo, 1998).

Units of the southern segment of the BFB

The units described in this segment are the Canastra, Ibiá, Araxá and Vazante groups.

Canastra Group

The Canastra Group, defined by Barbosa (1955) and Barbosa *et al.* (1970), represents an association of psammitic and pelitic metasediments frequently containing carbonate, and consisting essentially of phyllite and quartzite, metamorphosed in the greenschist facies. The lithostratigraphy of the Canastra Group (Fig. 13), still not well known, was established by Campos Neto (1984a,b), and recently described in more detail by Pereira (1994) and Pereira *et al.* (1993). Freitas-Silva and Dardenne (1994) divided the Canastra Group into three formations in the northwestern region of Minas Gerais:

• The Serra do Landim Formation. The base of this formation was defined by Madalosso and Valle (1978), Madalosso (1980), as part of the Vazante Formation, being attributed to the Canastra Group by Freitas-Silva and Dardenne (1994). It consists of calc-phyllite or calc-schist of pale green to greenish grey color.

• The Paracatu Formation, initially defined by Almeida (1969), is represented by thick beds of grey carbonaceous phyllite with some intercalations of white quartzite. The Paracatu Formation was subdivided into two members by Freitas-Silva and Dardenne (1994): the lower Morro do Ouro Member begins with a relatively continuous level of quartzite with thickness varying from some metres to more than 100 m, passing to thick beds of carbonaceous phyllite with thin levels of quartzite; the upper Serra da Anta Member is represented by thick grey, greenish grey phyllite with some carbonaceous intercalations and thin quartzite layers.

• The Chapada dos Pilões Formation, the type area of which

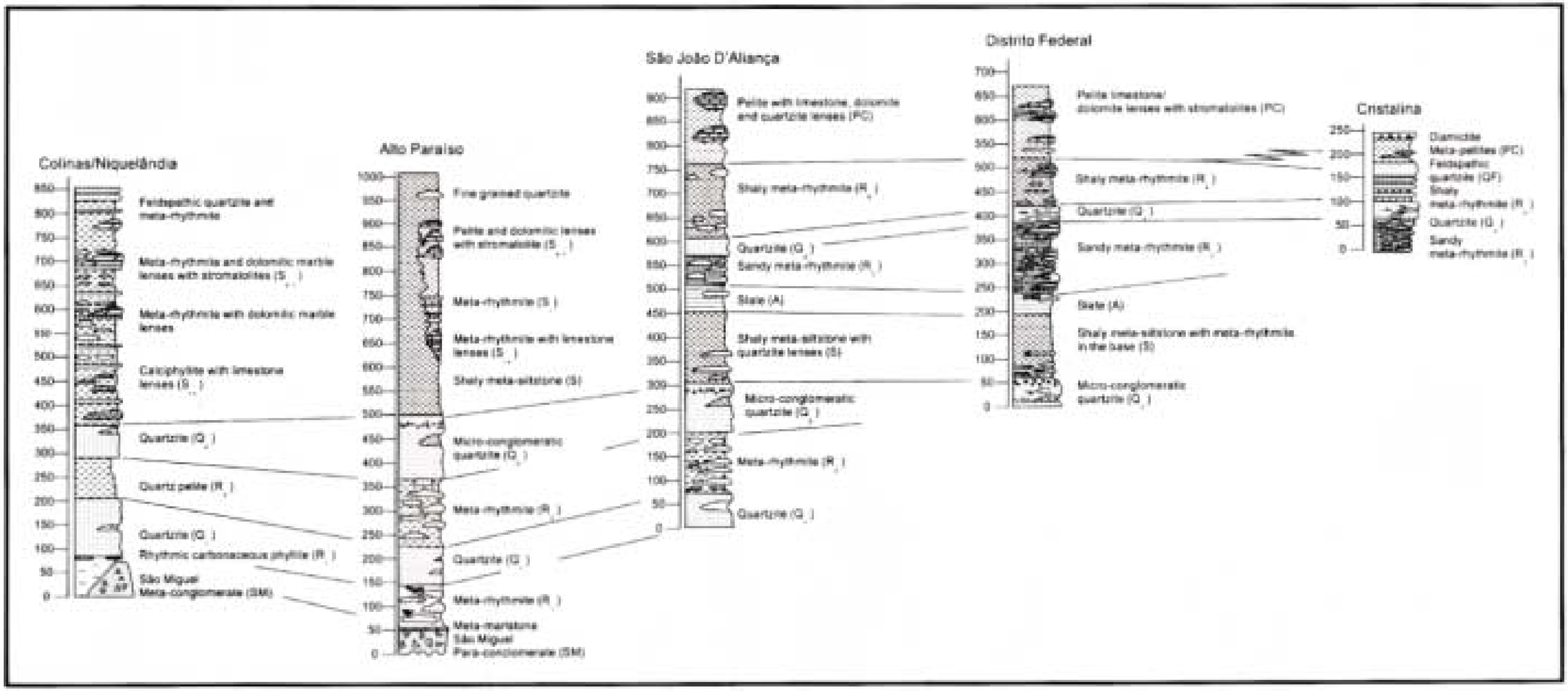


Fig. 9 - Lithostratigraphic correlations for the Paranoá Group in the areas of Colinas, Alto Paraíso, São João D'Aliança, Distrito Federal and Cristalina (from Faria, 1995).

FIGURE 9 - Lithostratigraphic correlations for the Paranoá Group in the areas of Colinas, Alto Paraíso, São João D'Aliança, Distrito Federal and Cristalina (after Faria, 1995).

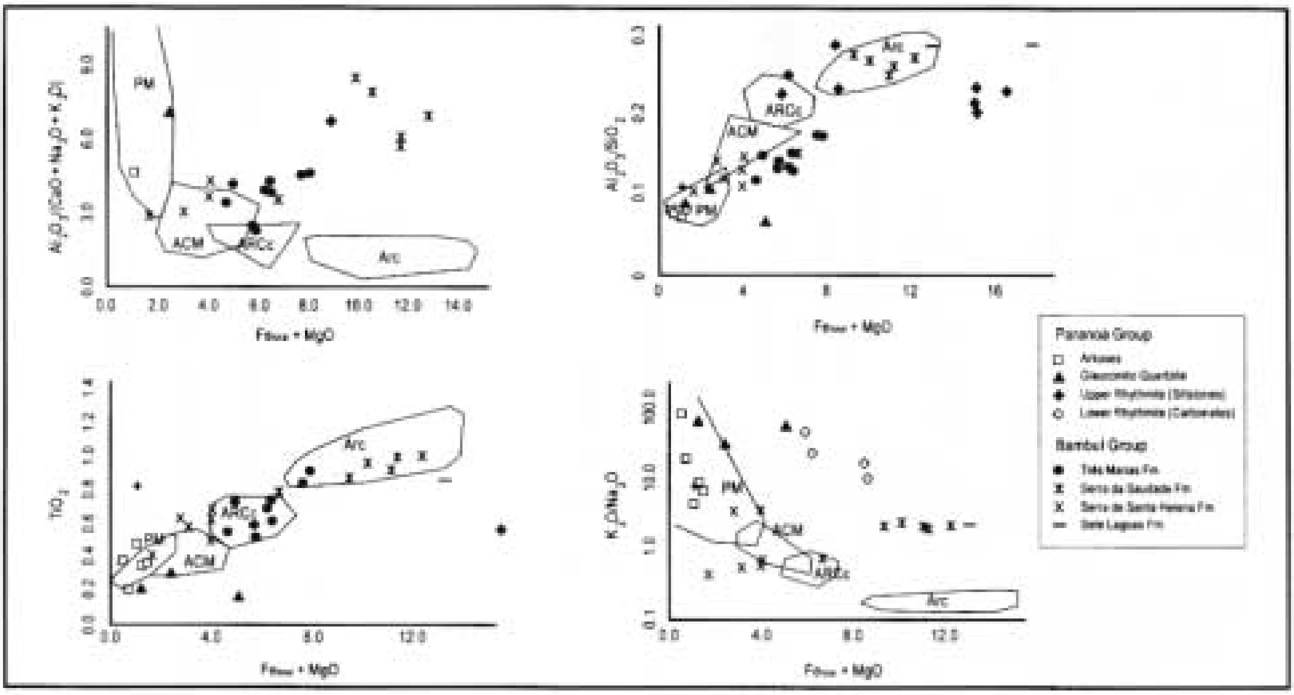


FIGURE 10 - Composition of the Paranoá and Bambuí rocks. PM - Passive Margin; ACM - Active Continental Margin; ARCc - Continental Island Arc; Arc - Oceanic Island Arc (after Guimarães, 1997).

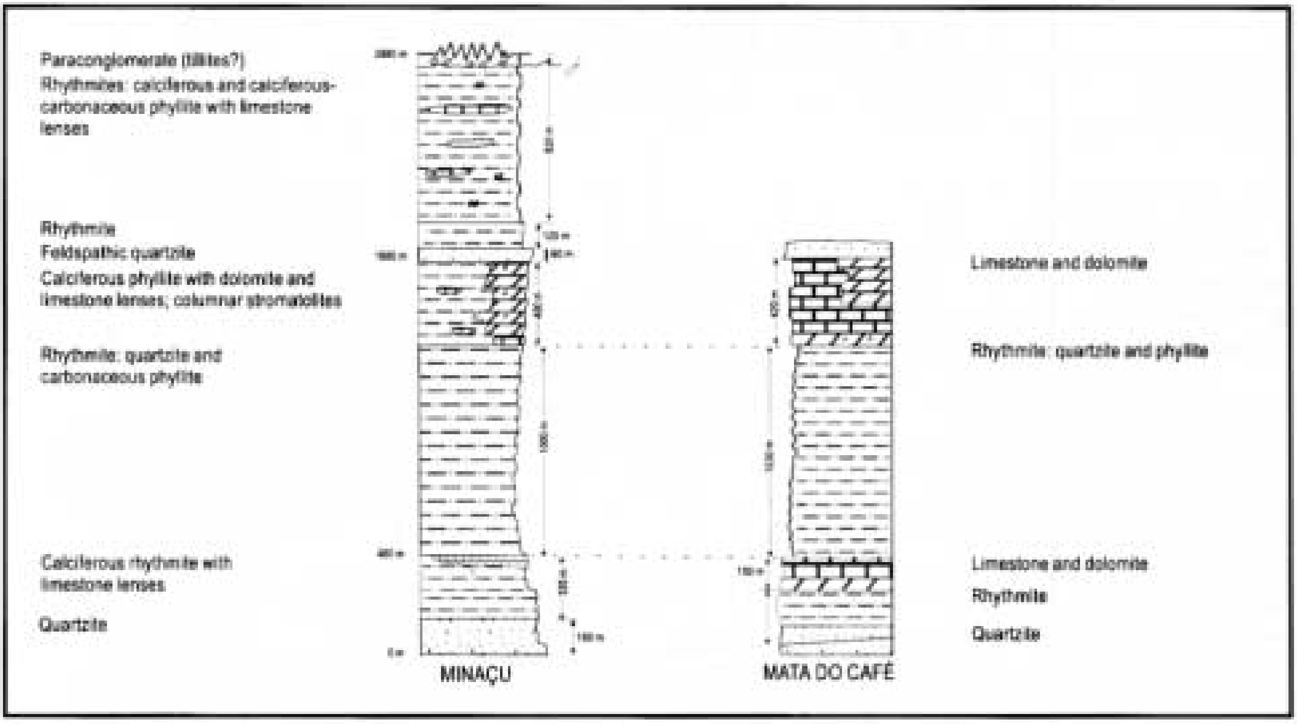


FIGURE 11 - Lithostratigraphic columns of the Minaçu Formation (after Marini and Fuck, 1981).

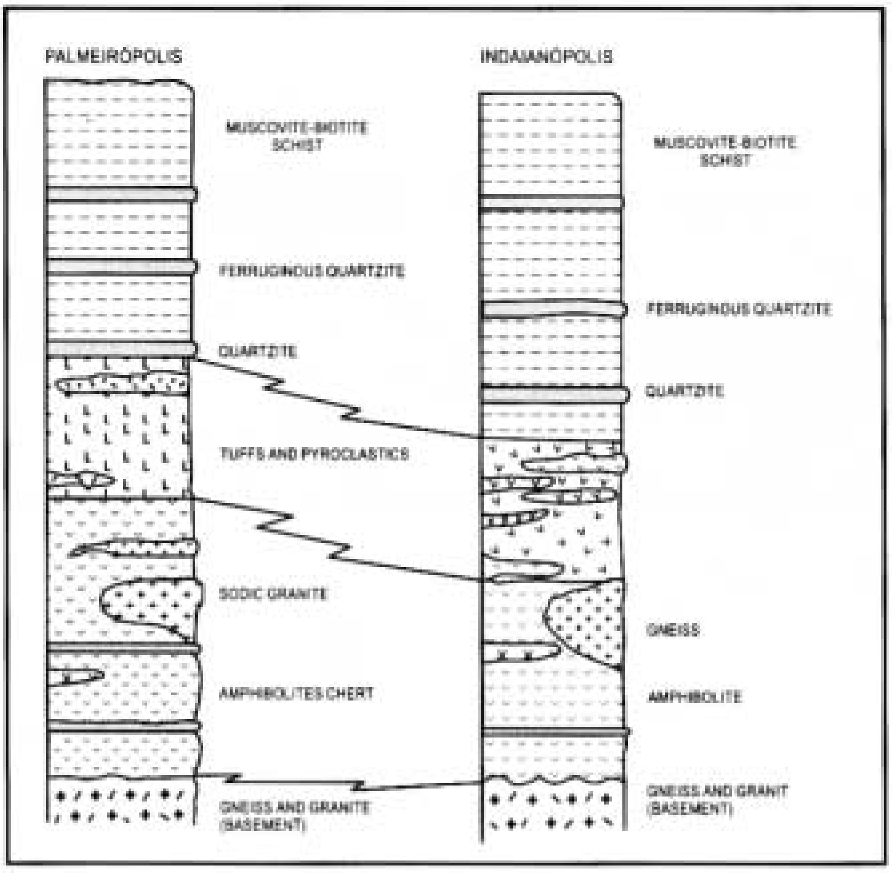
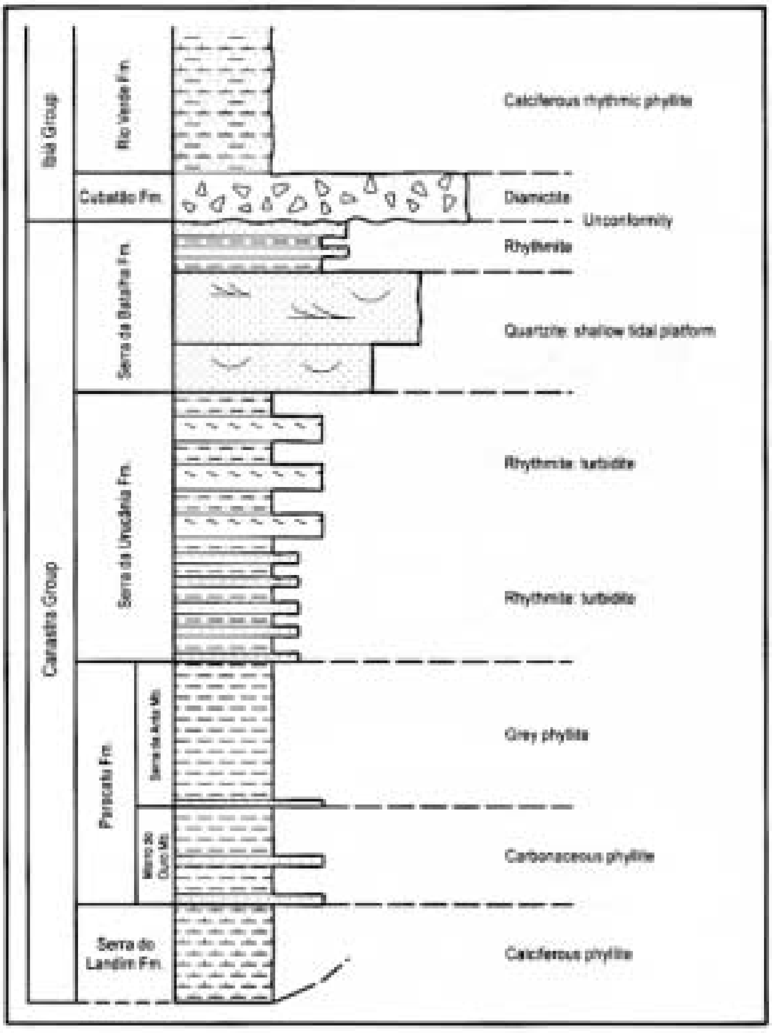


FIGURE 12 - Volcano-sedimentary sequences of Palmeirópolis and Indaianópolis (after Marini et al., 1984).

FIGURE 13 - Lithostratigraphy of Canastra and Ibiá groups in the region of Paracatu - MG (after Freitas-Silva and Dardenne, 1994; Pereira et al., 1994).



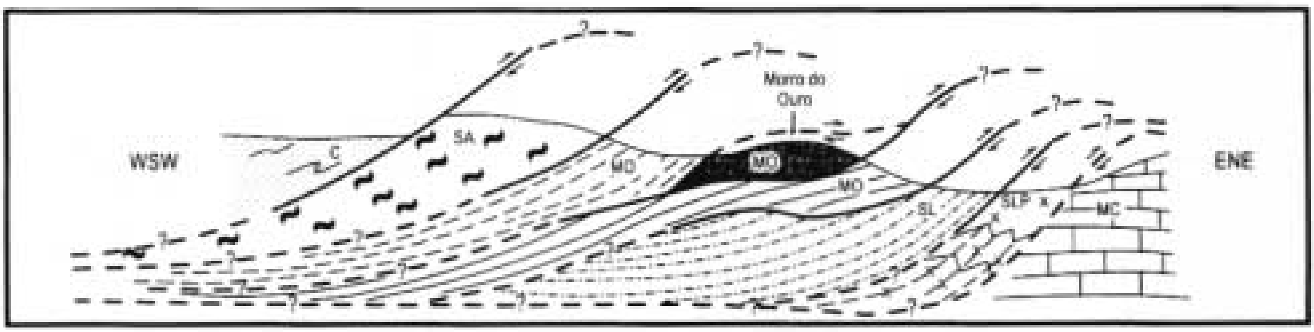


FIGURE 14 - Geological cross-section of the Paracatu region. C - Canastra Group; facies: MC - Morro do Calcário; SLP - Serra da Lapa; SL - Serra do Landim; MO - Morro do Ouro (ore deposit in black); and SA - Serra da Anta (after Freitas-Silva and Dardenne, 1991).

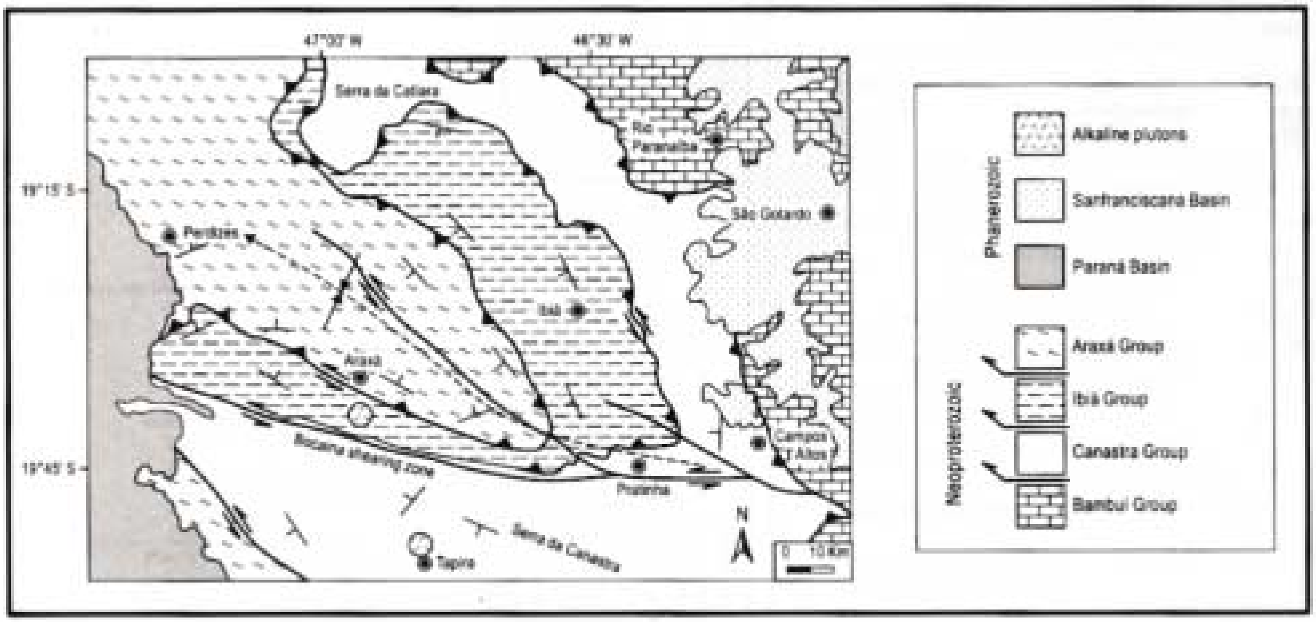


FIGURE 15 - Geologic map of the Araxá Sinform (after Seer, 1999).

occurs in the homonymous region to the W of Paracatu and Vazante, was subdivided in two members: the lower Serra da Urucânia Member consisting of regular intercalations of quartzite and phyllite; and the upper Hidroelétrica Batalha Member consisting mainly of quartzite. The Paracatu and Chapada dos Pilões formations show a coarsening upward succession, suggestive of a regressive megacycle. The base of the megacycle consists of beds rich in organic matter and diagenetic pyrite interpreted to have been deposited in deep water. These beds pass gradually upwards to turbidite beds deposited on a talus by gravity currents. Near the top there are facies typical of those found on platforms dominated by the action of storm currents (hummocky cross stratification). Finally, there are sediments associated with a shallow platform, dominated by tidal currents with cross stratification indicating a transport from E to W.

Generally, the base of the Canastra Group is not observed, being obliterated by the overthrusts that place this group in contact with lower grade metamorphites of the Vazante, Paranoá and Bambuí groups. The relationship between the Canastra Group and the Araxá Group is not clear in function of the tectonic imbrication observed between these sequences, that are generally interpreted as lateral equivalents. In like manner, the Canastra Group is often considered as a more metamorphosed equivalent of the Paranoá Group in the southern segment of the BFB (Campos Neto, 1984a, b; Dardenne, 1978, 1979; Pereira, 1992; Pereira *et al.*, 1994; Pedrosa-Soares *et al.*, 1994).

The carbonaceous phyllite of the Morro do Ouro deposit near the town of Paracatu is a very low-grade (*c.* 0.4 g/t Au), bulk mineable gold deposit, with remaining reserves estimated at >200t gold metal. The annual production in that area is given as around 5.2 t. The ore paragenesis of the gold mineralization, associated with quartz boudins, is pyrite, arsenopyrite, galena, sphalerite, sericite and Fe-carbonate. The main control of the mineralization is structural (Fig. 14), being developed along a low angle shear thrust (Freitas-Silva, 1996; Freitas-Silva *et al.*, 1991).

Araxá Group

The Araxá Group (Fig. 15), defined in the homonymous region by Barbosa (1955), Barbosa *et al.* (1970), occurs mainly in the southern segment of the BFB, being differentiated from the Serra da Mesa Group by Marini *et al.* (1984a, b) in function of the distinct characteristics of the two groups to the N and S of the Pirineus mega-inflection. It consists mainly of micaceous quartzite and micaschist (calc-schist, muscovite-quartz schist, muscovite-clorite schist, biotite-garnet schist, staurolite schist and feldspathic schist). At the base of these schist units paragneiss with biotite and hornblende, frequently described in the literature, may be observed. Lenses of calcitic and/or dolomitic marble occur locally (Fig. 16).

The presence of volcanic rocks associated with the micaschist of the Araxá Group is frequently observed in some areas (Lacerda Filho and Oliveira, 1995). These include the Rio do Peixe, Silvânia (Freitas, 1994), Ipameri (Dardenne *et al.*, 1992), Santa Cruz and Pires do Rio (Leonardos *et al.*, 1990), Abadia dos Dourados (Brod *et al.*, 1992), Araxá (Seer, 1999; Ferrari, 1981, 1989a, b), Passos (Valeriano and Simões, 1997). These volcanic rocks are amphibolite (metabasalt), meta-andesite and meta-rhyolite, which stand is evidence for the volcano-sedimentary character of the Araxá sequence (Fig. 17). The presence of lenses of serpentinite, amphibolite, actinolite and talc schist, with podiform chromite associated, tectonically interbedded in the Araxá micaschist, led Drake Jr. (1980) and later Strieder and Nilson (1992) to consider this assemblage as an ophiolitic melange, obducted over a continental margin by nappes transported from W to E. The occurrence of this volcanism in restricted areas turns evident the appearance of oceanic crust related to discontinuous openings in the southern segment of the BFB. In the Araxá (Seer, 1999), Abadia dos Dourados (Brod *et al.*, 1991) and Ipameri (Dardenne *et al.*, 1992; Pimentel *et al.*, 1995) regions, the volcanic and metasedimentary rocks are intruded by metagabbro, metadiorite and granitoid with a clear syn-tectonic character.

Ibiá Group

This unit, defined as Ibiá Formation by Barbosa *et al.* (1970), was raised to group status by Pereira (1992), Pereira *et al.* (1994), being subdivided in two formations (Fig. 13):

- The Cubatão Formation, at the base, lying on an erosional unconformity or by tectonic contact over the Canastra Group, is represented by thick units of diamictite, interpreted as deposited in glacio-marine conditions by gravity flows (Dardenne *et al.*, 1978; Pereira, 1992; Pereira *et al.*, 1994). The pebbles found in the diamictite beds are extremely varied, and include those of quartzite, limestone, dolomite, schist, gneiss and different types of granitoid.
- The Rio Verde Formation, at the top, is composed essentially of calciferous phyllite or calc-schist with quartzose laminations and some intercalations of fine-grained quartzite and grey phyllite, locally carbonaceous. It is characterized by the presence of chlorite, which gives it an apple-green color. It shows a great uniformity of facies, corresponding to deposition in relatively deep water.

The diamictite beds of the Cubatão Formation have been frequently correlated with the Jequitaí Formation, whereas the rhythmic phyllite units of the Rio Verde Formation were considered as equivalent of the metasediments of the Macaúbas Group and/or the Araxá Group (Dardenne *et al.*, 1978; Pereira, 1992; Pereira *et al.*, 1994; Pedrosa-Soares *et al.*, 1994; Ferrari, 1989a, b). The provenance studies (Seer, 1999) indicate a volcanic arc source (Fig. 18) for the sediments of the Ibiá Group, which is reflected by the Sm/Nd model ages (1.1 Ga) obtained for these sediments (Seer, 1999; Pimentel *et al.*, 1999).

Vazante Group

In the northwestern part of the State of Minas Gerais, the metasediments of the Vazante Group cover an area some 250 km long, the general orientation of which is N-S, passing near the towns of Coromandel, Lagamar, Vazante, Paracatu and Unaí (Fig. 19). These metasediments comprise a thick marine pelitic-dolomitic sequence that can be divided into seven formations (Fig. 20), from base to top: Retiro, Rocinha, Lagamar, Serra do Garrote, Serra do Poço Verde, Morro do Calcário and Serra da Lapa formations (Dardenne *et al.*, 1997, 1998).

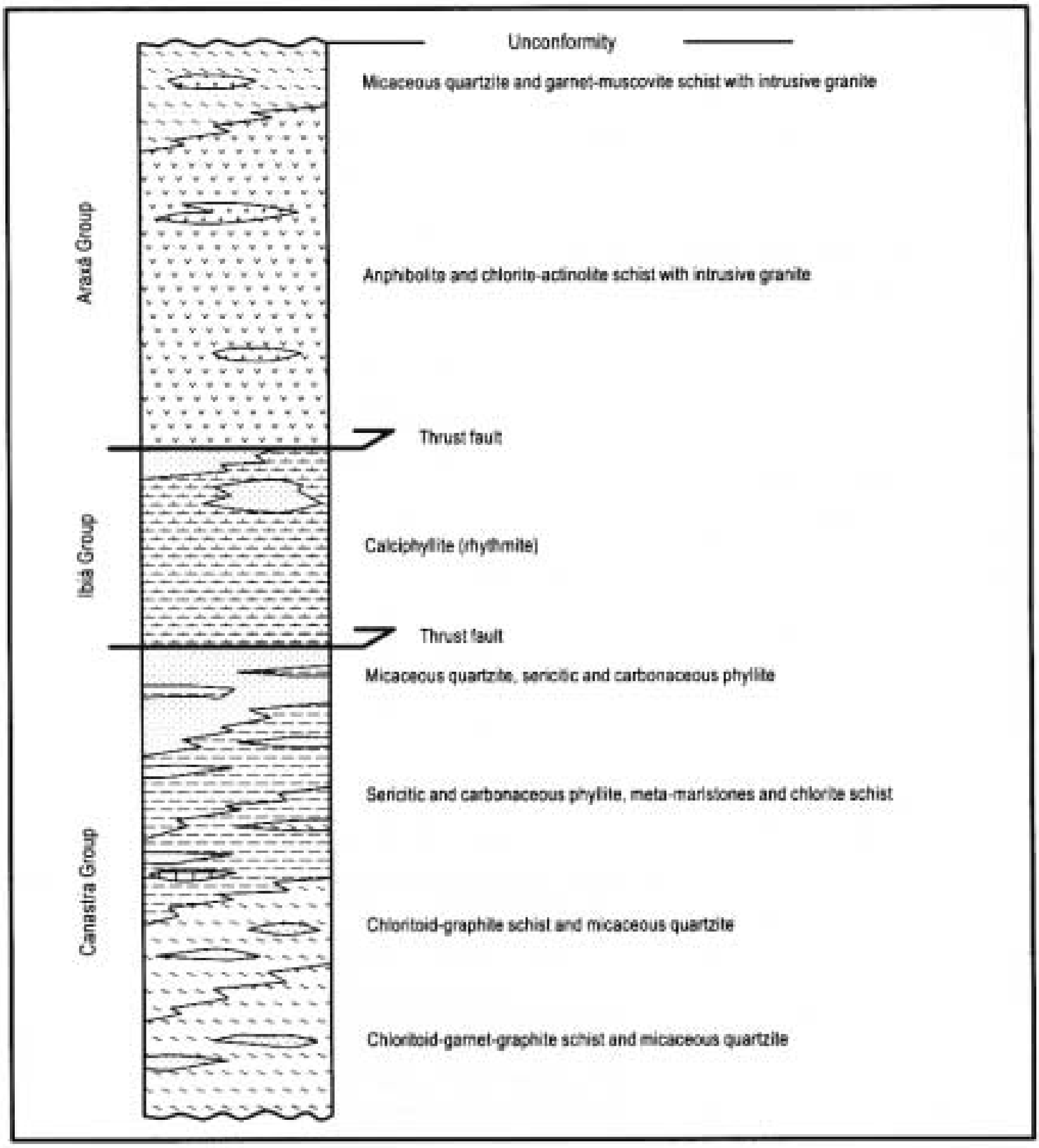


FIGURE 16 - Tectonostratigraphy of the Araxá Sinform (after Seer, 1999).

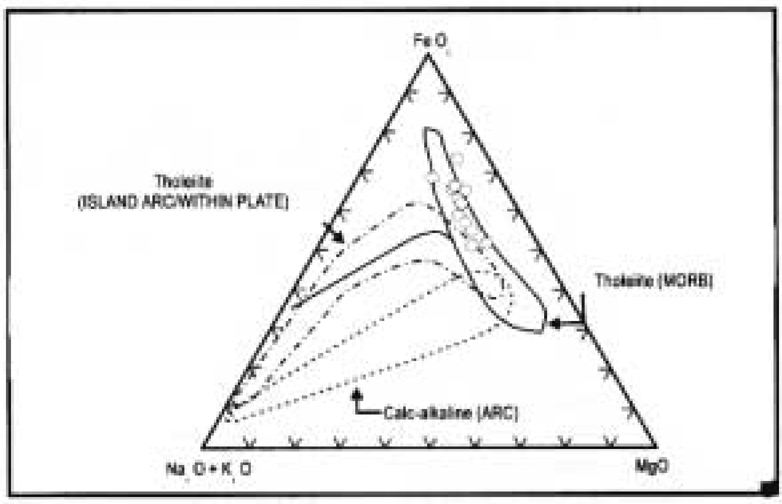


FIGURE 17 - AFM diagram showing the tholeiitic trend for the Araxá Anfibolite (open circles) (after Seer, 1999).

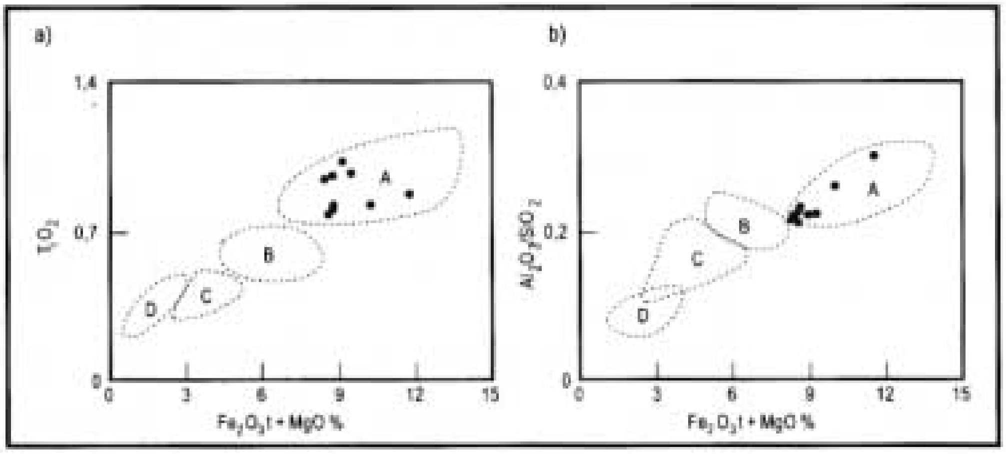


FIGURE 18 - Diagrams TiO_2 x Fe_2O_3 + MgO (a) e Al_2O_3/SiO_2 x Fe_2O_3 + MgO (b) for geotectonic environment of the Ibiá Group (black circles) (after Seer, 1999). A - oceanic island arcs; B - continental island arcs; C - active continental arcs and D - passive margins.

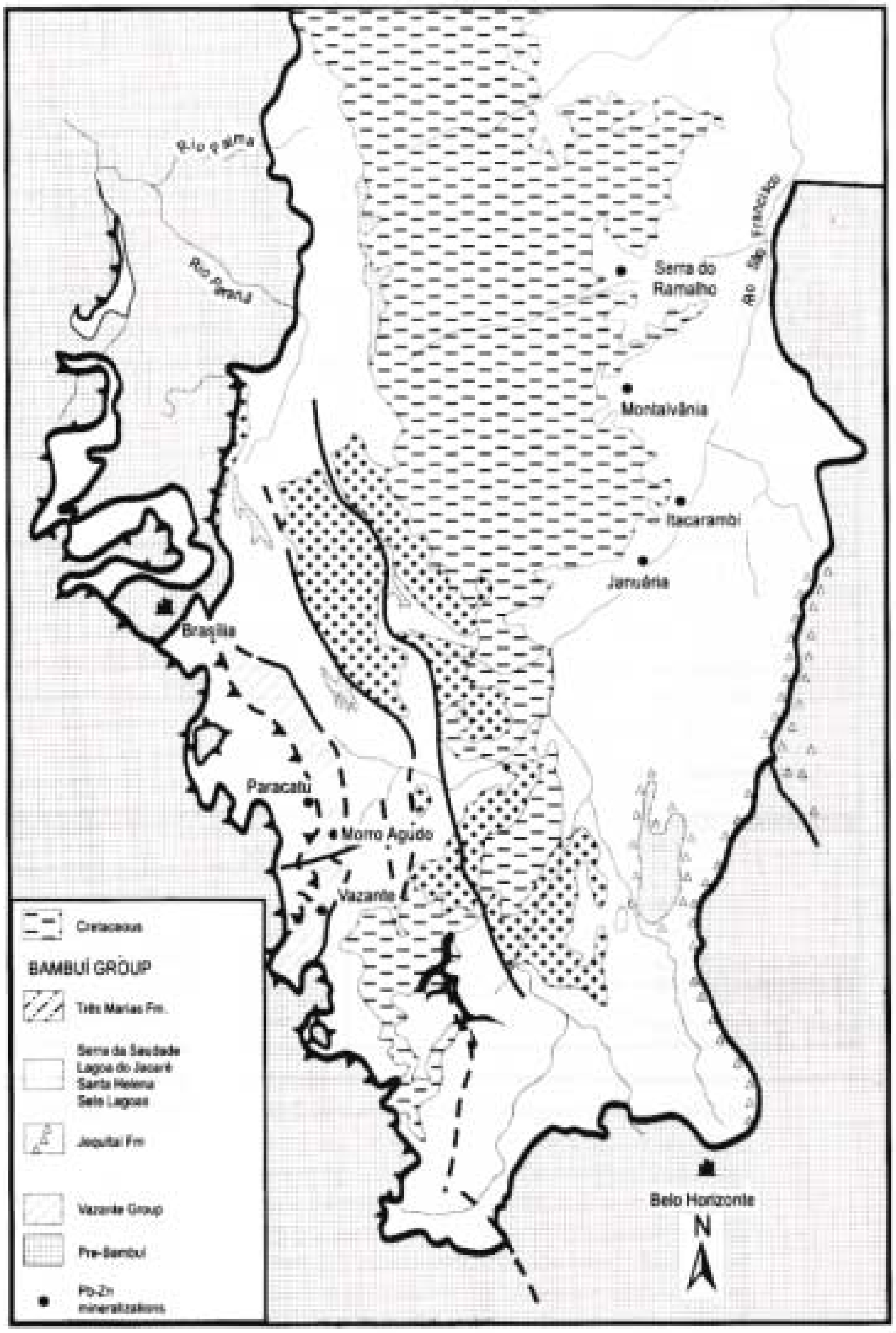


FIGURE 19 - Situation of the Pb-Zn deposits in Vazante and Bambuí groups (after Dardenne, 1978).

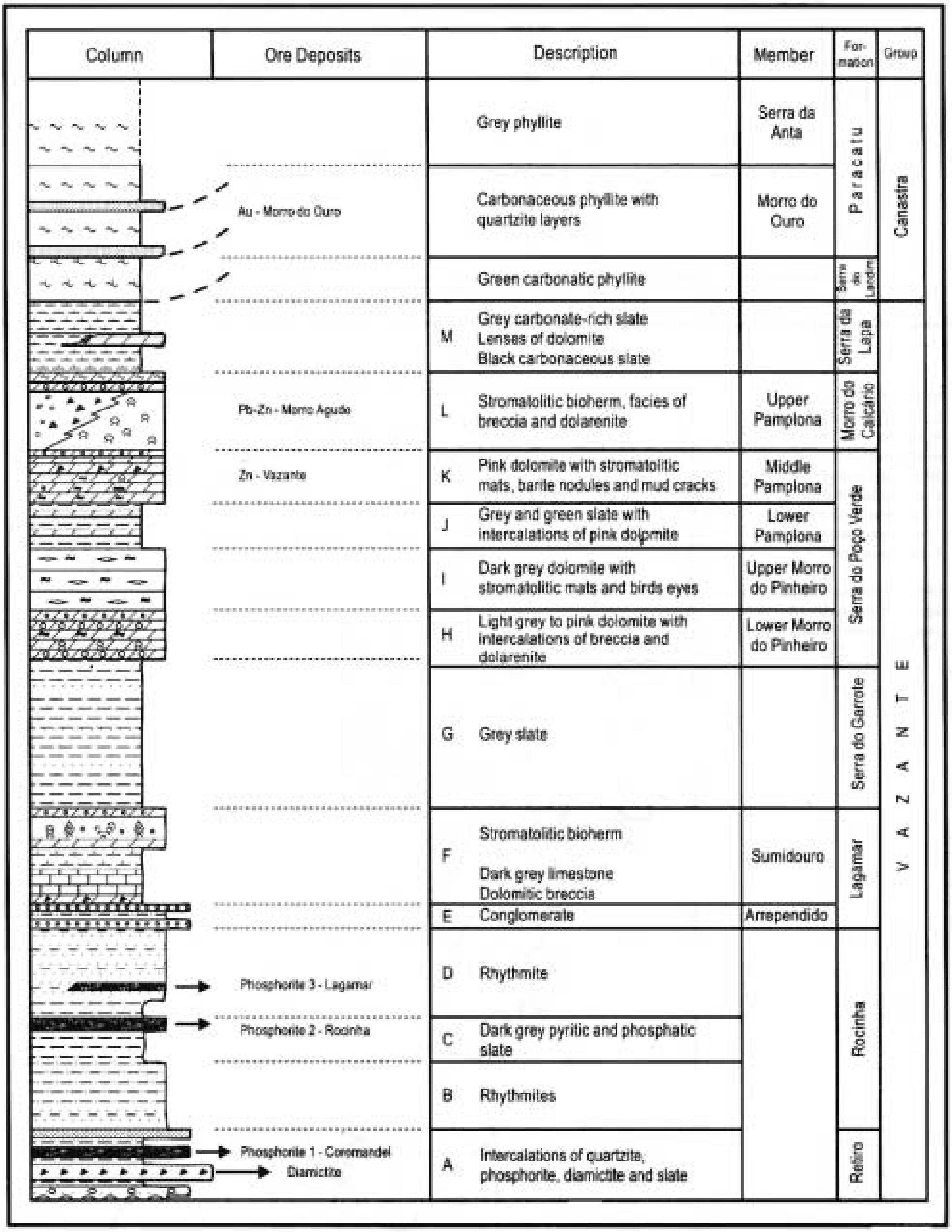


FIGURE 20 - Lithostratigraphic column of the Vazante Group (after Dardenne et al., 1998).

• Retiro Formation: this is considered to be the basal formation and consists of metric beds of white quartzite, locally conglomeratic, intercalated with slate. In the Santo Antônio do Bonito and Santo Inácio rivers, this formation is characterized by the presence of diamictite beds with pebbles of quartzite, limestone, dolomite, metasiltstone and granitoid in the pelitic matrix, which is locally phosphatic. Larger concentrations of phosphate are found in the slaty facies and in the phospharenite layers, rich in intraclasts and pellets (Phosphorite 1). The diamictite beds represent debris flows deposited in relatively deep water by gravity currents (Dardenne *et al.*, 1998; Souza, 1997).

• Rocinha Formation: at the base this formation consists of a rhythmic sandy and pelitic sequence that grades upwards to the Retiro Formation. In its upper part, it consists of a thick sequence of slates and metasiltstone beds regularly intercalated with yellow to red alteration colours. It passes vertically to dark grey, carbonate and pyrite-bearing slate, with fine phosphatic laminations that slowly change to intraclasts and pellet rich phospharenite (Phosphorite 2), forming the Rocinha phosphate deposit, which has reserves in the order of 400 Mt at 12% P_2O_5. In the upper part of the formation, rhythmic sediments (quartzite and siltstone) host the Lagamar phosphate deposit (Nogueira, 1993; Dardenne *et al.*, 1997) composed essentially by phospharenite (Phosphorite 3). Reserves of the deposit have been estimated at 5 Mt at 25% P_2O_5.

• Lagamar Formation: the psammo-pelitic carbonate unit of Lagamar (Dardenne *et al.*, 1976; Dardenne, 1978a, b, 1979; Campos Neto, 1984a, b; Dardenne *et al.*, 1997, 1998) is represented in its basal part by alternating beds of conglomerate, quartzite, metasiltstone and slate. The conglomerate units show a framework supported by the quartzite, metasiltstone and dark grey limestone clasts and are known as the Arrependido Member. These psammite beds are overlain by dolomitic intraformational breccia passing to dark grey, well-stratified limestone units with intercalations of lamellar breccia followed by stromatolitic dolomite. The stromatolitic beds form very beautiful beige to pale pink bioherms, composed of laminated dolomite (cyanobacteria mats), oncolitic dolarenite and dolorudite, and columnar stromatolites with convex and conical laminations of the *Conophyton* and *Jacutophyton* type. Laterally and vertically, these bioherms interdigitate with carbonate-bearing metasiltstone beds and slate.

• Serra do Garrote Formation: this formation consists of a thick sequence of dark grey to greenish grey slate, locally rhythmic, carbonaceous and containing pyrite, with fine quartzite intercalations (Madalosso and Valle, 1978; Madalosso, 1980; Dardenne, 1978; Campos Neto, 1984; Dardenne *et al.*, 1997, 1998).

• Serra do Poço Verde Formation: this formation corresponds to a dominantly dolomitic sequence, first defined by Dardenne (1978a, b, 1979) and subsequently assigned to the Vazante Formation by Rigobello *et al.* (1988). It is divided into four members described successively from the base to the top:

– Lower Morro do Pinheiro Member: this member consists of light grey and/or pink laminated dolomite with cyanobacteria mats, intercalated with levels of oncolitic dolarenite and intraformational breccia associated with lenses of dolomite with columnar stromatolites (thickness of about 500m).

– Upper Morro do Pinheiro Member: this member consists of medium to dark grey laminated dolomite with cyanobacteria mats and frequent birdseyes features, intercalated with some layers of dolarenite, lamellar breccia and carbonaceous shale (thickness ranging from 300 to 500m).

– Lower Pamplona Member: this member consists of grey, green and purple siltstone beds intercalated with pink micritic laminated dolomite with cyanobacteria mats and small lenses of fine-grained to conglomeratic sandstone (thickness between 100 and 200 m).

– Medium Pamplona Member: this member consists of light grey to pink dolomite with cyanobacteria mats laminations, intercalated with layers of dolarenite, lamellar breccia and dolomite with columnar stromatolites, and with shale lenses (thickness around 400m).

• Morro do Calcário Formation: this formation corresponds to the Upper Pamplona Member of Rigobello *et al.* (1988). It is characterized by the presence of pink stromatolitic dolomite forming biostromes and bioherms with convex lamination columns, associated with oolitic and oncolitic dolarenite and dolorudite (thickness of about 200 to 300m). In the region of Morro Agudo, Paracatu and Unai, the Morro do Calcário Formation exceeds 900 m of thickness, being composed essentially of dolorudite, indicating the reworking of stromatolitic bioherms partially preserved and associated with oolitic and oncolitic dolarenite facies. This unusual thickness suggests that, in this region, the Morro do Calcário and Serra do Poço Verde formations represent a single dolomitic sequence deposited continuously, not permitting the subdivision defined in the region of Vazante.

• Serra da Lapa Formation: this formation constitutes the upper part of the Vazante Group, which has been described by Madalosso and Valle (1978) and Madalosso (1980) in the region of Paracatu, as the Serra do Velosinho and Serra da Lapa members. These are represented by carbonaceous phyllite, carbonate-bearing metasiltstone, dolomite lenses and quartzite layers. The dolomitic lenses show various facies as laminated dolomite with cyanobacteria mats, dolomite with columnar stromatolites and dolomite with intraformational breccia, interdigitating with the pelitic sequence that covers regionally the dominantly dolomitic formations of Morro do Calcário and Serra do Poço Verde.

In the Unaí region, this formation appears as a very peculiar facies, consisting of lithic sandstone and conglomerates intercalated with dark grey slate. The clasts are fragments of phyllite, fine quartzite and chert (Laranjeira, 1992; Laranjeira and Dardenne, 1993; Dardenne *et al.*, 1998). The age of the Vazante Group is not yet known. The information obtained from the columnar *Conophyton* type stromatolites indicate a very long interval between 1.35 Ga and 950 Ma (Cloud and Dardenne, 1973; Moeri, 1972; Dardenne *et al.*, 1976), suggesting a correlation with the Paranoá Group. On the other hand, the occurrence of diamictite units at the base of the sequence, very similar to those found in the Jequitaí Formation favours a correlation with the Bambuí Group. The absence of volcanism associated with the sedimentary sequence restricts the quality of the geochronological data which remain restricted

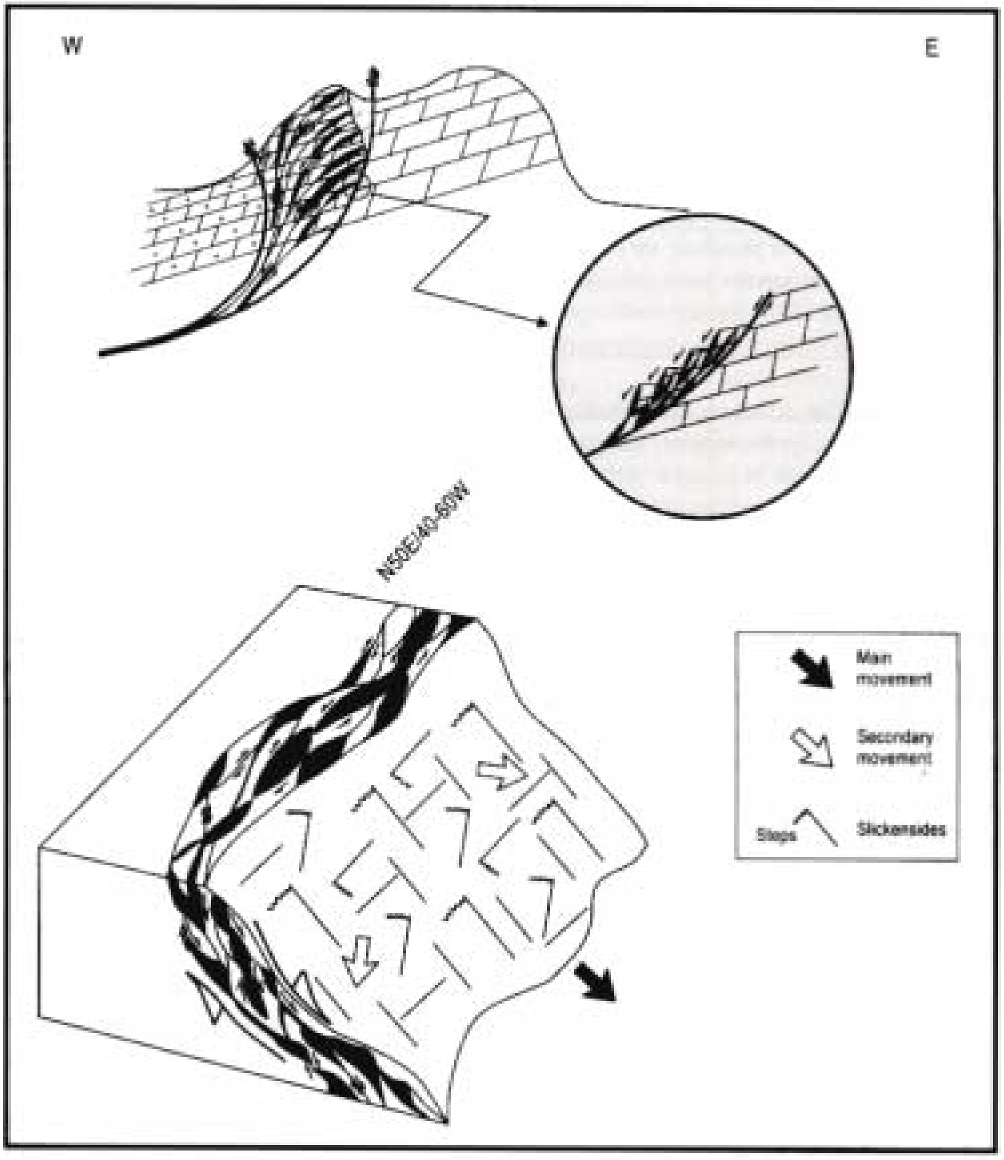
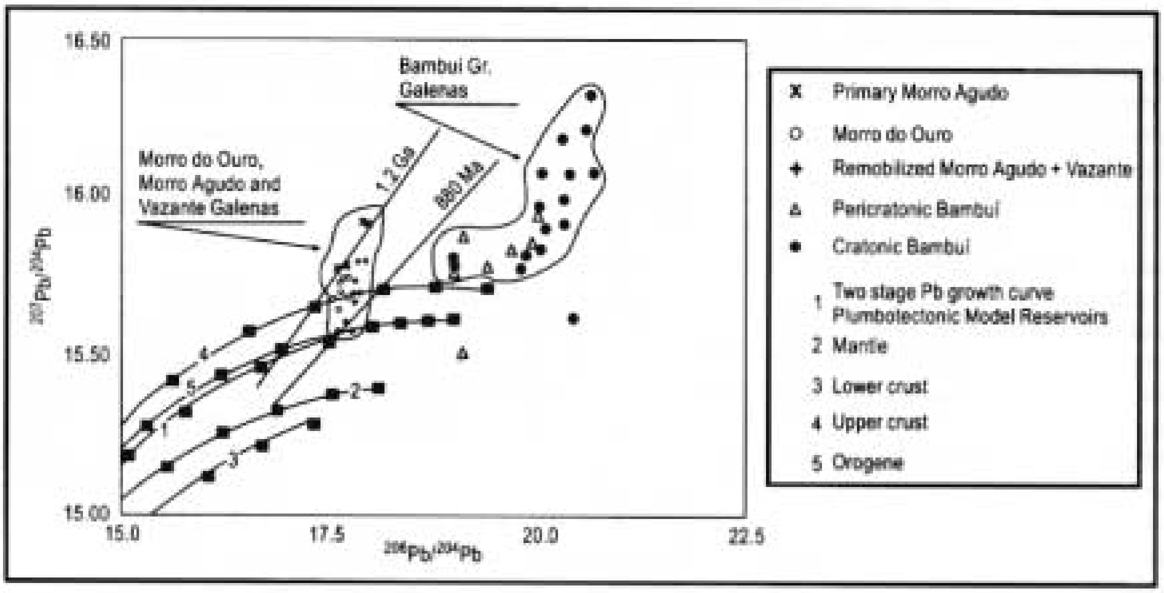


FIGURE 21 - Lenticular aspect of the zinc ore in the listric fault of Vazante (after Dardenne and Freitas-Silva, 1998).

FIGURE 22 - Pattern of Morro do Ouro, Vazante, Morro Agudo and Bambuí galenas in relation to Plumbotectonic Model reservoirs and the two stages of the Pb growth curve intervals in the Plumbotectonic Model = 400 Ma and stage = 250 Ma (after Freitas-Silva and Dardenne, 1997; Dardenne and Freitas-Silva, 1998).



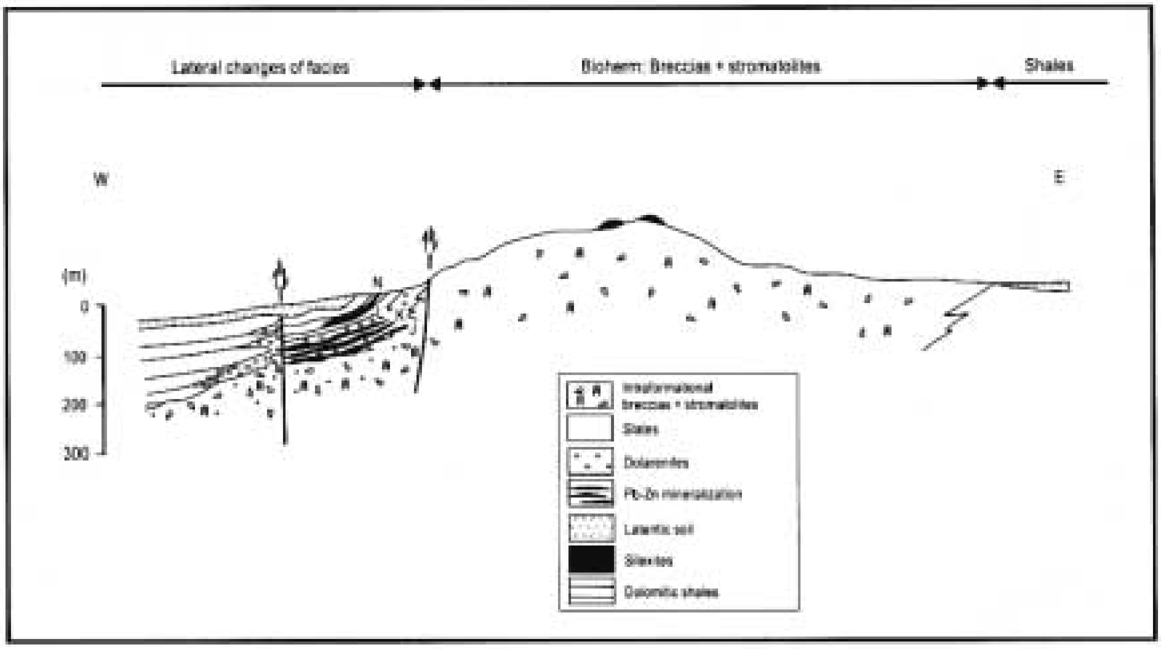


FIGURE 23 - Schematic geological section of Morro Agudo lead-zinc deposit (after Dardenne, 1978; Dardenne et al., 1998).

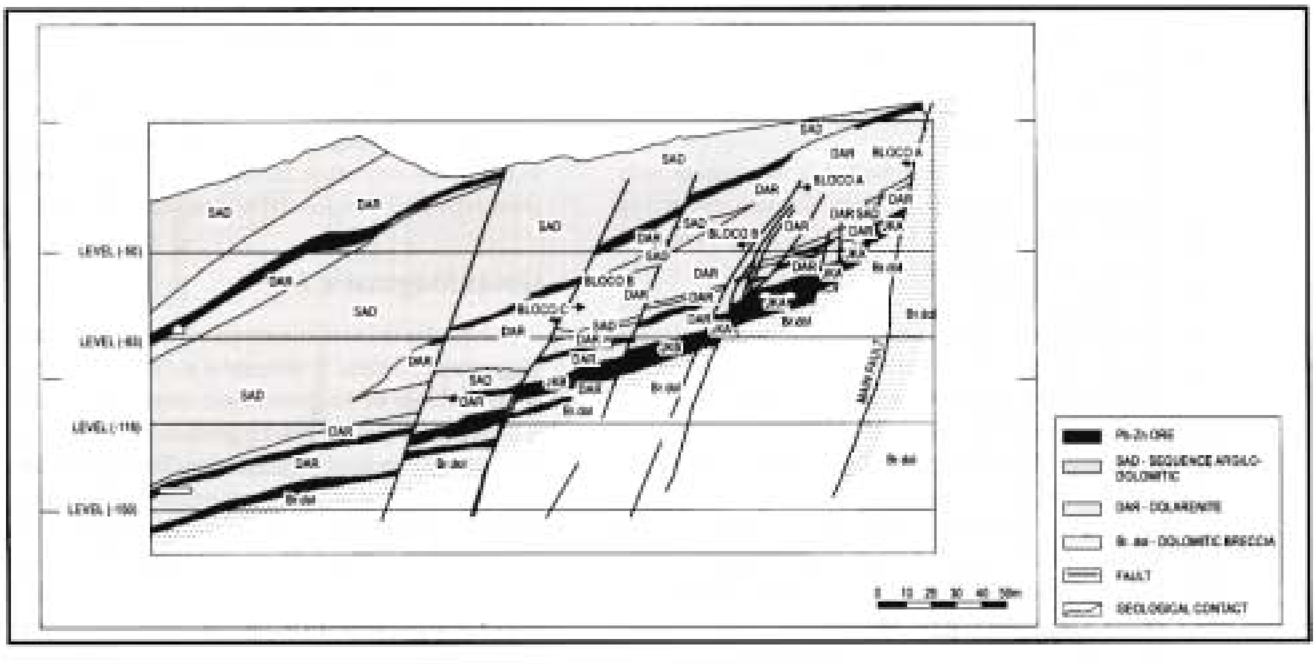


FIGURE 24 - E-W cross-section of the Morro Agudo Pb-Zn ore deposit (after Oliveira, 1998).

to K/Ar and Rb/Sr data on the pelitic sediments, indicating the age of the metamorphism (*c.* 600 Ma), and the Pb/Pb dating on galena of the Vazante and Morro Agudo deposits, which give ages varying between 1.2 Ga and 650 Ma (Freitas-Silva and Dardenne, 1997). In the case of a correlation with the Bambuí Group, the oldest ages could indicate the moment of the separation of the Pb and not the age of the formation of the galena and, consequently, of the sedimentation. In this context, the sedimentary sequence of the Vazante Group would represent an anomalous sedimentation in a rapidly subsiding zone forming a depression in front of the uplift of the BFB.

The most important deposits of Pb and Zn in Brazil (the Vazante and Morro Agudo mines) are hosted in the dolomitic rocks of the Vazante Group. Other known occurrences (*e.g.*,Biboca, Ambrósio, Fagundes) are being evaluated. The Vazante Deposit (Dardenne, 1979; Rigobello *et al.*, 1988; Oliveira, 1998) is associated with an important tectonic structure, the Vazante Fault. The Vazante Fault is some 12 km long, with strike N50°E and dip 60°- 70°NW. It has been interpreted as originally being a syn-sedimentary growth fault, many times reactivated during the Brasiliano tectonic event (Fig. 21). Initially it was an inverse and transcurrent fault during the compressive phase, and laterally a normal fault at the end of the Brasiliano Cycle (Dardenne, 1979; Pinho *et al.*, 1990; Dardenne and Freitas-Silva, 1998a, b). The mineralization is essentially composed of willemite (Zn_2SiO_4) associated with hematite, zincite, franklinite, sphalerite and galena (Monteiro, 1997; Monteiro *et al.*, 1996). It is accompanied by intense hydrothermal alteration consisting of silicification and sideritization (siderite/ankerite) of the dolomite host, while the open fractures and veinlets in the dolomite are filled with siderite/ankerite and red jasper.

This mineralization is hosted in the siltstone beds intercalated with thin layers of pink dolomite corresponding to the Lower Pamplona Member. The analyses of fluid inclusions indicate an aqueous hydrothermal fluid with medium salinity (3 to 15% wt eq. NaCl) and homogenization temperatures between 65 and 180 °C. The Pb isotopes in the galena (Fig. 21) show a constant $^{206}Pb/^{204}Pb$ ratio, around 17.7, similar to those found in Morro Agudo (Bez, 1980; Freitas-Silva, 1996; Iyer *et al.*, 1992; Freitas-Silva and Dardenne, 1997; Dardenne and Freitas-Silva, 1998a, b). $\delta^{34}S$ of the sulfides show positive values varying between + 12 and + 14‰. $\delta^{18}O$ values are negative (- 9 to 0‰) for the carbonate of the mineralized zone (Monteiro, 1997) suggesting a mixture of hydrothermal fluids with meteoric water. These data suggest hydrothermal mineralization formed by the migration of connate fluids from basin sediments to the Vazante growth fault, which was subsequently submitted to the intense shearing during the Brasiliano event.

The Morro Agudo Deposit (Dardenne, 1978a, b, 1979; Romagna and Costa, 1988; Oliveira, 1998) is associated with back-reef facies situated on the western flank of the stromatolitic bioherm of the Morro do Calcário (figs. 23 and 24). Breccia, dolarenitic breccia and dolarenite constitute the host rocks of the main disseminated sulfide levels (sphalerite and galena), denominated I, J, K and L. The M level has a strata-bound character with mineralized fractures regularly distributed inside the same dolomitic zone. The N level mineralization is stratiform, and consists of regular and alternate laminations of chert, galena, sphalerite and pyrite.

The mineralized facies are limited by a N10°W normal fault that was active during the sedimentation and borders the stromatolitic bioherm on the western side. This fault, which may represent a preferential conduct to the mineralizing fluids, is filled in its upper part by massive vein mineralization of galena associated to sphalerite, pyrite and barite, resulting from the remobilization of the primary mineralization during the Brasiliano tectonic event. The disseminated mineralization shows patterns of diagenetic substitution of the dolomitic material by sphalerite and galena from the syn-diagenetic stage when the sediment was not consolidated until tardi-diagenetic and epigenetic stages (Dardenne, 1978a, b, 1979). The galena has a very low Ag content, whereas the sphalerite, of pale yellow color, is rich in Cd (300 ppm). At the base of the J and K levels, colloform structures of sphalerite are observed.

The fluid inclusion studies developed on the sphalerite show an aqueous hydrothermal fluid with variable salinity from 0 to 18% wt eq. NaCl, and homogenization temperatures varying between 70 and 160 °C (Dardenne and Freitas-Silva, 1998a, b). The lowest values are observed for the stratiform sulfides of the N level (- 0,5%), showing the reduction of sulfides by bacterial activity. The Pb isotopes in the galena show constant values for the $^{206}Pb/^{204}Pb$ (*c.* 17.7) and $^{207}Pb/^{204}Pb$ (*c.* 15.3) ratios, corresponding to a model age of 1.1 to 1.0 Ga for the syn-diagenetic galena (Freitas-Silva and Dardenne, 1997). These data suggest that the Morro Agudo Deposit is diagenetic, originated by continuous expulsion of the connate fluids from the basin sediments. In this way, the Morro Agudo Deposit is similar to the Navan Deposit in Ireland that shows the same characteristics (Hitzmann, 1995; Hitzmann *et al.*, 1995).

Goiás Magmatic Arc

Situated in the westernmost part of the BFB, the Goiás Magmatic Arc, defined by Pimentel *et al.* (1991), consists of an assemblage of Neoproterozoic meta-igneous and metasedimentary rocks showing geochemical and isotopic characteristics similar to the associations formed in modern island arcs and in active margin environments. It consists, therefore, of a segment of the juvenile continental crust in the central region of Brazil. This arc is represented by narrow belts of volcanic and metasedimentary rocks with structural strike varying between NNW-SSE and NNE-SSW, separated by orthogneiss of dioritic to granitic composition, frequently mylonitized. The available isotope data indicate that the igneous protoliths of these rocks were formed between 900 Ma and 640 Ma.

The Mara Rosa (Arantes *et al.*, 1991), Arenópolis and Bom Jardim de Goiás volcano-sedimentary sequences were formed in an oceanic island arc environment that was intruded by tonalite and granodiorite, with which diverse gold and base metals deposits are associated. The volcano-sedimentary sequence of Mara Rosa, formed at *c.* 860 Ma, deformed and metamorphosed between 790 and 630 Ma (Richardson *et al.*, 1986), contains several Au and Cu-Au

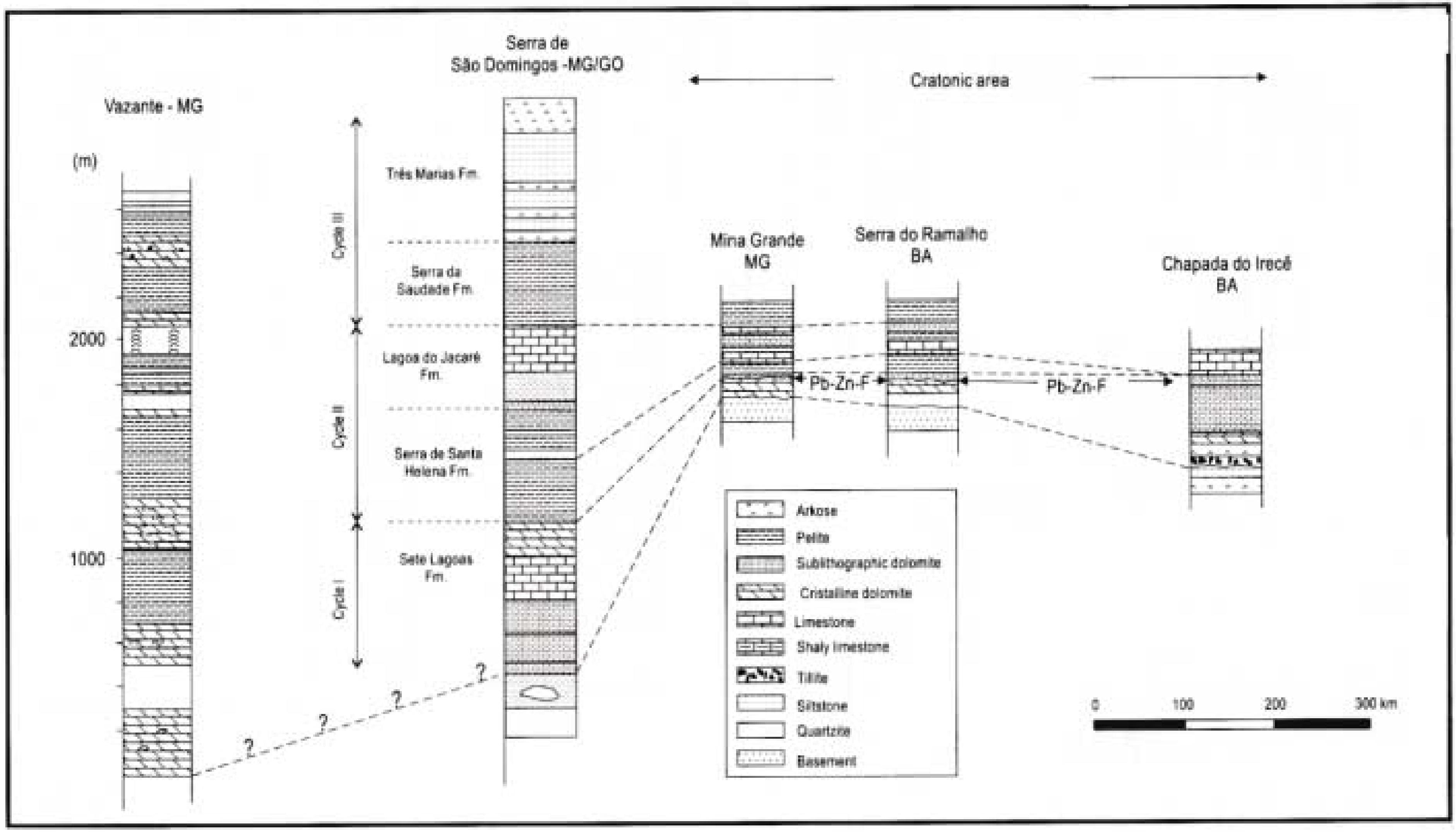


FIGURE 25 - Thickness variation of the Bambuí Group and localization of Pb-Zn-F mineralizations in the cratonic area (after Dardenne, 1978).

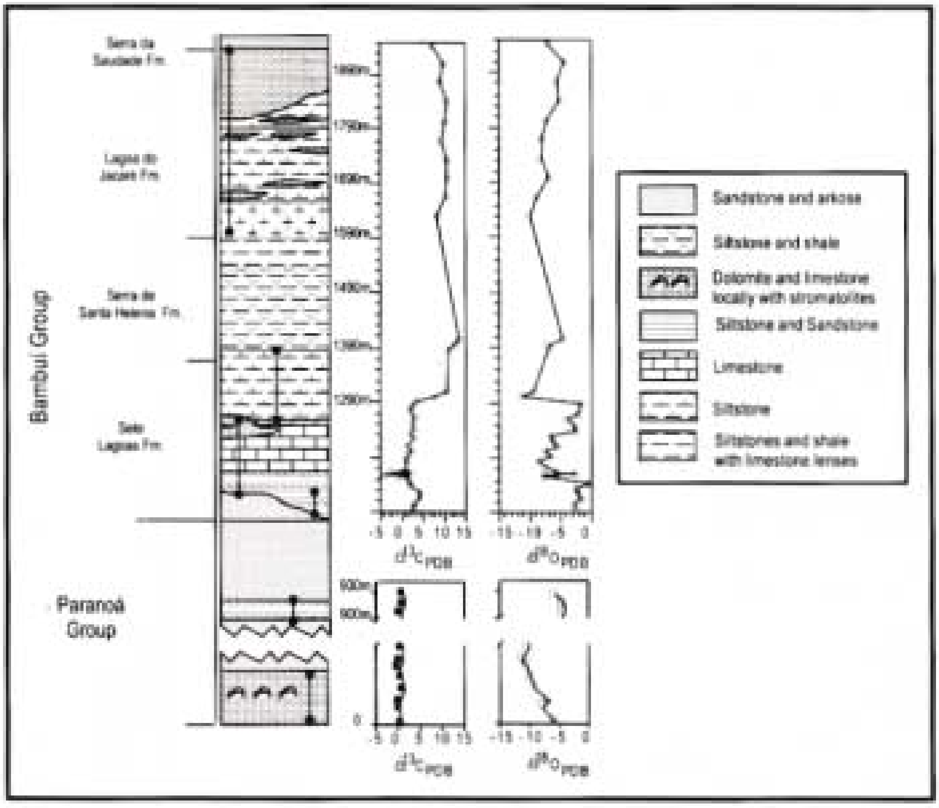


FIGURE 26 - C and O isotopic variations of the Bambuí Group in the Serra de São Domingos, MG (after Santos et al., 1996).

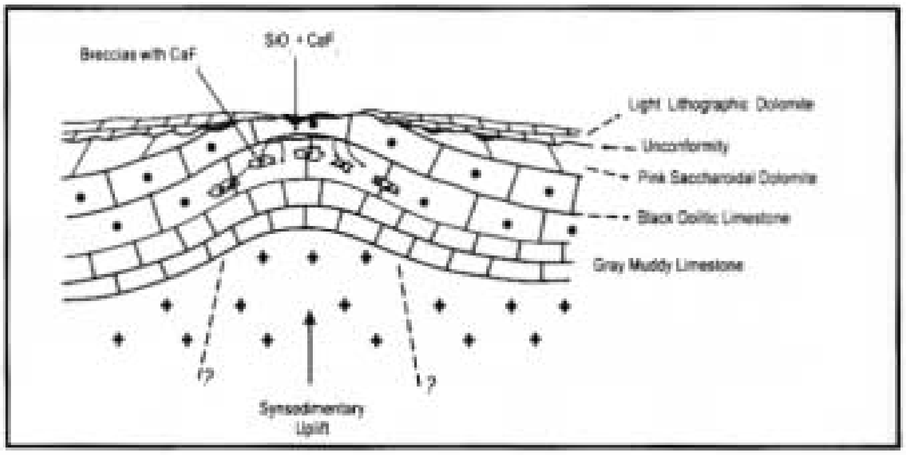


FIGURE 27a - CaF_2 deposits in Campo Alegre (Serra do Ramalho - BA).

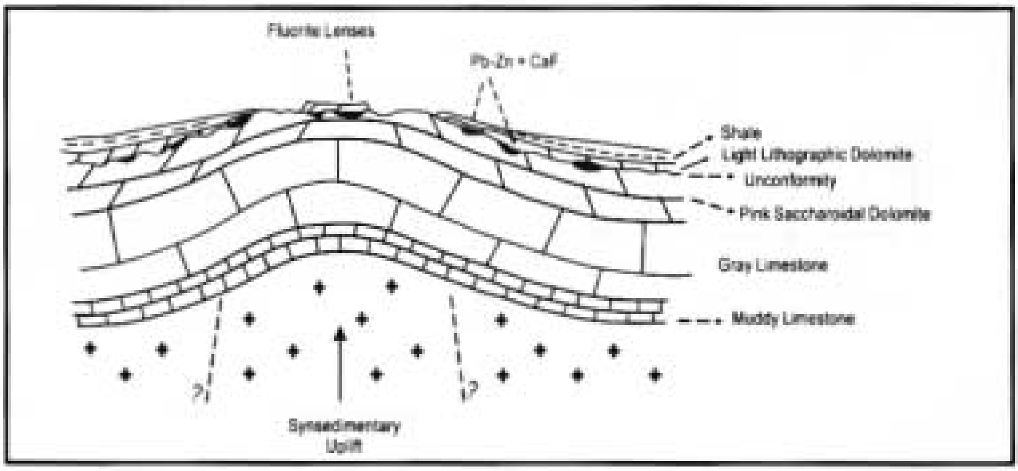


FIGURE 27b - Pb-Zn-CaF_2 deposits in Mina Grande (Itacarambi - MG).

deposits: the Chapada (Cu-Au) Deposit, interpreted as an exhalative volcanogenic deposit (Kuyumjian, 1995) or a porphyry-copper type (Richardson *et al.*, 1986); Posse (Au) Deposit, epigenetic, associated with hydrothermalism in shear zones (Arantes *et al.*, 1991a, b); Zacarias (Au-Ag-Ba) Deposit of the VMS-type of exhalative origin (Arantes *et al.*, 1991a, b). In this region, Oliveira *et al.* (1997) underline the importance of the hydrothermalism associated with shear zones for the evolution of the gold mineralization. In the Bom Jardim volcano-sedimentary sequence, the gold mineralization is considered as exhalative, being remobilized in transcurrent fault zones (Seer, 1995).

Bambuí Group

Originally suggested by Rimann (1917), the name Bambuí has been widely used in the Brazilian geological literature, and is applied to the pelitic and carbonate sediments that belong to the Neoproterozoic, occupying all the eastern side of the BFB and covering large areas of the SFC. The lithostratigraphy, originally described by Branco and Costa (1961), was detailed with the research developed by Oliveira (1967), Braun (1968), Barbosa *et al.* (1969, 1970), Schöll (1973) and Dardenne (1978a, b, 1979). Presently the Bambuí Group is divided into six formations (Fig. 25): Jequitaí, Sete Lagoas, Serra de Santa Helena, Lagoa do Jacaré, Serra da Saudade and Três Marias. This lithostratigraphic sequence has been recognized in the states of Goiás, Minas Gerais and Bahia. The group covers unconformably the granite-gneiss basement, and the metasediments of the Araí and Paranoá groups.

The Jequitaí Formation, at the base of the Bambuí Group, represents the expression of a glacial episode that occurred over a large part of the SFC, marked by the presence of diamictite beds containing pebbles of limestone, dolomite, quartzite, gneiss and diverse types of granitoid in a matrix that is frequently greenish-grey in colour and contains carbonate. The melting of the ice at the end of this glacial period provided the necessary conditions for the installation of a marine environment and the beginning of the deposition of the carbonate-bearing pelitic sediments over the stable area of the São Francisco Craton (Dardenne, 1978a, b). This sedimentary association, which followed the Jequitaí glaciation, is represented by three regressive megacycles (Fig. 25). Each of these megacycles begins with a rapid marine transgression of regional extent, associated with a sudden subsidence of the basin as is shown by the deep-water pelitic marine beds that pass to shallow platform facies and finally to tidal and supratidal beds.

From the base to the top, these megacycles are as follows:

- Megacycle I: pelitic-carbonate sediments, corresponding to the Sete Lagoas Formation, showing a coarsening upward sequence with dark grey to black calcilutite at the base, passing to limestone and dolomite at the top.
- Megacycle II: pelitic-carbonate sediments, constituting the Serra de Santa Helena Formation, essentially pelitic, indicating a sudden and generalized subsidence of the basin, followed by the Lagoa do Jacaré Formation, consisting of dark grey limestone deposited on a platform dominated by storms and tidal currents.
- Megacycle III: pelitic-sandy sediments, represented by the Serra da Saudade Formation, pelitic beds deposited on a platform lying at a mean depth that was periodically subject to the influence of storms; and by the Três Marias Formation, consisting mainly of arkose deposited on shallow platform dominated by storm currents with tidal to supratidal facies occurring locally (Chiavegatto, 1992).

The lithostratigraphic sequences from the Januária-Itacarambi cratonic area towards the Vazante region in the W show a considerable increase in the thickness of the carbonate and pelitic beds. This reflects a westward deepening of the basin that was related to synsedimentary movements along the great regional faults with strike varying from N-S to N10°-20°W (Dardenne, 1978a, b; Alvarenga and Dardenne, 1978; Dardenne and Freitas-Silva, 1998a, b). The isotopic studies of $\delta^{13}C$ and $\delta^{18}O$, carried out on the carbonate rocks collected throughout the Bambuí Group (Fig. 26), confirm the evolutionary model proposed for the Bambuí Group and the lithostratigraphic correlation with the presence of glaciogenic dolomite (cap-dolomite) at the base of the sequence (negative $\delta^{13}C$) that evolved to marine facies (positive $\delta^{18}O$). The extremely high values for $\delta^{13}C$ (up to 12 $^{o}/_{oo}$) in the upper part of this group suggests confined marine conditions induced by tectonic processes (Santos *et al.*, 1997; Chang, 1997; Chang *et al.*, 1993; Martins, 1999; Misi *et al.*, 1997a, b).

The age of the Bambuí Group is still debatable in function of the absence of coeval volcanics. It falls within the broad interval between 950 Ma, corresponding to the opening of the Macaúbas Rift, and 600 Ma, which marks the end of the Brasiliano tectonic event. Historically, the first attempts to date the Bambuí rocks were based on Pb/Pb isotopic ratios determined in galena crystals collected from ore deposits, but these gave very variable ages without much significance (Amaral, 1968; Cassedanne and Lassere, 1969; Bez, 1980; Iyer, 1984; Iyer *et al.*, 1992). Lately, the K/Ar and Rb/Sr data for the pelitic sediments of the Bambuí Group show ages between 650 and 600 Ma (Amaral and Kawashita, 1967; Parenti Couto *et al.*, 1981; Thomaz Filho and Lima, 1981; Thomaz Filho *et al.*, 1998). In the 1990s, the first studies using the isotopic $^{87}Sr/^{86}Sr$ ratios in samples of limestone and dolomite of the Bambuí Group as a geochronometer by analogy with the reference seawater variation curves of variation gave surprisingly young ages in the order of 600 Ma (Misi *et al.*, 1997a, b; Chang, 1997; Chang *et al.*, 1993), corresponding to a Vendian age for this group.

The Pb-Zn-Ag-CaF_2 deposits associated with the Bambuí Group sediments occur near the towns of Januária, Itacarambi, Montalvânia and Serra do Ramalho on the left bank of the São Francisco River (Fig. 19). All these deposits occur in the same lithostratigraphic position, associated with a zone of pink saccharoidal dolomite, regionally anomalous in Pb and Zn. This zone occurs in the upper part of the first pelitic-carbonate regressive cycle of the Bambuí Group (Fig. 26). It is characterized by the presence of intraclasts and ooliths, and by bi-directional through cross-bedding, indicating a shallow platform under the influence of tidal currents. In some places, evidence of emersion can be observed at the top of the dolomitic zone, indicating the presence of a regional unconformity within the sedimentary sequence. This unconformity is covered by pink, beige, light grey micritic dolomite, showing cyanobacteria mats and teepee structures characteristic of a supratidal environment.

The mineralization, described by Dardenne (1979) is related to this internal unconformity, forming discontinuous or linked elongated lenses (Fig. 27). The ore bodies consist of silver-rich galena, sphalerite, fluorite and barite. Numerous secondary silver minerals appear where the primary mineralization undergoes supergene alteration. The migration of the mineralizing fluids along the unconformity is accompanied by dissolution and substitution of the dolomitic host. Generally, an important silicification event affects the dolomite in the upper part of the mineralized lenses. Fluorite appears as cement of the dolomitic fragments that show partial to total silicification. Regionally, a mineralogical zonation is observed from SW to NE along the São Francisco River, with increasing amounts of fluorite in relation to galena and sphalerite which predominate in the Januária region.

In this stable area of the SFC, the carbonate sequence frequently overlies directly the granite-gneiss basement. The fluid inclusion studies performed on fluorite of Mina Grande/Mina do Fabião near Itacarambi show carbonic monophase, aqueous biphase and aqueous triphase inclusions with halite daughter crystals. The salinity observed is high, varying from 15 to 30% wt eq. NaCl, while the homogenization temperatures are between 100 and 200 °C (Dardenne and Freitas-Silva, 1998a, b). The determination of the Pb/Pb isotope ratios in galena (Cassedanne and Lassèrre, 1969; Iyer *et al.*, 1992; Freitas-Silva, 1996; Freitas-Silva and Dardenne, 1997) show an incorporation of radiogenic Pb during the migration of mineralizing fluids (Fig. 22), which probably comes from the granitic basement of the SFC. These data allow the classification of the Pb-Zn-Ag-CaF_2 mineralization of the cratonic area as MVT strata-bound epigenetic deposits.

Geotectonic Evolution of the Brasília Fold Belt

The stabilization of the continental crust at the end of the Transamazonian Cycle led to the formation of the new Atlantica supercontinent, the relicts of which are represented in the BFB (figs. 28 and 29) by:

• Archean terranes of the Old Nucleus or Goiás Massif, with ages between 3.0 and 2.5 Ga (Queiroz *et al.*, 1999). This occupies an oval area in the northwestern part of the State of Goiás and is underlain by greenstone belts, such as Crixás, Pilar de Goiás, Guarinos and Goiás Velho, preserved along synformal throughs and isolated by the granite-gneiss complexes of Anta, Caiamar, Hidrolina and Uvá.

• Santa Terezinha Volcano-Sedimentary Sequence accreted to the Old Nucleus during the Transamazonian Cycle (Kuyumjian *et al.*, 1999; Oliveira and Pimentel, 1998).

• SFC appendix, situated in the NNE of Goiás and the SSE of Tocantins, consists of granite-gneiss and the volcano-sedimentary sequences of Almas, Dianópolis and São Domingos, of probable Paleoproterozoic age (*c.* 2.2Ga).

• Metasedimentary Sequence of the Ticunzal Formation deformed and metamorphosed during the Transamazonian event.

The fragmentation of the Atlantica supercontinent is marked by a series of events that occurred between the Paleoproterozoic (*c.* 2.0 Ga) and the Neoproterozoic up to the close of the Brasiliano Cycle (*c.* 600 Ma).

Paleoproterozoic Intracontinental Rift

The various stages, which mark the development of the intracontinental rift (Nilson *et al.*, 1994) can be schematically organized according to the following succession:

Intrusion of the Cana Brava, Niquelândia and Barro Alto mafic-ultramafic complexes

This event, of fundamental importance to the geotectonic framework in the northern segment of the BFB, occurred *c.* 2.0 Ga (Correia *et al.*, 1996, 1997). The layered complexes show an association of dunite, peridotite, pyroxenite and gabbro characteristic of an intracontinental rift environment. They present important potential for PGM mineralization in the transition between the peridotite and the pyroxenite (Lima, 1997). The Brasiliano tectonic event was responsible for the formation of a large asbestos deposit in the ultramafic rocks of the Cana Brava Complex. Lateritic alteration acting since the beginning of the Tertiary on dunite, peridotite and pyroxenite resulted in the formation of important lateritic nickel deposits known in the Niquelândia and Barro Alto complexes.

Intrusion of the Anorogenic Tin Granites of the Rio Paranã Sub-Province

In the N of Goiás, the granitic intrusions of Mendes, Mocambo, Pedra Branca, Mangabeira, Sucuri and Soledade, dated at 1.77 Ga (Pimentel *et al.*, 1991), precede the opening of the Araí Rift. These anorogenic granites show important mineralization of tin and indium, associated with greisenization and albitization processes (Marini and Botelho, 1986; Botelho, 1992; Botelho and Moura, 1998).

Araí Rift

The evolution of the Araí Rift may be subdivided according to the following stages:

– Opening and filling of the Araí Rift: this stage is marked by the continental alluvial and fluvial character of the sedimentation, as shown by the presence of intraformational breccia and conglomerate, coarse-grained quartzite and bimodal volcanism (basalt, rhyolite and ignimbrite), which is genetically related to the anorogenic granitic magmatism. The syn-rift sedimentation is observed in the Cavalcante, Teresina de Goiás, Monte Alegre and Arraias regions.

– Intrusions of Anorogenic Tin Granites in the Rio Tocantins Sub-Province. The granitic intrusions of Serra Branca, Serra da Mesa and Serra do Encosto, dated in 1.56 Ga (Pimentel *et al.*, 1991), probably correspond to a reactivation of the Araí Rift and show some tin mineralization associated mainly to greisenization processes (Botelho and Moura, 1998).

– Post-Rift Marine Transgression. After the syn-rift continental sedimentation, an important marine

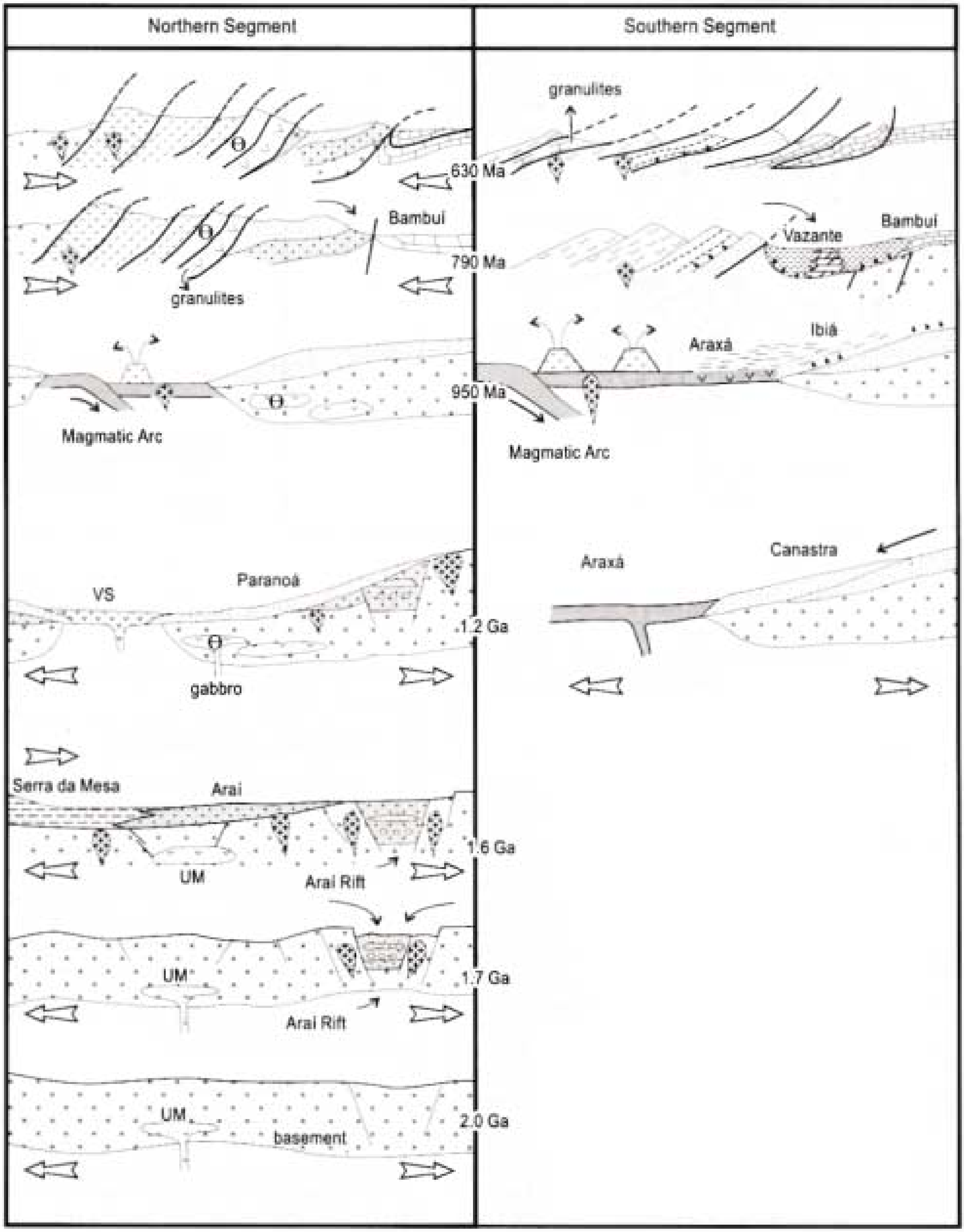


FIGURE 28 - Schematic tectonic evolution of the Brasília Fold Belt

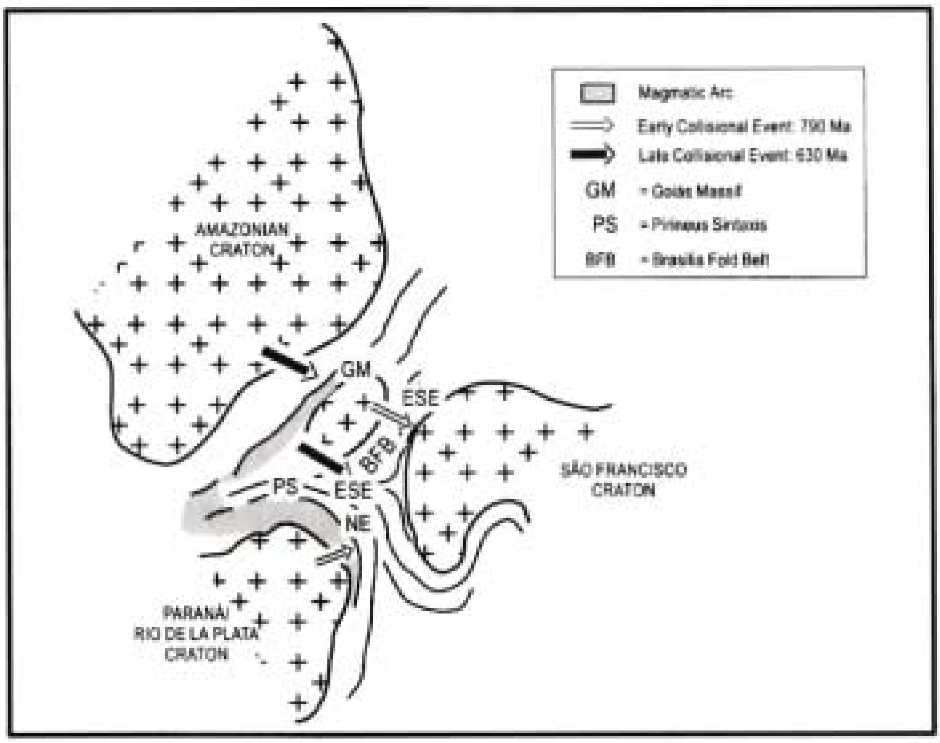


FIGURE 29 - Neoproterozoic evolution of the Brasília Fold Belt showing position of the cratons and of the sucessive compressional trends.

transgression occurred, which is represented by quartzite and metapelite beds deposited on a shallow platform. It is very likely that the metasediments of the Serra da Mesa Group constitute a metamorphic equivalent of the Araí marine sediments.

Meso-Neoproterozoic Basin

In the northern segment of the BFB, the development of a passive margin involved the deposition of the Paranoá Group sediments including quartzite, metapelite, metarhythmite, slate, limestone and dolomite on a marine platform, dominated by storm and tide currents. To the W of the Cana Brava, Niquelândia and Barro Alto complexes, the volcano-sedimentary sequences of Palmeirópolis, Indaianópolis and Juscelândia show ocean floor basaltic volcanism, dated at 1.3 - 1.25 Ga (Araújo *et al.*, 1995; Correia *et al.*, 1999), suggesting that the ocean opened during the Mesoproterozoic. Small volcanic massive sulfide (Cu-Pb-Zn) deposits are associated with this volcanism. The differentiated mafic units (mainly gabbro and anorthosite), dated at 1.3 Ga (Suita, 1998) and situated in the western part of the Cana Brava, Niquelândia and Barro Alto complexes, represent intrusions contemporaneous with this volcanism. The Castelã metasediments, which contain stratiform Pb-Zn occurrences and have been dated at 1.2 Ga (Freitas-Silva and Dardenne, 1997) are generally considered as lateral basin equivalent to the Paranoá Group.

In the southern segment of the BFB, the Canastra Group may be correlated to the Paranoá sediments (Dardenne, 1978a, b, 1979; Campos Neto, 1984a, b). Generally, this group is interpreted as the passive margin sediments equivalent to the deep basin Araxá rocks (Dardenne, 1978a, b, 1979; Campos Neto, 1984a, b; Marini *et al.*, 1984a, b; Fuck *et al.*, 1993). However, new Sm/Nd isotopic data obtained for the Araxá metasediments indicates a Neoproterozoic age for this group.

Neoproterozoic Brasiliano Evolution

The Neoproterozoic Brasiliano tectonic evolution is especially well illustrated in the southern segment of the BFB.

Development of the Magmatic Arc

Defined originally in the western and northwestern parts of Goiás (Pimentel and Fuck, 1991, 1992), the magmatic arc extends to the southern segment of the BFB, with prolongation under the sedimentary cover of the Paraná Basin. Dated between 950 and 600 Ma (Pimentel *et al.*, 1997), it is represented by arc-type volcano-sedimentary sequences and juvenile tonalitic/granodioritic rocks, frequently metamorphosed (orthogneiss) and mylonitized. Gold and base metal deposits (VMS) are associated with the volcano-sedimentary sequences. The individualization of this magmatic arc is related to an intra-oceanic subduction zone, plunging eastwards.

Araxá Oceanic Expansion

In the southern segment of the BFB, the Araxá Group is represented by volcano-sedimentary and sedimentary assemblages characteristic of distinct paleogeographic situations during the basin extensional stage. The psammo-pelitic micaschist and quartzite represent deep-water turbidite marine sediments, formed probably on the slope of the continental platform. The fine-grained amphibolite, of basaltic tholeiitic composition, associated with banded iron formation units and pelitic carbonaceous metassediments, is evidence for oceanic sea-floor expansion during the Araxá sedimentation. Geochemical data, including major, trace and rare earth elements, confirm the existence of an Araxá oceanic basin in the regions of Passos, Araxá, Abadia dos Dourados, Abadiânia and Silvânia, while psammo-pelitic sediments predominate in other regions such as Ipameri. The discontinuous distribution of volcanic occurrences may suggest that the opening of the ocean was localized and heterogeneous.

The Nd isotope data for the Araxá metasediments display a bimodal model age with a T_{DM} group between 1.5 and 1.0 Ga and another T_{DM} group from 2.0 to 1.8 Ga, indicating two independent and distinct sources for the origin of the sediments (Pimentel *et al.*, 1999). These data can indicate a Neoproterozoic age for the Araxá Group.

Ibiá Back-arc Basin

The Ibiá turbidite metasediments were deposited in a deep marine environment. They may represent a stratigraphic equivalent of the Araxá Group and possibly may be equivalent to the glacio-marine sediments correlated with the Jequitaí/Macaúbas units. Geochemical data indicate island arc characteristics for the sources (Seer, 1999; Seer *et al*, 1999), whereas Nd isotopes show model ages T_{DM} between 1.5 and 1.0 Ga and 2.3 - 1.8 Ga (Pimentel *et al.*, 1999). These results suggest the presence of a younger source in addition to the Paleoproterozoic one situated on the SFC. This younger source may be related to the erosion of the volcanic areas situated to the W and to the sedimentation of the products in a back-arc basin environment.

Early Brasiliano Collisional Event at 790 Ma

In the southern segment of the BFB, acid volcanic, sub-volcanic and plutonic rocks of Ipameri and Pires do Rio regions are associated with metasediments belonging to the Araxá Group. The sub-volcanic and plutonic intrusions of the Maratá Sequence, which were dated (U/Pb zircon age) at 794 Ma (Pimentel *et al.*, 1992), result from the re-melting of an older Paleoproterozoic continental crust (*c.* 2.0 Ga) during this early Brasiliano collisional event. The granodioritic to granitic rocks are syn to tardi-tectonic in relation to the main deformational event. They have a slight to strong peraluminous character, and some of the tin mineralization has been classified as being associated with S-type granites (Pimentel *et al.*, 1999).

In the northern segment of the BFB, the early Brasiliano collisional event is well described. The granulite metamorphism has been dated at 790 Ma in the Cana Brava, Niquelândia and Barro Alto complexes (Ferreira Filho *et al.*, 1994; Correia *et al.*, 1997). Garnet Sm/Nd ages between 730 and 700 Ma are also observed for metasediments of the Mara Rosa Arc.

Foreland Bambuí Basin

Following the initial BFB uplift, a large depression was formed in front of the belt, where the deposition of the Bambuí Group took place. Sedimentological and geochemical data (Guimarães, 1997; Castro, 1997; Castro and Dardenne, 1996) placed emphasis on the influence of the BFB uplift on the sedimentation, specifically in the southern segment. The Nd model ages display a more uniform distribution with T_{DM} values ranging between 1.9 and 1.4 Ga (Pimentel *et al.*, 1999). All these data favour a foreland basin as the depositional site for the Bambuí Group sediments.

The Vazante Group represents probably a transitional sequence between the Paranoá and the Bambuí groups.

Late Brasiliano Collisional Event at 630 Ma

In the southern segment of the BFB, new structural, geochemical and isotopic data indicate the existence of a late Brasiliano collisional event around 630 - 610 Ma. This event is suggested by the overlapping of Araxá, Ibiá and Canastra metasediments through low angle thrusts on the Bambuí Group. Furthermore, Araxá and Canastra clasts are incorporated in the Bambuí sediments in front of the nappes. Finally, Sm/Nd dating of the Araxá metamorphism and the syn-tectonic granites (Seer, 1999) lends support for this late Brasiliano collision that was also responsible for the granulite-grade metamorphism of the Anápolis-Itauçu Belt, dated at 610 Ma (Fischel *et al.*, 1999). Similar geochronological results have been obtained to the S of the BFB for metamorphism and syn-tectonic granites of the Socorro-Guaxupé nappe system (Campos Neto and Caby, 1999).

In the northern segment of the BFB, the late Brasiliano collisional event is reflected by overthrusting towards SE observed along the Pireneus mega-inflexion (Araújo Filho, 1999). During this second collisional event, tardi and post-tectonic intrusions mark the end of the Brasiliano Cycle. To this event are related a series of mafic-ultramafic differentiated complexes occurring to the S of the Serra Dourada Group. Between these complexes, the Americano do Brasil and Mangabal complexes, dated *c.* 610 Ma (Winge, 1995; Nilson *et al*, 1997) display typical arc geochemical signatures and contain important Cu-Ni-Co sulfide mineralization (Nilson, 1981). At the same time, numerous calc-alkaline K-rich granitic intrusions occur in the western part of the BFB, between 630 and 590 Ma (Pimentel *et al.*, 1999).

Several gold deposits are related to low and high angle shear zones developed at the end of the Brasiliano Cycle: Burracão, Santa Rita, Rio do Carmo, Cavalcante, Luziânia, Morro do Ouro, Santa Cruz, and Araxá (Dardenne and Freitas-Silva, 1998a, b). The last tectonic event related to the Brasiliano Cycle corresponds to the intracontinental, N30°E-striking Transbrasiliano transcurrent fault system (Marini *et al.*, 1984a, b).

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REFERENCES

Alkmin, F.F., Brito Neves, B.B. and Alves, J.A.C. (1993). Arcabouço tectônico do Cráton do São Francisco, uma revisão. In: Dominguez, J.M.L. and Misi, A. (eds). *Simpósio sobre o Cráton do São Francisco, Salvador, SBG/SGM/CNPq*, 45-62.

Alkmin ,F.F., Chemale Jr., F., Bacellar, L.A.P., Oliveira, J.R.P. and Magalhães, P.M. (1989). Arcabouço estrutural da porção sul da bacia do São Francisco. *Anais 5º Simp. Geol. Minas Gerais, Belo Horizonte, SBG*, 289-293.

Almeida, F.F.M. (1969). Evolução tectônica do Centro-Oeste brasileiro. *Anais da Academia Brasileira de Ciências*, 40, 280-296.

Almeida, F.F.M. (1977). O Cráton do São Francisco. *Revista Brasileira de Geociências*, 7, 349-364.

Alvarenga, C.J.S. and Dardenne, M.A. (1978). Geologia dos grupos Bambuí e Paranoá na Serra de São Domingos, MG. *An. 30º Cong. Bras. Geol., Recife, SBG*, 2, 546-556.

Amaral, G. (1966). Isótopos de chumbo e gênese das jazidas de Vazante e Itacarambi. *Anais 20º Cong. Bras. Geol., Rio de Janeiro, SBG, Publ. Esp.*, 1, 45-46.

Amaral, G. (1968). Resultados preliminares do estudo sobre a mineralização e composição isotópica do chumbo em galena de depósitos encaixados no Grupo Bambuí. *Anais 22º Cong. Bras. Geol., Belo Horizonte, SBG, Publ. Esp.*, 1, 51-53.

Amaral, G. and Kawashita, K. (1967). Determinação da idade do Grupo Bambuí pelo método Rb/Sr. *Anais 21º Cong. Bras. Geol., Curitiba, SBG*, 214-217.

Arantes, D., Buck, P.S., Osborne, G.A. and Porto, C.G. (1991a). A sequência vulcano-sedimentar de Mara Rosa e mineralizações associadas. *Bol. Centro-Oeste, SBG, Núcleo CO*, 14, 27-40.

Arantes, D., Osborne, G.A. and Buck, P.S. (1991b). The Mara Rosa volcano-sedimentary sequence and associated gold mineralization. In: *Brazil Gold'91, Proceedings*, ed. Ladeira, E.A., 221-229.

Araújo Filho, J.O. (1999). *Structural characteristics and tectonic evolution of the Pirineus sintaxis, Central Brazil*. PhD Thesis, Univ. of Illinois, 418 p. (unpublished).

Araújo, S.M. (1986). *Petrologia e mineralizações sulfetadas da sequência vulcano-sedimentar de Palmeirópolis, Goiás*. MSc. Diss., IG/UnB, Brasília, 196 p.

Araújo, S.M. (1998). The Palmeirópolis volcanogenic massive sulfide deposit, Tocantins State. *Workshop: Depósitos Minerais Brasileiros de Metais-Base, Salvador, CAPES-PADCT-ADIMB*, 75-79.

Araújo, S.M. and Nilson, A.A. (1987). Caracterização petroquímica dos anfibolitos da sequência vulcano-sedimentar de Palmeirópolis, Goiás. *Anais 1º Cong. Bras. Geoq., Porto Alegre, SBGq*, 1, 335-348.

Araújo, S.M. and Nilson, A.A. (1988). Depósito de zinco, cobre e chumbo de Palmeirópolis, Goiás. In: Schobbenhaus, C. and Coelho, C.E.S. (coords.)., *Principais Depósitos Minerais do Brasil, MME-DNPM-CVRD*, 3, 171-180.

Araújo, S.M., Fawcett, J.J. and Scott, S.D. (1995). Metamorphism of hydrothermally altered rocks in a volcanogenic massive sulfide deposit: the Palmeirópolis, Brazil example. *Revista Brasileira de Geociências*, 25, 173-184.

Babinski, M., Von Schmus, W.R., Chemale, Jr. and Kawashita, K. (1993). Evolução da porção sul da bacia do São Francisco baseada na geoquímica isotópica de Pb em rochas carbonáticas. *Anais 2º Simp. Craton São Francisco, Salvador, SBG*, 182-185.

Babinski, M., Von Schmus, W.R., Chemale, Jr. and Kawashita, K. (1997). Pb isotope behaviour in carbonate rocks from Bambuí Group, Brazil, during post-depositional events: a study case. *South-American Symp. on Isotope Geology, Campo do Jordão, SP, Ext. Abst.*, 38-40.

Baeta, Jr., J.D.A., Fogaça, C.C., Ramos, J.B., Jo, J.A.G. and Barberi, M. (1978). Projeto manganês no centro-sul de Goiás. *Final Report, DNPM/CPRM, Goiânia*, 7, 98 p.

Barbosa O. (1955). Guia das excursões. *9º Cong. Bras. Geol., São Paulo, Noticiário*, 3, 3-5.

Barbosa, O., Baptista, M.B., Dyer, R.C., Braun, O.P.G. and Cotta, J.C. (1969). Geologia e inventário dos recursos minerais do Projeto Brasília. *PROSPEC/DNPM, Final Report* (unpublished).

Barbosa, O., Braun, O.P.G., Dyer, R.C. and Cunha, C.A.B.R. (1970). Geologia da região do Triângulo Mineiro. (Projeto Chaminés). *DNPM/DFPM, Rio de Janeiro, Bol.* 136, 140 p.

Bez, L. (1980). Isótopos de chumbo e sua aplicação na prospecção mineral: uma reavaliação dos dados existentes. *Anais 31º Cong. Bras. Geol., Camboriú, SBG*, 3, 1417-1430.

Botelho, N.F. (1992). *Les ensembles granitiques subalcalins à peralumineux minéralisés Cu, Sn et In de la sous-province Paranã, état de Goiás, Brésil*. Doctorate Thesis, Univ. Paris VI, 343 p. (unpublished).

Botelho, N.F. and Moura, M.A. (1998). Granite-ore deposit relationships in Central Brazil. *Journal of South American Earth Sciences*, 11(5), 427-438.

Branco, J.J.R. (1957). O conglomerado do Samburá, Minas Gerais. *Anais da Academia Brasileira de Ciências*, 28(3), 295-301.

Branco, J.J.R. and Costa, M.T. (1961). Roteiro de Excursão Belo Horizonte- Brasilia. *14º Cong. Bras. Geol., Belo Horizonte, Public.* 15, 25 p.

Braun, O.P.G. (1968). Contribuição à estratigrafia do Grupo Bambuí. *An. 22º Cong. Bras. Geol., Belo Horizonte, SBG*, 155-166.

Brito Neves, B.B. and Cordani, U.G. (1991). Tectonic evolution of South America during the late Proterozoic. *Precambrian Research*, 53, 23-40.

Brod, J.A. (1988). *Geologia das rochas vulcano-sedimentares da região de Indaianópolis, GO, e lineamentos adjacentes*. MSc Diss., IG/UnB, Brasilia, 267 p. (unpublished).

Brod, J.A. and Jost, H. (1991). Características estruturais, litológicas e magmáticas da zona de cisalhamento dúctil do Rio Traíras, bloco do complexo de Niquelândia, Goiás. *Revista Brasileira de Geociências*, 21(3), 205-217.

Brod, J.A. and Jost, H. (1994). Revisão estratigráfica da região de Indaianópolis, GO. *An. 4º Simp. Geol. Centro-Oeste, Brasília, SBG*, 138-141.

Brod, J.A., Leonardos, O.H. and Meneses, P.R. (1991). Tectono-estratigrafia da Faixa Brasília na região do Alto Paranaíba. *An. 3º Simp. Geol. Centro-Oeste, Cuiabá, SBG*, 155-168.

Brod, J.A., Leonardos, O.H. and Meneses, P.R. (1992). Geoquímica da sequência vulcano-sedimentar de Abadia dos Dourados e Complexo chapada das Perdizes, Triângulo Mineiro, MG. *Revista Escola de Minas*, 45(1/2), 164-166.

Campos Neto, M.C. (1984a). Geometria e fases de dobramentos brasilianos superpostos no oeste de Minas Gerais. *Revista Brasileira de Geociências*, 14, 60-68.

Campos Neto, M.C. (1984b). Litoestratigrafia, relações estratigráficas e evolução paleogeográfica dos grupos Canastra e Paranoá (região Vazante-Lagamar, MG). *Revista Brasileira de Geociências*, 14, 81-91.

Campos Neto, M.C. and Caby, R. (1999). Neoproterozoic high-pressure metamorphism and tectonic constraint from the nappe system south of the São Francisco Craton, southeast Brazil. *Precambrian Research*, 97, 3-26.

Cassedanne, J. and Lassèrre, M. (1969). Análise isotópica pelo método do chumbo em algumas galenas brasileiras: descrição do método utilizado. *Mineração e Metalurgia*, 49(293), 215-224.

Cassedanne, J. and Lassèrre, M. (1970). Análise isotópica pelo método do chumbo de uma segunda série de galenas brasileiras. *Mineração e Metalurgia*, 51(301), 31-40.

Cassedanne, J., Duthou, J.L. and Lassèrre, M. (1972). Análise isotópica pelo método do chumbo de uma terceira série de galenas brasileiras. *Mineração e Metalurgia* (335), 12-19.

Castro, P.T.A. (1997). *Os conglomerados associados ao Grupo Bambuí na porção sudoeste do Cráton do São Francisco: sedimentologia, estratigrafia e implicações tectônicas.* Doctorate Thesis, IG/UnB, Brasilia, 267 p.

Castro, P.T.A. and Dardenne, M.A. (1996). O conglomerado Samburá (Grupo Bambuí, Neoproterozóico) e rochas sedimentares associadas no flanco leste da Serra da Pimenta, SW de Minas Gerais: um sistema de fan-delta. *Geonomos*, 3(2), 35-41.

Chang, H.K. (1997). *Isótopos estáveis (C,H,O) e $^{87}Sr/^{86}Sr$: implicações na estratigrafia e na paleo-circulação de fluidos na bacia do São Francisco.* Livre-Docência Thesis, 129 p. (unpublished).

Chang, H.K., Kawashita, K., Alkmin, F.F. and Moreira, M.Z. (1993). Considerações sobre a estratigrafia isotópica do Grupo Bambuí. *Anais 2° Simp. Craton São Francisco, Salvador, SBG*, 195-196.

Chang, H.K., Miranda, F.P., Magalhães, L. and Alkmin, F.F. (1988). Considerações sobre a evolução tectônica da Bacia do São Francisco. *Anais 35° Cong. Bras. Geol., Belém, SBG*, 7, 2076-2090.

Chiavegatto, J.R.S. (1992). *Análise estratigráfica das sequências tempestíticas da Formação Três Marias (Proterozóico Superior) na porção meridional da Bacia do São Francisco.* MSc Diss., UFOP, 216 p. (unpublished).

Chiavegatto, J.R.S. and Dardenne, M.A. (1997). Contribuição à sedimentologia e estratigrafia do Grupo Bambuí no norte de Minas Gerais. *Anais 9° Simp. Geol. Minas Gerais, Ouro Preto, SBG*, **14**, 81-82.

Cloud, P. and Dardenne, M.A. (1973). Proterozoic age of the Bambuí Group in Brazil. *Geological Society of America Bulletin*, 84(5), 1673-1676.

Correia, C.T., Girardi, V.A.V., Lambert, D.D., Kinny, P.D. and Reeves, S.J. (1996). 2 Ga U-Pb (SHRIMP II) and Re-Os ages for the Niquelândia basic-ultrabasic layered intrusion, Central Goiás, Brazil. *An. 39° Cong. Bras. Geol., Salvador, SBG*, 6, 187-189.

Correia, C.T., Girardi, V.A.V., Tassinari, C.C.G. and Jost, H. (1997). Rb-Sr and Sm-Nd geochronology of the Cana Brava layered mafic-ultramafic intrusion, Brazil, and considerations regarding its tectonic evolution. *Revista Brasileira de Geociências*, 27, 163-168.

Correia, C.T., Tassinari, C.C.G., Lambert, D.D., Kinny, P. and Girardi, V.A.V. (1997). U-Pb (SHRIMP), Sm-Nd and Re-Os systematics of the Cana Brava, Niquelândia and Barro Alto layered intrusions in Central Brazil, and constraints on the tectonic evolution. *South-American Symp. on Isotope Geology, Campo do Jordão, SP, Ext. Abst.*, 88-89.

Correia, C.T., Jost, H., Tassinari, C.C.G., Girardi, V.A.V. and Kinny, P.D. (1999). Ectasian Mesoproterozoic U-Pb ages (SHRIMP II) for the metavolcano-sedimentary sequences of Juscelândia and Indaianópolis and for high grade metamorphosed rocks of Barrro Alto stratiform igneous complex, Goiás State, Central Brazil. *Actas 2X South American Symposium on Isotope Geology, Cordoba, Argentina*, 31-33.

Costa, L.A.M. and Angeiras, A.G. (1971). Geosynclinal evolution in the epi-Baikalian platform of Central Brazil. *Geologische Rundschau*, **60**(2), 1024-1050.

Danni, J.C.M. and Kuyumjian, R.M. (1984). A origem dos anfibolitos basais da seqüência vulcano-sedimentar de Juscelândia, Goiás. *Anais 34° Cong. Bras. Geol., Rio de Janeiro, SBG*, **9**, 4126-4136.

Danni, J.C.M. and Leonardos, O.H. (1980). The Niquelândia Mafic-ultramafic granulites and gabro-anortosites-metavolcanic associations (unpublished).

Danni, J.C.M., Dardenne, M.A., Fuck, R.A. and Ribeiro, M.J. (1973). Geologia da extremidade sudoeste da Serra Dourada (Goiás). *Revista Brasileira de Geociências*, 3(3), 160-180.

Dardenne, M.A. (1978). Geologia da região de Morro Agudo (MG). *Bol. Núcleo Centro-Oeste, SBG*, 7-8, 68-94.

Dardenne, M.A. (1978a). Zonação tectônica na borda ocidental do Cráton São Francisco. *An. 30° Cong. Bras. Geol., Recife, SBG*, 1, 299-308.

Dardenne, M.A. (1978b). Síntese sobre a estratigrafia do Grupo Bambuí no Brasil Central. *An. 30° Cong. Bras. Geol., Recife, SBG*, 2, 597-610.

Dardenne, M.A. (1979). *Les minéralisations de plomb, zinc, fluor du Protérozoïque Supérieur dans le Brésil Central.* Doctorate Thesis, Univ. Paris VI, 251 p. (unpublished).

Dardenne, M.A. (1981). Os grupos Paranoá e Bambuí na Faixa Dobrada Brasília. *An. Simp. Cráton São Francisco e suas Faixas Marginais, Salvador, SBG*, 140-157.

Dardenne, M.A. and Faria, A. (1985). Estratigrafia do Grupo Paranoá na região de Alto Paraíso, GO. *Anais 2° Simp. Geol. Centro-Oeste, Goiânia, SBG*, 65-71.

Dardenne, M.A. and Freitas-Silva, F.H. (1998a). Depósito Pb-Zn-F-Ba do Supergrupo São Francisco. *Anais 40° Cong. Bras. Geol., Belo Horizonte, SBG*, 133.

Dardenne, M.A. and Freitas-Silva, F.H. (1998b). Modelos genéticos dos depósitos Pb-Zn nos grupos Bambuí e Vazante. *Workshop: Depósitos Minerais Brasileiros de Metais-Base. Salvador, CAPES-PADCT-ADIMB*, 86-93.

Dardenne, M.A., Campos, J.E., Alvarenga, C.J.S., Martins, F.L. and Botelho, N.F. (1999). A sequência sedimentar do Grupo Araí na região da Chapada dos Veadeiros, Goiás. *Anais 7° Simp. Geol. Centro-Oeste* (submitted).

Dardenne, M.A., Campos, J.E., Martins, F.L. and Alvarenga, C.J.S. (1997). Geologia da região de Alto Paraíso de Goiás/Parque Nacional da Chapada dos Veadeiros, GO. *Relatório FAPDF, Geologia do Distrito Federal e Entorno Próximo*, 87-103 (unpublished).

Dardenne, M.A., Danni, J.C.M. and Fuck, R.A. (1981). Revisão estratigráfica do Grupo Araxá na região da Serra Dourada, GO. *Anais 1° Simp. Geol. Centro-Oeste, Goiânia, SBG*, 664-678.

Dardenne, M.A., Faria, A. and Andrade, G.F. (1976). Occurrence de stromatolithes columnaires dans le Groupe Bambuí (Goiás, Brésil). *Anais da Academia Brasileira de Ciências*, **48**(3), 555-566.

Dardenne, M.A., Faria, A., Magalhães, L.F. and Soares, L.A. (1978). O tilito da base do Grupo Bambuí na borda ocidental do Cráton do São Francisco. *Bol. Núcleo Centro-Oeste, SBG*, 7-8, 85-97.

Dardenne, M.A., Freitas-Silva, F.H., Nogueira, G.M.S. and Souza, J.F.C. (1997). Depósitos de fosfato de Rocinha e Lagamar, Minas Gerais. In: Principais Depósitos Minerais do Brasil, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S. *DNPM/CPRM*, 4C, 113-122.

Dardenne, M.A., Freitas-Silva, F.H., Souza, J.C.F and Campo, J.E.G (1998). Evolução tectono-sedimentar do Grupo Vazante no contexto da Faixa de Dobramentos Brasília. *Anais 40° Cong. Bras. Geol., Belo Horizonte, SBG*, 26.

Dardenne, M.A., Fuck, R.A., Pimentel, M.M., Meneses, P.R. and Strieder, A.J. (1992). Geologia da região de Ipameri, GO. *Bol. Res. 37° Cong. Bras. Geol., São Paulo, SBG*, 2, 149-150.

Dardenne, M.A., Magalhães, L.F. and Soares, L.A. (1978). Geologia do Grupo Bambuí no vale do Rio Paranã. *An. 30° Cong. Bras. Geol., Recife, SBG*, 2, 611-622.

Dardenne, M.A., Nogueira, G.M.S. and Campos, J.E.G. (1989). Litoestratigrafia e ambientes de sedimentação na região de Lagamar, MG. *An. 5° Simp. Geol. Minas Gerais, Belo Horizonte, SBG-Núcleo MG*, 10, 271-274.

Drake Jr., A.A. (1980). The Serra de Caldas windows. *U.S. Geological Survey, Professional Paper*, **1119**-A.

Dyer, R.C. (1970). Notas sobre o conglomerado São Miguel, basal da Formação Paranoá. *Revista Escola de Minas*, **28**, 27-30.

Fairchild, T.R., Schopf, J.W., Shen-Miller, J., Guimarães, E.M., Edwards, M.D., Lagstein, A., Li, X., Pabst, M. and Melo Filho, L.S. (1996). Recent discoveries of Proterozoic microfossils in south-central Brazil. *Precambrian Research*, **80**, 125-152.

Faria, A. (1995). *Estratigrafia e sistemas deposicionais do Grupo Paranoá nas áreas de Cristalina, Distrito Federal e São João D'Aliança – Alto Paraíso de Goiás*. Doctorate Thesis, IG/UnB, Brasilia, 199 p. (unpublished).

Faria, A. and Dardenne, M.A. (1995). Estratigrafia do Grupo Paranoá na região de Alto Paraíso de Goiás – São João D'Aliança. *Anais 5° Simp. Geol. Centro-Oeste, Goiânia, SBG, Bol. Res. Exp.*, 75-77.

Ferrari, P.G. (1981). Sequência vulcano-turbidítica de margem continental-borda ocidental da plataforma São Francisco. *An. 1° Simp. Geol. Centro-Oeste, Goiânia, SBG*, 702-715.

Ferrari, P.G. (1989a). Grupo Araxá- associações litológicas e metamorfismo na região do Triângulo Mineiro. *An. 5° Simp. Geol. Minas Gerais, Belo Horizonte, SBG*, 253-256.

Ferrari, P.G. (1989b). A Formação Ibiá e sua pertinência ao Grupo Araxá. *An. 5° Simp. Geol. Minas Gerais, Belo Horizonte, SBG*, 257-261.

Ferreira Filho, C.F., Kamo, S.L., Fuck, R.A., Krogh, T.E. and Naldrett, A.J. (1994). Zircon and rutile U-Pb geochronology of the Niquelândia layered mafic and ultramafic intrusion, Brazil: constraints for the timing of the magmatism and high grade metamorphism. *Precambrian Research*, **68**, 241-255.

Ferreira Filho, C.F., Nilson, A.A. and Naldrett, A.J. (1992). The Niquelândia mafic-ultramafic complex, Goiás, Brazil: a contribution to the ophiolite ¥ stratiform controversy based on new geological and structural data. *Precambrian Research*, **57**, 1-19.

Figueiredo, J.A., Leão Neto, R. and Valente, C.R. (1981). Depósitos de sulfetos maciços de Zn, Cu e Pb da região de Palmeirópolis, GO. *Atas 1° Simp. Geol. Centro-Oeste, Goiânia, SBG*, 422-441.

Fischel, D.P., Pimentel, M.M. and Fuck, R.A. (1999). Idade do metamorfismo granulítico do Complexo Anápolis-Itauçu, determinada pelo método Sm-Nd. *Revista Brasileira de Geociências*, **28(4)**, 543-544.

Fonseca, M.A. (1996). *Estilos estruturais e arcabouço tectônico do segmento setentrional da Faixa Brasília*. Doctorate Thesis, IG/UnB, Brasília, 172 p. (unpublished).

Fonseca, M.A. and Dardenne, M.A. (1993). Sistemas de falhamentos transcorrentes na porção norte da Faixa de Dobramentos Brasília. *An. 2° Simp. Cráton São Francisco e suas Faixas Marginais, Salvador, SBG*, 280-283.

Fonseca, M.A. and Dardenne, M.A. (1995). Estilos estruturais e a compartimentação tectônica do segmento setentrional da Faixa Brasília. *Anais 5° Simp. Nac. Est. Tect., Gramado, SBG*, 36-38.

Fonseca, M.A., Dardenne, M.A. and Uhlein, A. (1995). Faixa Brasília, Setor Setentrional: estilos estruturais e arcabouço tectônico. *Revista Brasileira de Geociências*, **25**(4), 267-278.

Freitas, M.E. (1994). *Hidrotermalismo na região de Vianópolis, Goiás*. MSc Diss., IG/UnB, Brasília, 106 p. (unpublished).

Freitas-Silva, F.H. (1996). *Metalogênese do depósito do Morro do Ouro, Paracatu, MG*. Doctorate Thesis, IG/UnB, Brasília, 339 p. (unpublished).

Freitas-Silva, F.H. and Dardenne, M.A. (1992). Evolução estrutural das formações Paracatu e Vazante na região de Paracatu, MG. *REM*, **45**(1/2), 91-93.

Freitas-Silva, F.H. and Dardenne, M.A. (1994). Proposta de subdivisão estratigráfica formal para o Grupo Canastra no oeste de Minas Gerais e leste de Goiás. *An. 4° Simp. Geol. Centro-Oeste, Goiânia, SBG*, 161-163.

Freitas-Silva, F.H. and Dardenne, M.A. (1997). Pb/Pb Isotopic patterns of galenas from Morro do Ouro (Paracatu Formation), Morro Agudo/Vazante (Vazante Formation) and Bambui Group Deposits. South-American Symp. *On Isotope Geol., Campos do Jordão, SP, Ext. Abst.*, 118-120.

Freitas-Silva, F.H., Dardenne, M.A. and Jost, H. (1991). Lithostructural control of the Morro do Ouro, Paracatu, Minas Gerais, Gold Deposit. In: *Brazil Gold'91*, ed. Ladeira E.A., pp. 681-683. Belkama.

Fuck, R.A. (1994). A Faixa Brasília e a compartimentação tectônica na Província Tocantins. *Anais 4° Simp. Geol. Centro-Oeste, Brasília, SBG*, 184-187.

Fuck, R.A. and Marini, O.J. (1981). O Grupo Araxá e unidades homotaxiais. *An. 1° Simp. Cráton São Francisco e suas Faixas Marginais, Salvador, SBG/CPM*, 118-130.

Fuck, R.A., Danni, J.C.M., Winge, M., Andrade, G.F., Barreira, C.F., Leonardos, O.H. and Kuyumjian, R. (1981). Geologia da região de Goianésia. *Atas 1° Simp. Geol. Centro-Oeste, Geologia do Pré-Cambriano, Goiânia, SBG*, 447-469.

Fuck, R.A., Jardim de Sá, E.F., Pimentel, M.M., Dardenne, M.A. and Pedrosa-Soares, A.C. (1993). As faixas de dobramentos marginais do Cráton do São Francisco: síntese dos conhecimentos. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and Misi, A., pp. 165-181. SBG/SGM/CNPq, Salvador.

Fuck, R.A., Marini, O.J., Dardenne, M.A. and Figueiredo, A.N. (1988). Coberturas metassedimentares do Proterozóico Médio: os grupos Araí e Paranoá na região de Niquelândia-Colinas, Goiás. *Revista Brasileira de Geociências*, **18**, 54-62.

Fuck, R.A., Pimentel, M.M. and Botelho, N.F. (1987). Granitoid rocks in west-central Brazil: a review. *Symp. Granites and Associated Mineralizations, Salvador, SME-CPM/SBG, Ext. Abstr.*, 118-130.

Fuck, R.A., Pimentel, M.M. and Silva, J.H.D. (1994). Compartimentação tectônica na porção oriental da Província Tocantins. *Anais 38° Cong. Bras. Geol., Camboriú, SBG*, **1**, 215-216.

Guimarães, E.M. (1997). *Estudos de Proveniência e diagênese com ênfase na caracterização dos filossilicatos* dos grupos Paranoá e Bambuí na região de Bezerra – Cabeceiras, GO. Doctorate Thesis, IG/UnB, Brasília, 269 p. (unpublished).

Guimarães, E.M. and Dardenne, M.A. (1989). Geologia da região de Bezerra e Cabeceiras, GO. *Anais 5° Simp. Geol. Minas Gerais, Belo Horizonte, SBG/NMG, Bol.* **10**, 266-270.

Hitzmann, M.W. (1995). Mineralization in the Irish Zn-Pb(Ba-As) ore field. In: Irish Carbonate-Hosted Zn-Pb Deposits, eds. Anderson, K., Ashton, J., Earls, G., Hitzmann, M. and Tear, S. *SEG Guide book*, **21**, 132-137.

Hitzmann, M.W., Thorman, C.H., Romagna, G., Oliveira, T.F., Dardenne, M.A. and Drew, L.J. (1995). The Morro Agudo Zn-Pb deposit, Minas Gerais, Brazil: a proterozoic Irish-Type Carbonate-Hosted SEDEX – Replacement deposit. *Annual Meeting, New Orleans, GSA, Abstract*, 408.

Iyer, S.S. (1984). A discussion on the lead isotope geochemistry of galenas from the Bambuí Group, Brazil. Miner. *Deposita*, **19**, 132-137.

Iyer, S.S.; Hoefs, J. and Krouse, H. (1992). Sulfur and lead isotope geochemistry of galenas from Bambuí Group, Minas Gerais, Brazil. implications for ore genesis. *Economic Geology*, **87**, 437-443.

Kuyumjian, R. (1989). *The geochemistry and tectonic significance of amphibolites from the Chapada sequence, Central Brazil*. PhD Thesis, Univ. of London, 289 p. (unpublished).

Kuyumjian, R.M. (1994). A seqüência Mara Rosa na região da Chapada, Goiás. *Bol. Geol. Centro-Oeste, SBG*, 34-38.

Kuyumjian, R.M. (1995). Diversity of fluids in the origin of the Chapada Cu-Au deposit, Goiás. *Revista Brasileira de Geociências*, **25**(3), 203-205.

Kuyumjian, R.M. (1998). The magmatic arc of western Goiás: a promising exploration target. *Workshop: Depósitos Minerais Brasileiros de Metais-Base, Salvador, CAPES-CNPq-ADIMB*, 80-85.

Kuyumjian, R.M. (1989). Geoquímica e significado do posicionamento geotectônico de rochas plutônicas de Chapada, Goiás, Brasil. *An. 2° Cong. Bras. Geoq., Rio de Janeiro, SBGq*, 195-201.

Kuyumjian, R.M. (1991). A suggested hydrothermal exalative origin for the Chapada copper-gold deposit, Brazil. In: *Brazil Gold'91 Proceedings*, ed. Ladeira E.A. pp. 231-234.

Kuyumjian, R.M. and Danni, J.C.M. (1991). Geoquímica dos anfibolitos da seqüência de Juscelândia, Goiás: implicações geotectônicas. *Revista Brasileira de Geociências*, **21**(3), 218-223.

Kuyumjian, R.M., Campos, J.E.E., Oliveira, C.G. and Queiroz, C.L. (1999). Registros de evolução Transamazônica na província estrutural do Tocantins: exemplo da região de Campinorte-Alto Horizonte (GO). *Revista Brasileira de Geociências*, (in press).

Kuyumjian, R.M., Campos, J.E.G., Oliveira, C.G. and Queiroz, C.L. (1999). Registros da evolução Transamazônica na província estrutural do Tocantins: exemplo da região de Campinorte-Alto Horizonte (GO). *Revista Brasileira de Geociências*, (in press).

Lacerda Filho, J.V. and Oliveira, C.C. (1995). Geologia da região centro-sul de Goiás. *Bol. Geol. Centro-Oeste*, **18**(1/2), 3-19.

Laranjeira, N.P.F. (1992). *Geologia do Grupo Paranoá na região de Unaí: uma plataforma siliciclástica-carbonática no Proterozóico de Minas Gerais*. MSc Diss., IG/UnB, Brasília, 213 p. (unpublished).

Laranjeira, N.P.F. and Dardenne, M.A. (1993). Relação estratigráfica e aspectos estruturais das sequências supracrustais aflorantes a NE de Unaí, MG. *An. 36° Cong. Bras. Geol., Natal, SBG*, **1**, 23-33.

Leonardos, O.H., Jost, H., Fuck, R.A. and Barros, J.C. (1990). Geologia da região aurífera de Santa Cruz de Goiás. *Resumos 36° Cong. Bras. Geol., Natal, SBG*, 338 p.

Lima, T.M. (1997). *Geologia, estratigrafia e petrologia da porção sul do complexo máfico-ultramáfico de Cana Brava, Goiás*. MSc. Diss, IG/UnB, Brasília, 312 p. (unpublished).

Madalosso, A. (1980). Considerações sobre a paleogeografia do Grupo Bambuí na região de Paracatu, MG. *An. 31° Cong. Bras. Geol., Camboriú, SBG*, **2**, 772-785.

Madalosso, A. and Valle, C.R.O. (1978). Considerações sobre a estratigrafia e sedimentologia do Grupo Bambuí na região de Paracatu – Morro Agudo (MG). *An. 30° Cong. Bras. Geol., Recife, SBG*, **2**, 622-634.

Magalhães, P.M., Chemale, Jr. and Alkmin, F.F. (1989). Estilo tectônico da porção sudoeste da bacia do São Francisco. *An. 5° Simp. Geol. Minas Gerais, Belo Horizonte, SBG*, 284-288.

Marchese, H.G. (1974). Estromatólitos "*Gymnoselenides*" en el lado oriental de Minas Gerais, Brasil. *Revista Brasileira de Geociências*, **4**(4), 257-271.

Marini, O.J. and Botelho, N.F. (1986). A província de granitos estaníferos de Goiás. *Revista Brasileira de Geociências*, **16**(1), 19-131.

Marini, O.J. and Fuck, R.A. (1981). A Formação Minaçu: estratigrafia, tectônica e metamorfismo. *Atas 1° Simp. Geol. Centro-Oeste, Goiânia, SBG/NCO*, 716-745.

Marini, O.J., Fuck, R.A., Danni, J.C.M. and Dardenne, M.A. (1981). A evolução geotectônica da Faixa Brasília e do seu embasamento. *An. Simp. Cráton São Francisco e suas Faixas Marginais, Salvador, SME-CBPM/SBG/NBA-SE*, 100-115.

Marini, O.J., Fuck, R.A., Danni, J.C.M., Dardenne, M.A., Loguércio, S.O. and Ramalho, R. (1984a). As faixas de dobramento Brasília, Uruaçu, Paraguai-Araguaia e o Maciço Mediano de Goiás. In: *Geologia do Brasil*, coords. Schobbenhaus, C., Campos, D.A, Derze, G.R. and Asmus, H.E., pp. 251-303. Brasília, DNPM.

Marini, O.J., Fuck, R.A., Dardenne, M.A. and Danni, J.C.M. (1984b). Província Tocantins, Setores Central e Sudeste. In: *O Pré-Cambriano do Brasil*, eds. Almeida, F.F.M. and Hasui, Y., pp. 205-264. São Paulo, Edgar Blücher.

Marini O.J., Fuck R.A., Dardenne M.A. and Faria A. (1977). Contribuição à geologia do Pré-cambriano da porção central de Goiás. *Revista Brasileira de Geociências*, 7(4), 304-324.

Marini, O.J., Liberal, C.S., Reis, L.T., Trindade, C.A.H. and Souza, S.L. (1978). Nova unidade litoestratigráfica do Pré-cambriano do Estado de Goiás. *Resumos 30° Cong. Bras. Geol., Recife, SBG, Bol. Esp.*, 1, 126-127.

Martins, M. (1999). *Análise estratigráfica das sequências mesoproterozóicas (borda oeste) e neoproterozóicas da bacia do São Francisco*. MSc Diss., UFRGS, Porto Alegre, 214 p. (unpublished).

Melo Filho, L.S. (1996). *Estromatólitos do Grupo Paranoá na região de São Gabriel e Mato Seco/Mimoso, Goiás*. MSc Diss., IG/UnB, Brasília, 77 p. (unpublished).

Misi, A., Iyer, S.S. and Tassinari, C.C.G. (1996). Boquira (2,5Ga) and Morro Agudo (0,65Ga) lead-zinc deposits, Brazil: New SEDEX Subtypes? *Anais 39° Cong. Bras. Geol., Salvador, SBG*, 7, 251-253.

Misi, A., Iyer, S.S., Tassinari, C.C.G., Coelho, C.E.S., Kyle, J.R., Franca-Rocha, W.J.S., Gomes, A.S.R., Cunha, I.A., Carvalho, I.G. and Mônaco, Filho V. (1998). Integrated studies and metallogenic evolution of the Proterozoic sediment-hosted Pb-Zn-Ag sulfide deposits of the São Francisco Craton, Brazil. *Workshop: Depósitos Minerais Brasileiros de Metais-Base, Salvador, CAPES-PADCT-ADIMB-UFBA*, 94-101.

Misi, A., Tassinari, C.C.G. and Iyer, S.S. (1997a). New isotope data from the Proterozoic lead-zinc (Ag) sediment-hosted sulfide deposits of Brazil: implications for their metallogenic evolution. *South-American Symp. on Isotope Geology, Campo do Jordão, SP, Ext. Abst.*, 201-203.

Misi, A., Veizer, J., Kawashita, K. and Dardenne, M.A. (1997b). The age of the Neoproterozoic carbonate platform sedimentation based on $^{87}Sr/^{86}Sr$ determinations, Bambuí and Una groups, Brazil. *South-American Symp. on Isotope Geology, Campos do Jordão, SP, Ext. Abst.*, 199-200.

Moeri, E. (1972). On a columnar stromatolite in the Precambrian Bambuí Group of Central Brazil. Eclog. Geol. Helv., 65(1), 185-195.

Monteiro, L.V.S. (1997). *Contribuição à gênese das mineralizações de Zn da mina de Vazante, MG*. MSc. Diss., USP, 159 p. (unpublished).

Monteiro, L.V.S., Bettencourt, J.S. and Grasa, R. (1996). Contribuição à gênese das mineralizações de Zn e Pb da mina de Vazante (MG). *Anais 39° Cong. Bras. Geol., Salvador, SBG*, 5, 501-503.

Moore, S.L. (1956). Zinc and copper deposits of the Vazante area, Minas Gerais, Brazil. *USGS Report*, **386**, 16 p.

Morães, R., Vasconcellos, A.C.B.C. and Fuck, R.A. (1994). Uma descontinuidade metamórfica no Complexo Barro Alto: evidência para sua compartimentação? *An. 4° Simp. Geol. Centro-Oeste, Brasília, SBG*, 135-137.

Nascimento, F.S., Veloso, F.D.L.I. and Sabóia, L.A. (1981). Caracterização e discussão sobre a sequência vulcano-sedimentar da borda oeste do maciço de Niquelândia (oeste do rio Traíras). *Atas 1° Simp. Geol. Centro-Oeste, Goiânia, SBG*, 470-492.

Nilson, A.A. (1981). *The nature of the Americano do Brasil mafic-ultramafic complex and associated sulfide mineralization*. PhD Thesis, Univ. of Western Ontario, 460 p. (unpublished).

Nilson, A.A., Gioia, S.M. and Pimentel, M.M. (1997). Idade Sm-Nd do Complexo máfico-ultramáfico de Americano do Brasil, Goiás, e características isotópicas das rochas gnáissicas encaixantes. *Cong. Bras. Geoquim.* (in press).

Nilson, A.A., Botelho, N.F. and Ferreira Filho, C.F. (1994). Rifteamento mesoproterozóico no Centro-Oeste de Goiás. *Resumos 38° Cong. Bras. Geol., Camboriú, SBG*, 258-259.

Nogueira, G.S.M. (1993). *Enquadramento litoestratigráfico, sedimentologia e evolução geoquímica do depósito fosfático de Lagamar, MG, Formação Vazante.* MSc Diss., IG/UnB, Brasília, 165p. (unpublished).

Oliveira, A.M. (1993). *Petrografia, estratigrafia, petroquímica e potencialidade para elementos do Grupo da Platina (EGP) do Complexo Barro Alto, na região de Goianésia, Goiás.* MSc. Diss., IG/UnB, Brasília, 86 p. (unpublished).

Oliveira, C.G., Sintia, A.V. and Barbosa, I.O. (1997). Influência da deformação transcorrente NS na mineralização aurífera na sequência vulcano-sedimentar de Mara Rosa. *An. 6° Simp. Geol. Centro-Oeste, Cuiabá, SBG*, 59-61.

Oliveira, G.G. and Pimentel, M.M. (1998). Geology of the northern part of the Hidrolina Dome and Sm-Nd data for metavolcanic and granitoid rocks in the Campinorte region, Goiás, Central Brazil. *Abstr. 14th Int. Conf. on Basement Tectonics, Ouro Preto, Brazil*, 55-56.

Oliveira, M.A. (1967). Contribuição à geologia da parte sul da bacia do São Francisco e áreas adjacentes. *Coletânea de Rel. de Exploração, PETROBRÁS/ DEPIN, CENPES, Rio de Janeiro*, 1, 71-105.

Oliveira, T.F. (1998). As minas de Vazante e Morro Agudo, Minas Gerais. *Workshop: Depósitos Minerais Brasileiros de Metais-Base, Salvador, CAPES-PADCT-ADIMB*, 48-57.

Oliveira, W.R., Leonardos, O.H., Kuyumjian, R.M. and Martins, E.S. (1992). Ocorrências de ouro de Silvânia, Goiás. *Res. Exp. 37° Cong. Bras. Geol., São Paulo, SBG*, 227-228.

Parenti-Couto, J.G. and Bez, L. (1980). A glaciação Jequitaí: um guia estratigráfico para o Pré-Cambriano Superior no Brasil. *Revista Brasileira de Geociências*, 11(1), 17-21.

Parenti-Couto, J.G., Cordani, U.G., Kawashita, K., Iyer, S.S. and Morães, N.M.P. (1981). Considerações sobre a idade do Grupo Bambuí com base em análises isotópicas de Sr e Pb. *Revista Brasileira de Geociências*, 11(1), 5-16.

Pedrosa-Soares, A.C., Dardenne, M.A., Hasui, Y. and Castro, F.D.C. (1994). *Mapa metalogenético de Minas Gerais, 1/1000.000.* COMIG, Belo Horizonte.

Pedrosa-Soares, A.C., Dardenne, M.A., Hasui, Y., Castro, F.D.C. and Carvalho, M.V.A. (1994). *Nota explicativa dos mapas geológico, metalogenético e de ocorrências minerais do Estado de Minas Gerais, 1/1000.000.* COMIG, Belo Horizonte, 97 p.

Pereira, E.L., Poli, A.J.C., Teixeira, N.A. and Ferreira, M.C.B. (1981). Prospecção e caracterização genética das mineralizações estaníferas da porção sudeste do Estado de Goiás. *Atas 1° Simp. Geol. Centro-Oeste, Goiânia, SBG/NCO*, 746-770.

Pereira, L.F. (1992). *Relações tectono-estratigráficas entre as unidades Canastra e Ibiá na região de Coromandel, MG.* MSc Diss., IG/UnB, Brasília, 73 p. (unpublished).

Pereira, L.F., Dardenne, M.A., Rosiére, C.A. and Pedrosa-Soares, A.C. (1994). Evolução geológica dos grupos Canastra e Ibiá na região entre Coromandel e Guarda-Mor, MG. *Geonomos*, 2(1), 22-32.

Pimentel, M.M. and Fuck, R.A. (1991). Origin of orthogneiss and metavolcanic rock units in western Goiás: Neoproterozoic crustal accretion. *Geochimica Brasiliensis*, 5, 133-152.

Pimentel, M.M. and Fuck, R.A. (1992). Características geoquímicas e isotópicas de unidades metavulcânicas e ortognáissicas neoproterozóicas do Oeste de Goiás. *Bol. Núcleo Centro-Oeste, SBG, Bol.*, 15, 1-22.

Pimentel, M.M. and Fuck, R.A. (1992). Neoproterozoic crustal accretion in Central Brazil. *Geology*, 20(4), 375-379.

Pimentel, M.M., Dardenne, M.A., Viana, M.G., Costa, S.M., Gioia, L., Junges, S. and Seer, H.J. (1999). Nd isotopes and the provenance of sediments from the Neoproterozoic Brasília Belt, Central Brazil: Geodynamic implications. *South-American Symp. on Isotope Geology* (in press).

Pimentel, M.M., Fuck, R.A. and Botelho, N.F. (1999). Granites and the geodynamic history of the neoproterozoic Brasília belt, Central Brazil: a review. *Lithos*, 46, 463-483.

Pimentel, M.M., Fuck, R.A., Dardenne, M.A., Silva, L.J.H.D and Meneses, P.R. (1995). O magmatismo ácido peraluminoso associado ao Grupo Araxá na região entre Pires do Rio e Ipameri, Goiás: características geoquímicas e implicações geotectônicas. *Res. Exp. 5° Cong. Geol. Centro-Oeste, Goiânia, SBG*, 68-71.

Pimentel, M.M., Heaman, L. and Fuck, R.A. (1991). U-Pb zircon and sphene geochronology of late Proterozoic volcanic arc rock units from southwestern Goiás, Central Brazil. *Journal of South American Earth Sciences*, 4, 329-339.

Pimentel, M.M., Heaman, L. and Fuck, R.A. (1992). Idade do metariolito da Sequência Maratá, Grupo Araxá, Goiás: estudo geocronológico pelos métodos U-Pb em zircão, Rb-Sr e Sm-Nd. *Anais da Academia Brasileira de Ciências*, 64, 19-28.

Pimentel, M.M., Heaman, L., Fuck, R.A. and Marini, O.J. (1991). U-Pb zircon geochronology of Precambrian tin-bearing continental - type acid magmatism in Central Brazil. *Precambrian Research*, 52, 321-335.

Pimentel, M.M., Whitehouse, M.J., Viana, M.G., Fuck, R.A. and Machado, N. (1997). The Mara Rosa arc in the Tocantins Province: further evidence for Neoproterozoic crustal accretion in Central Brazil. *Precambrian Research*, 81, 299-310.

Pinho, J.M.M. (1990). *Evolução tectônica da mineralização de zinco de Vazante.* MSc Diss., IG/UnB, Brasília, 115 p. (unpublished).

Pinho, J.M.M. and Dardenne, M.A. (1989). Evolução tectônica da mineralização de zinco de Vazante, MG. *An. 5° Simp. Geol. Minas Gerais, Belo Horizonte, SBG, Bol.*, 10, 275-276.

Pinho, J.M.M. and Dardenne, M.A. (1993). Deformação do Grupo Canastra na região de Coromandel, MG. *An. 4° Simp. Nac. Est. Tect., Belo Horizonte, SBG, Bol.*, 12, 288-291.

Pinho, J.M.M., Dardenne, M.A. and Rigobello, A.E. (1990). Caracterização da movimentação transcorrente da falha Vazante, MG. *An. 36° Cong. Bras. Geol., Natal, SBG*, 5, 2284-2295.

Pinho, J.M.M., Dardenne, M.A., Souza, J.C.F. and Rosiére, C.A. (1993). Caracterização da nappe do Grupo Canastra na região da Serra da Paciência, Oeste de Minas Gerais. *An. 4° Simp. Nac. Est. Tect., Belo Horizonte, SBG, Bol.*, 12, 277-280.

Pires, F.R.M. and Miano, S.C. (1994). O depósito estanífero de Ipameri, Goiás, discussões sobre a origem. *Anais 4° Simp. Geol. Centro-Oeste, Brasília, SBG*, 31-33.

Queiroz, C.L., Jost, H. and Mc Naughton, N.J. (1999). U-Pb Shrimp ages of Crixás granite-greenstone belt terranes: from Archean to Neoproterozoic. *Anais 7° Simp. Nac. Est. Tect., Lençois-Bahia, SBG-ABGP, Sessão* 1, 35-37.

Ribeiro Filho, W. and Teixeira, N.A. (1980). Sequência vulcano-sedimentar da borda entre os complexos de Niquelândia e Cana Brava. *Bol. Núcleo Centro-Oeste, SBG*, 10, 157-177.

Richardson, S.V., Kesler, S.I. and Essene, E.J. (1986). Origin and geochemistry of the Chapada Cu-Au deposit, Goiás, Brazil: a metamorphosed wall-rock porphyry copper deposit. *Economic Geology*, 81, 1884-1898.

Rigobello, A.E., Branquinho, J.A., Dantas, M.G.S., Oliveira, T.F. and Neves Filho, W. (1988). Mina de zinco de Vazante. In: *Principais Depósitos Minerais do Brasil*, 3, eds. Schobbenhaus, C. and Coelho, C.E.S., pp. 101-110. DNPM, Brasília.

Rimann, E. (1917). Kimberlita no Brasil. *R.E.M.*, 15, 27-32.

Romagna, G. and Costa, R.R. (1988). Jazida de zinco e chumbo de Morro Agudo, Paracatu, Minas Gerais. In: *Principais Depósitos*

Minerais do Brasil, 3, eds. Schobbenhaus, C. and Coelho, C.E.S., pp. 83-90. DNPM, Brasília.

Roscoe, J.S. and Araújo Filho, J.O. (1994). Estudo estrutural preliminar do front de empurrão do Araxá sobre o Paranoá na porção oriental da megainflexão dos Pirineus, Goiás Central. *An. 4º Simp. Geol. Centro-Oeste, Brasília, SBG*, 177-180.

Roscoe, J.S. and Araújo Filho, J.O. (1994). Estudo estrutural preliminar do front de empurrão do Araxá sobre o Paranoá na porção oriental da megainflexão dos Pirineus, Goiás Central. *Anais 4º Simp. Geol. Centro-Oeste, Brasília, SBG*, 177-180.

Rossi, P., Andrade, G.F. and Gaspar, J.C. (1996). The Peixe (Tocantins, Brazil) mesoproterozoic corundum-bearing syenite pegmatite complex: age, considerations on its origin and regional significance. *Anais 39º Cong. Bras. Geol., Salvador, SBG*, 6, 426-429.

Santos, R.V., Alvarenga, J.S., Dardenne, M.A., Sial, A.M. and Ferreira, V.P. (1996). Isótopos de carbono e oxigênio em carbonatos do Grupo Bambuí na Serra de São Domingos, GO. *Anais 39º Cong. Bras. Geol., Salvador, SBG*, 5, 516-520.

Schmidt, W. and Fleischer, R. (1978). Estilo estrutural do pré-cambriano no sudoeste de Minas Gerais. *An. 30º Cong. Bras. Geol., Recife, SBG*, 1, 431-434.

Schöll, W.U. (1973). Sedimentologie der Bambui Gruppe in SE-teil des São Francisco Beckens (Minas Gerais, Brasilien). *Münstersche Forschungen zur Geologie und Paläontologie*, 31-32, 71-91.

Seer, H.J. (1995). *Geologia, deformação e mineralização de cobre no complexo vulcano-sedimentar de Bom Jardim de Goiás*. MSc Diss., IG/UnB, Brasília, 181 p. (unpublished).

Seer, H.J. (1999). *Evolução tectônica dos grupos Araxá, Ibiá e Canastra na sinforma Araxá, Minas Gerais*. Doctorate Thesis, IG/UnB, Brasília, 267 p. (unpublished).

Seer, H.J., Dardenne, M.A. and Fonseca, M.A. (1998). Deformation and tectonic framework of the meso/neoproterozoic units in the southern Brasilia Fold Belt: implications on the geotectonic evolution. *Abstracts 14º Int. Conf. on Basem. Tectonics, Ouro Preto, EM – DEGEO – UFOP*, 57-61.

Seer, H.J., Dardenne, M.A. and Fonseca, M.A. (1999). O Grupo Ibiá na sinforma de Araxá: um terreno tectonoestratigráfico ligado à evolução de arcos magmáticos. *An. 7º Simp. Geol. Centro-Oeste, Brasília, SBG* (in press).

Silva, J.A. and Sá, J.A.G. (1988). Jazida de cobre de Chapada, Mara Rosa, Goiás. In: *Principais Depósitos Minerais do Brasil*, 3, eds. Schobbenhaus, C. and Coelho, C.E.S., pp. 55-60. DNPM, Brasília.

Simões, L.S.A. (1991). Dobras em bainhas macroscópicas relacionadas ao cisalhamento dúctil de baixo ângulo da nappe Araxá-Canastra no sudoeste de Minas Gerais. *Anais Simp. Nac. Est. Tect., Rio Claro, SBG*, 11-13.

Simões, L.S.A. (1995). *Evolução tectono-metamórfica da nappe de Passos, sudoeste de Minas Gerais*. Doctorate Thesis, IG/USP, São Paulo, 149 p. (unpublished).

Simões, L.S.A. and Navarro, G. (1996). Estruturação da Faixa Brasília na região de Araxá, MG. *Anais 39º Cong. Bras. Geol., Salvador, SBG*, 92-95.

Simões, L.S.A. and Valeriano, C.M. (1990). Porção meridional da Faixa Brasília: estágio atual do conhecimento e problemas de correlação tectono-estratigráfica. *Anais 36º Cong. Bras. Geol., Natal, SBG*, 6, 2564-2575.

Souza, J.C.F. (1997). *Litoestratigrafia e sedimentologia da Formação Vazante na região de Coromandel, MG*. MSc Diss. Thesis, IG/UnB, Brasília, 75 p. (unpublished).

Strieder, A.J. (1990). Análise estrutural da região de Abadiânia (GO). *Revista Brasileira de Geociências*, 20, 239-257.

Strieder, A.J. (1993). Tectônica colisional no Brasil Central: evolução e implicações no padrão estrutural. *Anais 4º Simp. Nac. Est. Tect., Belo Horizonte, SBG*, 297-301.

Strieder, A.J. (1994). *Deformação e metamorfismo na região de Santa Cruz de Goiás. Correlação tectono-estratigráfica e evolução tectônica regional*. Doctorate Thesis, IG/UnB, Brasília, 258 p. (unpublished).

Strieder, A.J. and Nilson, A.A. (1992). Melange ofiolítica nos metassedimentos Araxá de Abadiânia (GO) e implicações tectônicas regionais. *Revista Brasileira de Geociências*, 22(2), 204-215.

Suita, M.T.F. (1996). *Geoquímica e metalogenia de elementos do grupo da platina EPG + Au) em complexos máfico-ultramáficos do Brasil: critérios e guias com ênfase no complexo máfico-ultramáfico acamadado de alto grau de Barro Alto (CBA, Goiás)*. Doctorate Thesis, UFRGS, Porto Alegre, 525 p. (unpublished).

Suita, M.T.F. (1998). Late Paleo-Neoproterozoic PGE + Au metallogeny of giant layered high-grade mafic ultramafic intrusions (Barro Alto and Niquelândia), Tocantins Province, Goiás, Central Brazil. *Abstr. 14th Int. Conf. on Basement Tectonics, Ouro Preto, MG-Brazil*, 179-180.

Teixeira, N.A. and Danni, J.C.M. (1978). Contribuição à estratigrafia do Grupo Araxá na região de Passos, MG. *An. 30º Cong. Bras. Geol., Recife, SBG*, 2, 700-710.

Thomaz Filho, A. and Lima, V.Q. (1981). Datação radiométrica de rochas pelíticas pelo método Rb-Sr. *Bol. PETROBRÁS*, 24, 109-119.

Thomaz Filho, A., Kawashita, K. and Cordani, U.G. (1998). A origem do Grupo Bambuí no contexto da evolução geotectônica e de idades radiométricas. *Anais da Academia Brasileira de Ciências*, 70(3), 527-548.

Thorman, C.H. and Nahas, S. (1979). Reconnaissance geological study of the Vazante zinc district, Minas Gerais, Brazil. *USGS Prof. Paper*, 1126-I, 12 p.

Valeriano, C.M. (1992). *Evolução tectônica da extremidade meridional da Faixa Brasília, região da represa de Furnas, sudoeste de Minas Gerais*. Doctorate Thesis, IG/USP, São Paulo, 198 p. (unpublished).

Valeriano, C.M. and Simões, L.S.A (1997). Geochemistry of proterozoic mafic rocks from the Passos Nappe (Minas Gerais, Brazil): tectonic implications to the evolution of the southern Brasília Belt. *Revista Brasileira de Geociências*, 27(1), 99-110.

Valeriano, C.M., Almeida, J.C.H., Simões, L.S.A, Duarte, P.B., Roig, H.L. and Heilbron, M. (1997). Evolução estrutural do domínio externo da Faixa Brasília no sudoeste de Minas Gerais: registros de uma tectônica pré-brasiliana. *Revista Brasileira de Geociências* (in press).

Viana, M.G., Pimentel, M.M., Whitehouse, M.J., Fuck ,R.A. and Machado, N. (1996) O Arco Magmático de Mara Rosa, Goiás: dados geoquímicos e geocronológicos e suas implicações regionais. *Revista Brasileira de Geociências*, 25, 111-123.

Winge, M. (1995) Idade da supersuíte Americano do Brasil. *An. 5º Simp. Geol. Centro-Oeste, SBG, Goiânia*, 99-101.

Evolution of the Araçuaí Belt and its Connection to the Ribeira Belt, Eastern Brazil

Antônio Carlos Pedrosa-Soares and Cristina Maria Wiedemann-Leonardos

This chapter focuses on the mobile belt extending over the region from the eastern boundary of the São Francisco Craton to the Atlantic Ocean, named the Araçuai Belt (Almeida, 1977), and its connection to the Ribeira Belt (Almeida *et al.*, 1973), in eastern Brazil (Fig. 1). The later belt is also called the Atlantic (Leonardos and Fyfe, 1974) or Coastal Mobile Belt (Wiedemann, 1993).

In the early fifties, Rosier and Ebert recognized an orogenic belt along the coastline of Rio de Janeiro (Ebert, 1968); the Paraíba Belt, named after the prominent Paraíba do Sul River. According to their observations, this belt consists mainly of metasediments such as paragneiss, quartzite, and marble, metamorphosed in the amphibolite facies. In 1974, Leonardos and Fyfe included the Paraíba Belt of Ebert in the Atlantic Belt, outlined as a low pressure-high temperature metamorphic belt with common signs of ultrametamorphism and partial melting. The Ribeira Belt of Almeida *et al.* (1973) and Hasui *et al.* (1975), characterized by a sheaf of NE-trending strike-slip shear zones, also included the Paraíba Belt.

Almeida (1977) defined the southeastern limit of the São Francisco Craton as being the Neoproterozoic arch-shaped Araçuai Fold Belt. He accurately outlined the boundary between the craton and the belt, but the eastern and southern limits of the Araçuai Belt remained undefined. Nevertheless, the correlation of the Araçuai and West-Congo belts is well established in the geological literature (Brito-Neves and Cordani, 1991; Pedrosa-Soares *et al.*, 1992, 1998; Trompette, 1994, 1997). Both belts are counterparts of the Araçuaí-West-Congo Orogen, now separated by the Atlantic Ocean (Figs. 2 and 3). Consequently, we use the name Araçuaí Belt to designate the entire Brazilian portion of the Araçuaí-West Congo Orogen.

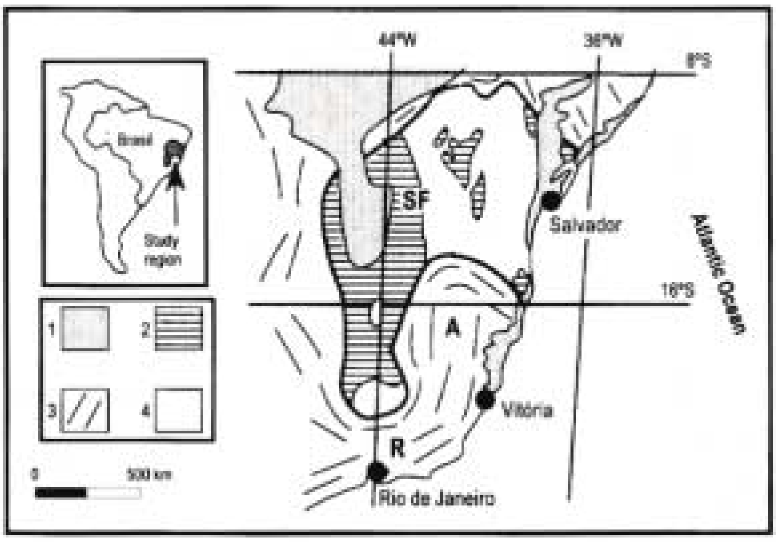
Araçuaí and Ribeira belts are the prevalent names in the current geological literature, although used in different senses. For instance, several authors call the internal orogenic domain of the Araçuaí Belt as the northern Ribeira Belt, while others use the name Ribeira to designate the southern part of the Araçuaí Belt. We do not regard Araçuai and Ribeira as separated parallel belts. As discussed bellow, they are understood as longitudinal segments connected along a tectonic trend of the same orogenic system.

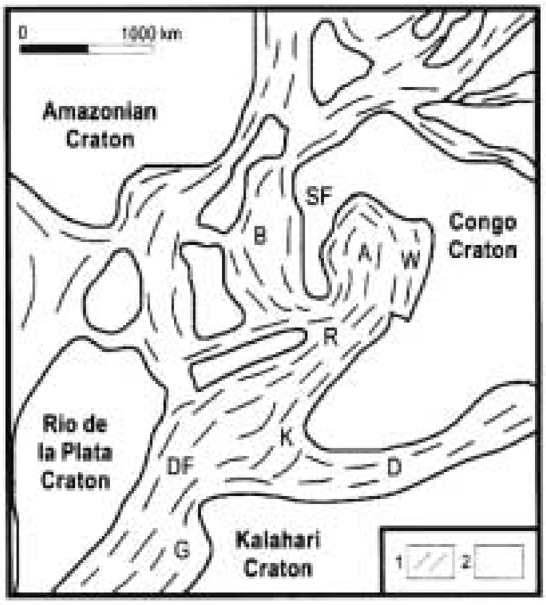
The combined examination of geological, geochemical, geophysical and geochronological data led us to a more complete interpretation of the Araçuaí-West-Congo Orogen (AWCO) as a continuous geotectonic unit. In the southern limits, the evolution of the Araçuaí Belt will be discussed together with its connection to the Ribeira Belt.

Araçuai Belt limits, tectonic domains and its connection to the Ribeira Belt

Craton and orogen are relative concepts in terms of the tectonic behavior of adjacent lithosphere sectors, during a specific time interval. The borders of rigid (cratonic) sectors

FIGURE 1 - Location of the Araçuaí (A) and Ribeira (R) belts in relation to the São Francisco Craton (SF). 1 - Phanerozoic cover; 2 - Neoproterozoic cratonic cover; 3 - structural trend of Neoproterozoic mobile belts; and 4 - cratonic basement, including late Paleoproterozoic and Mesoproterozoic rocks.





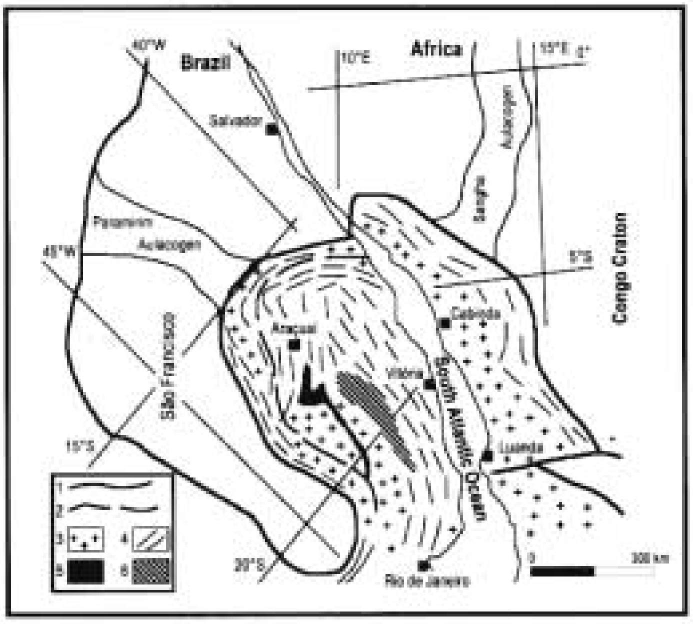


FIGURE 2 - The Neoproterozoic Araçuaí (A), Ribeira (R), and West-Congo (W) belts in the geotectonic framework of western Gondwana (modified after Brito-Neves and Cordani, 1991). 1 - structural trend of Neoproterozoic mobile belts (B, Brasília; D, Damara; DF, Dom Feliciano; G, Gariep; K, Kaoko), and 2 - major cratons (SF, São Francisco) and other landmasses.

FIGURE 3 - Tectonic sketch map of the Araçuaí-West-Congo Orogen in a predrift reconstruction (modified after Trompette, 1994). 1 - cratonic boundary; 2 - limit between external and internal tectonic domains; 3 - Archean-Paleoproterozoic basement reworked during Brasiliano-Pan African orogeny; 4 - Neoproterozoic structural trend outlined by the regional ductile foliation; 5 - zone of slivers of Neoproterozoic oceanic lithosphere; 6 - zone of remnants of the Neoproterozoic precollisional to syncollisional magmatic arc (calc-alkaline G1 plutons).

FIGURE 4 - Geological map of the Araçuaí Belt and cratonic surroundings, highligthing the Neoproterozoic units (after Pinto et al., 1998, and Pedrosa-Soares et al., 1999b).
1 - Phanerozoic covers. Late Cambrian granitoid suites:
2 - I-type G5 (black dots in plutons with mafic cores); 3 - S-type G4. Late Neoproterozoic to Cambrian granitoid suites:
4 - I-type G3-I; 5 - S-type G3-S and G2; 6 - I-type G1. Neoproterozoic: 7 - alkaline intrusions in cratonic region;
8 - sedimentary cratonic covers; 9 - Capelinha Formation;
10 - Macaúbas Group proximal unit; Macaúbas Group distal unit [11 - Salinas Formation metasedimentary unit;
12 - Salinas Formation metavolcanic-sedimentary unit (correlated with Dom Silvério Group, ds), (u, location of meta-ultramafic bodies)]; 13 - Rio Doce Group;
14 - Jequitinhonha Complex; 15 - high-amphibolite facies domain of Paraíba do Sul Complex; 16 - granulite facies domain of Paraíba do Sul Complex. Late Paleoproterozoic and Mesoproterozoic: 17 - Espinhaço Supergroup;
18 - Borrachudos granitoid suite. Paleoproterozoic and Archean: 19 - Juiz de Fora Complex; 20 - TTG complexes, with greenstone belts remnants (g) and metasedimentary units (m). 21 - thrust and detachment faults or ductile shear zones;
22 - oblique to strike-slip faults or ductile shear zones;
23 - other faults. Rivers: Do - Doce; Je - Jequitinhonha; Mu - Mucuri; Pa - Pardo. Towns: A - Araçuaí; BH - Belo Horizonte; CS - Cândido Sales; D - Diamantina; GV - Governador Valadares; M - Manhuaçu; MC - Montes Claros; N - Nanuque; TO - Teófilo Otoni; S - Salinas; SD - Salto da Divisa; V - Vitória.

resisting compressional deformation fix the limits of orogens and their coeval cratons (Sengor, 1990; Brito-Neves and Alkmim, 1993). Orogens may be subdivided in external (or proximal) and internal (or distal) tectonic domains with different tectono-metamorphic-magmatic frames. One or more tectonic domains may constitute a mobile (orogenic) belt. In terms of structural and metamorphic features, the external tectonic domains are generally fold-and-thrust belts, and the internal tectonic domain correspond to the metamorphic-anatetic (or "crystalline") core of the orogen (*e.g.*,Condie, 1993; Park, 1993). In this chapter, we describe the focused orogenic belts in terms of their external and internal tectonic domains.

In the Neoproterozoic geotectonic scenario of Western Gondwana (Fig. 2), the Araçuaí-West-Congo Orogen occupies a peculiar place, confined in a cratonic embayment. For that reason, the embayment outlined by the São Francisco and Congo cratons clearly defines the western, northern and eastern AWCO limits (Almeida, 1977; Alkmim *et al.*, 1993; Trompette, 1994). In fact, the Araçuaí-West Congo Orogen is the northern branch of the Brasiliano-Pan-African orogenic system that includes the Ribeira, Kaoko, Dom Feliciano, Damara and Gariep belts (Fig. 2). However, the southern boundary of the Araçuaí Belt (*i.e.*, the AWCO's Brazilian portion) has been only vaguely referred to.

Taking into account the concepts of craton and orogen, the southeastern toe of the São Francisco Craton defines the AWCO's southern limit in Brazil, as proposed by Pedrosa-Soares and Noce (1998, 1999). Along the major portion of the Araçuaí Belt, the structural trend follows the NNE strike. At this southern boundary, the structural trend inflects from NNE to NE. This is the trend of the Ribeira Belt from 21°S southwards. Therefore, the 21°S parallel roughly marks the southern limit of the Araçuaí Belt (Figs. 1 and 3). This gentle inflection is also


São Francisco Craton
São Francisco Craton
N
42ºW
40ºW
15ºS
16ºS
17ºS
18ºS
19ºS
20ºS
21ºS
Pa
CS
S
Je
MC
A
TO
N
Mu
D
GV
Do
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Atlantic Ocean
0
50 km
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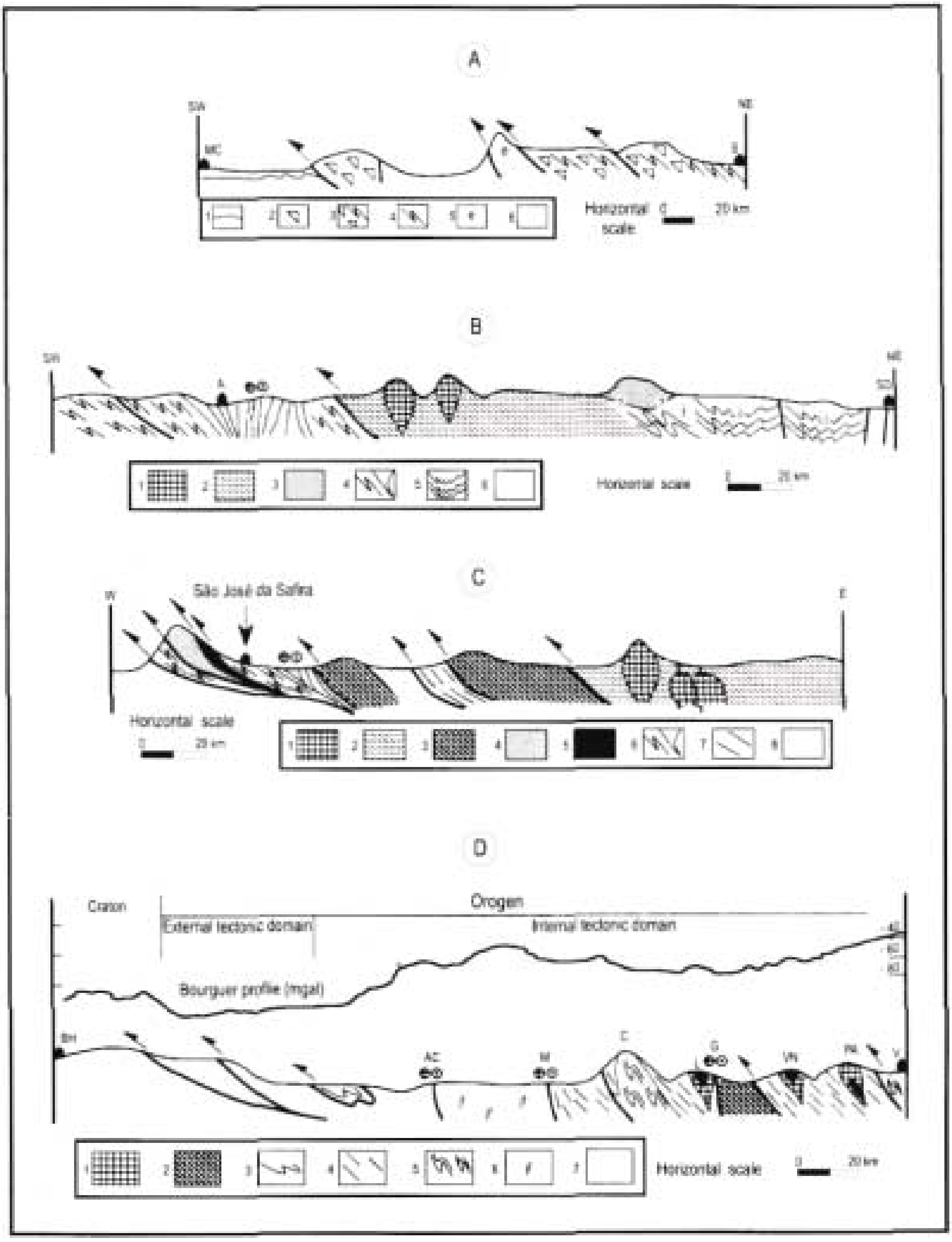


FIGURE 5 - Geological sections and a Bouguer profile cross-cutting the Araçuaí Belt. A) The external tectonic domain along the Montes Claros (MC) to Salinas (S) section. Neoproterozoic: 1 - pelite-carbonate cratonic cover; Macaúbas Group western formations: 2 - diamictite with minor intercalations of sandstone, metamorphosed in low greenschist facies; 3 - diamictite, sedimentary iron formation, sandstone, pelite, metamorphosed in greenschist to low amphibolite facies; Salinas Formation: 4 - greywacke and pelite, metamorphosed in low amphibolite facies. Older units: 5 - Espinhaço Supergroup; 6 - gneissic basement. B) The northern subdomain of the internal tectonic domain and eastern arm of the northern curvature in a section from the Araçuaí region (A) to Salto da Divisa (SD). Late Cambrian: 1 - G5 granitoid suite. Neoproterozoic: 2 - G2 and G3-S granitic suites; 3 - Capelinha Formation (quartzite, graphitic metapelite); 4 - Salinas Formation metasedimentary unit, outlining a dextral, positive flower-structure (f); 5 - Jequitinhonha Complex (kinzigitic gneiss, calc-silicate granulite). Paleoproterozoic: 6 - gneissic basement (Itabuna Belt). C) The northern subdomain of the internal tectonic domain in a section along 18°15'S, eastwards from 42°20'W (modified after Pedrosa-Soares et al., 1999b). Late Cambrian: 1 - G5 granitoid suite. Neoproterozoic: 2 - G2 and G3-S granitic suites; 3 - G1 granitoid suite (calc-alkaline orthogneiss); 4 - Capelinha Formation (mainly quartzite); 5 - tectonic slabs of meta-ultramafic rocks; 6 - Salinas Formation metavolcanic-sedimentary unit, outlining a positive flower-structure (f); 7 - Rio Doce Group. Older units: 8 - gneissic basement. D) External and internal tectonic domains along the Belo Horizonte (BH) to Vitória (V) section (modified after Wiedemann et al., in prep.). Cambro-Ordovician: 1 - G5 granitoid suite (L, Lajinha; PA, Pedra Azul; and VN, Venda Nova plutons). Neoproterozoic: 2 - G1 granitoid suite; 3 - volcanic-sedimentary Dom Silvério Group; 4 - Rio Doce Group and Paraíba do Sul Complex (amphibolite facies domain); 5 - Paraíba do Sul Complex (granulite facies domain; C, Caparaó ridge). Older units: 6 - Juiz de Fora Complex (granulite and high amphibolite facies); 7 - Archean and Paleoproterozoic units. Major steeply dipping, dextral shear zones: AC, Abre Campo; G, Guaçuí; M, Manhuaçu. In the Bouguer profile, note the negative gravity anomaly indicating crustal thickening in the external tectonic domain, and the striking increase in gravity values from the zone of preservation of Neoproterozoic oceanic slivers to the Juiz de Fora Complex.

clearly traced by magnetic and gravity anomalies (Ussami *et al.*, 1993; Ebert and Hasui, 1998). As a result, we call Araçuaí the mobile belt extending over the region from the eastern edge of the São Francisco Craton to the Atlantic Ocean, between the parallels 15°S and 21°S (Figs. 1 and 4).

Concerning the Brasiliano-Pan-African frame, the general features of the Araçuaí-West-Congo Orogen were synthesized from descriptions of Marshak and Alkmim, 1989; Uhlein, 1991; Pedrosa-Soares *et al.*, 1992, 1993, 1996, 1998, 1999a,b; Alkmim *et al.*, 1993, 1996; Costa *et al.*, 1993, 1998; Crocco-Rodrigues *et al.*, 1993; Wiedemann, 1993; Trompette, 1994, 1997; Campos-Neto and Figueiredo, 1995; D'El-Rey-Silva *et al.*, 1995; Pedrosa-Soares and Pedreira, 1996; Costa, 1989, 1998; Cunningham *et al.*, 1996, 1998; Grossi-Sad *et al.*, 1997; Nalini, 1997; Oliveira *et al.*, 1997; Pinto *et al.*, 1997; Wiedemann *et al.*, 1997; Fishel, 1998; Uhlein *et al.*, 1998a.

The following tectonic domains are highlighted:

- The external tectonic domain of the orogen is a symmetrical, arch-shaped feature (Fig. 3). It consists of the external tectonic domain of the Araçuaí Belt and the West-Congo Belt, linked through the northern curvature traced by the Araçuaí Belt. The external domain connects two partially inverted Neoproterozoic aulacogens, named Paramirim in Brazil (Schobbenhaus, 1996) and Sangha in Africa (Trompette, 1994). The opposite sides of the external domain show centrifugal vergences, outlined by shallow to moderately-dipping structures such as the regional ductile foliation, frontal thrusts and axial surfaces of asymmetric folds. The tectonic transport vectors point towards both the São Francisco (W vergence, Figs. 5a-d) and Congo (E vergence) cratons. Neoproterozoic sequences show metamorphism grading from low greenschist facies, at the cratonic edges to low amphibolite facies.

- The AWCO's internal tectonic domain is confined to the Araçuaí Belt. NNE to N-S structural trend dominates along this domain, being intersected by the bent structural trend that defines the northern sector of the Araçuaí Belt (Fig. 3). Neoproterozoic to Cambrian granitic plutonism is widespread along the internal tectonic domain. Two distinct subdomains are recognized in the internal domain. A northern subdomain (N of 19°S) shows well marked SW vergence formed by shallow to moderately-dipping oblique thrusts with dextral strike-component, where the anatectic zone of the orogen is better exposed (Fig. 5b - c). Prominent steeply dipping, dextral shear zones, granulite facies rocks and granitic plutons with mafic cores characterize a southern subdomain. This subdomain preserves consequently the deepest crustal levels of the orogen (Fig. 5d).

- Both the external and internal tectonic domains are represented in the northern curvature of the Araçuaí Belt. Along that arch-shaped zone the strike of the regional ductile foliation (schistosity or gneissic foliation) changes from NE, in the western region, to E-W along the central part, and to NW in the eastern arm of the curvature (Fig. 3). In general, this foliation dips steeply and displays an oblique to strike stretching lineation (Fig. 5b). The vergence toward the craton is only well defined in the neighboring cratonic covers.

Despite several local truncations, when the Brasiliano-Pan-African orogenic belt is regarded as a whole, progressive metamorphic assemblages record increasing temperatures: from W to E in the N-S segment of the Araçuaí Belt; from N to S in the northern curvature of this belt; and from E to W in the West-Congo Belt (Almeida *et al.*, 1978; Pedrosa-Soares *et al.*, 1992, 1999a; Trompette, 1994; Pedrosa-Soares and Pedreira, 1996; Carvalho and Pereira, 1997a; Pinto *et al.*, 1997; Cunningham *et al.*, 1998; Söllner *et al.*, in press). Therefore, the regional metamorphism shows a centripetal pattern, with increasing temperatures converging from the cratonic borders to the AWCO's anatectic-granulitic core. After all, these metamorphic assemblages continue along the strike, without any gap, from the southern part of the Araçuaí Belt into the Ribeira Belt (*e.g.*, Leonardos and Fyfe, 1974; Hasui and Oliveira, 1984; Campos-Neto and Figueiredo, 1995; Heilbron *et al.*, 1995).

The Neoproterozoic mobile belt that begins at 21 °S parallel and extends southwards, following the NE-SW trend to around 25 °S, is called Ribeira Belt. Because no metamorphic, structural or geophysical discontinuities were found in that Araçuaí-Ribeira boundary (Hasui and Oliveira, 1984; Ussami *et al.*, 1993; Wiedemann, 1993; Campos-Neto and Figueiredo, 1995; Heilbron *et al.*, 1995; Machado *et al.*, 1996; CPRM-SGA, 1997; Ebert and Hasui, 1998), the Ribeira Belt is thought to be longitudinally connected to the Araçuaí Belt. Despite being located outside the São Francisco-Congo embayment, this connection runs along the tectonic trend of the orogenic system (Figs. 2 and 3). The southern tectonic domains of the Araçuaí Belt extend into the northernmost Ribeira Belt, from approximately 43°30'W to the Atlantic coastline between the parallels 21 °S and 22 °S, where they show the following main characteristics (Hasui and Oliveira, 1984; Wiedemann, 1993; Campos-Neto and Figueiredo, 1995; Heilbron *et al.*, 1995, 1999; Figueiredo and Teixeira, 1996; Alkmim and Marshak, 1998; Ebert and Hasui, 1998; Fonseca, 1998; Campos-Neto *et al.*, 1999):

- The external tectonic domain is narrow and displays a gently bent structural trend bordering the southeastern toe of the São Francisco Craton. Rocks are mainly from the cratonic gneissic basement reworked by the Brasiliano NW vergent thrust tectonics.

- The internal tectonic domain comprises a western zone formed by granulite stacked with high amphibolite facies metasediments, flanked by the narrow coastal granulite complex. Prominent steeply dipping, oblique to transcurrent dextral shears zones trend NE and partially overprint the NW vergent thrust system.

Although the Ribeira and Araçuaí belts are segments of the same orogenic system, they are marked by some important differences, such as:

- The northern part of the Ribeira Belt shows fewer supracrustal greenschist to low-amphibolite metamorphic sequences, and more abundant high-amphibolite to granulite facies rocks than the Araçuaí Belt, implying deeper crustal levels along the former.

- The Ribeira Belt is strikingly marked by NE trending, dextral transcurrent-transpressive shear zones (Lammerer, 1987; Fritzer, 1991; Heilbron *et al.*, 1995, 1999; Ebert and Hasui, 1998). Although these zones continue farther into the southern Araçuaí Belt, thrust tectonics remains the most important regime in the Araçuaí frame.

- The African counterpart of the Ribeira Belt seems to be exclusively composed of pre-Neoproterozoic rocks, the polycyclic basement (Trompette, 1994). In fact, it is probable that the eastern terranes, at least the "Cabo Frio fragment" (Fonseca, 1998), of the Ribeira Belt represent the Congo Craton border strongly reworked during the Brasiliano-Pan-African collage.

In this scenario (Fig. 2), in comparison with the Araçuaí Belt, the Ribeira Belt has certainly resulted from a more complex tectonic process, involving convergence of at least one more rigid landmass against the São Francisco and Congo cratons (Brito-Neves and Cordani, 1991; Ebert and Hasui, 1998). Nevertheless, the northernmost segment of the Ribeira Belt seems to be developed in response to the convergent interaction of the São Francisco Craton and a landmass located along the coastal region of northern Rio de Janeiro - southern Espírito Santo (Wiedemann, 1993, Heilbron *et al.*, 1999). Although some authors consider this landmass a microplate (Campos-Neto and Figueiredo, 1995), it could be that the Congo Craton southwestern border is a piece of it. In this case, the Cabo Frio fragment is a remnant of this landmass, left in Brazil after the South Atlantic opening (Trompette, 1994).

Basement of the Araçuaí and Northern Ribeira belts

In the last eight years, geological mapping projects carried out by Universidade Federal de Minas Gerais, Universidade Federal do Rio de Janeiro, Companhia Mineradora de Minas Gerais and the Serviço Geológico do Brasil made possible the compilation of a regional map of the Araçuaí Belt (Fig. 4). Now, this belt is almost entirely mapped at 1:100 000 scale. Many segments were studied in greater detail.

Along the external tectonic domain, the basement consists mainly of Archean TTG gneissic complexes with remnants of greenstone belts, Paleoproterozoic supracrustal sequences and associated granitoid suites (Fig. 4). These units were affected by the Transamazonian Orogeny (*c.* 2.2 - 2.0 Ga), and were later reworked during the Brasiliano Orogeny (*e.g.*, Marshak and Alkmim, 1989; Dossin 1994; Figueiredo and Teixeira, 1996; Alkmim and Marshak, 1998; Noce *et al.*, 1998, 1999). Paleoproterozoic gneiss from the Itabuna Belt limits the eastern edge of the Araçuaí Belt northern curvature. This Itabuna Belt developed during the Transamazonian-Eburneam Orogeny (*e.g.*, Trompette, 1994; Barbosa and Dominguez, 1996).

In the internal tectonic domain, the Archean-Paleoproterozoic basement consists of high-amphibolite gneiss and granulite complexes. These rocks are best exposed in the southern Araçuaí and northern Ribeira belts. In the Juiz de Fora Complex, they were predominantly originated from arc-related magmatic protoliths of Paleoproterozoic age (*c.* 2.2 - 2.0 Ga) or represent the reworked Archean basement (Figueiredo and Teixeira, 1996; Machado *et al.*, 1996; Costa, 1998; Cunningham *et al.*, 1998; Fischel, 1998). Therefore, the Juiz de Fora Complex also records the roots of a Transamazonian magmatic arc, probably once linked to the Itabuna Belt (Alkmim and Marshak, 1998).

The late Paleoproterozoic to Mesoproterozoic units are the Espinhaço Supergroup and the Borrachudos Suite. The Espinhaço Supergroup was deposited in a long-lived rift-sag basin, which was successively filled with volcano-sedimentary and sedimentary units (Almeida-Abreu, 1993; Martins-Neto, 1998a; Uhlein *et al.*, 1998a). The bimodal volcanics yielded crystallization ages at about 1.75 - 1.7 Ga (Machado *et al.*, 1989; Dossin *et al.*, 1993). The Borrachudos Suite (*c.* 1.77 - 1.7 Ga) represents the anorogenic, granitic magmatism related to the Espinhaço continental rift (Dussin and Dussin, 1995; Fernandes *et al.*, 1999a,b). This suite consists of deformed granites, exhibiting a gneissic foliation and local evidences of migmatization. Zircon U/Pb data from a Borrachudos Pluton indicate that this regional deformation and metamorphic recrystallization took place around 620 Ma. Sphene U/Pb data from the same pluton are concordant at 507 Ma, suggesting that a regional reheating was induced by the significant postcollisional magmatism of the Araçuaí Belt (Fernandes *et al.*, 1999a,b). These U/Pb data suggest the Borrachudos Suite and the Espinhaço Supergroup only underwent the Brasiliano Orogeny, in the Araçuaí Belt.

Neoproterozoic rift and passive margin sedimentation

An extensional event, recorded by the intrusion of basic dyke swarms and anorogenic granite in the São Francisco-Congo paleocontinental region, started at the Mesoproterozoic-Neoproterozoic boundary. A continental rift opened in response to the rising of a mantle plume (Correa-Gomes and Oliveira, 1997; Martins-Neto, 1998b). Magmatic rocks related to this stage yielded ages from *c.* 1.05 Ga to 900 Ma (Machado *et al.*, 1989; D'Agrella *et al.*, 1990; Djama *et al.*, 1992; Vicat and Pouclet, 1995; Paes *et al.*, 1999). Widespread glacial deposits mark the Neoproterozoic sedimentation in the Araçuaí-West-Congo basin system and in the São Francisco-Congo paleocontinental region (Karfunkel and Hoppe, 1988; Trompette, 1994; Uhlein *et al.*, 1998a).

The Macaúbas Group is the glaciation-related Neoproterozoic unit in the Araçuaí Belt and records all major stages of a basin evolved from a continental rift to a passive margin (Pedrosa-Soares *et al.*, 1992, 1998; Noce *et al.*, 1997; Uhlein *et al.*, 1998a). In Figure 4, the Macaúbas Group formations have been divided into proximal and distal units. Detailed descriptions and synthesis on the Macaúbas Group are found in Karfunkel and Hoppe (1988), Pedrosa-Soares *et al.* (1992, 1998, 1999b,c), Trompette (1994), Barbosa and Dominguez (1996), Almeida-Abreu *et al.*, (1997), Grossi-Sad *et al.*, (1997), Noce *et al.*, (1997), and Ulhein *et al.*, (1998a).

The Macaúbas Group proximal unit includes pre-glacial and glaciation-related deposits. The preglacial sequences record the continental rift stage of the Macaúbas Basin and consist of sandstone, conglomerate, pelite and scarce dolomite beds, which were deposited in fluvial to shallow marine environments. The glaciation-related deposits consist of diamictite, sandstone and pelite. Volcanic rocks are also found in the Macaúbas proximal unit. Deposition of extensive glaciomarine sequences marks the transition from rift to passive margin stages. These deposits contain diamictite interfingered with graded sandstone, sedimentary iron formation units and sandy-pelitic rhythmite, which were deposited by debris flows and turbidity currents. The presence of scattered clasts up to one-metre in size, enveloped by sandy-pelitic sediments, is the evidence of iceberg discharge. In the Macaúbas proximal unit, metamorphism grades from low to high greenschist facies.

U/Pb ages of detrital zircon constrain the age of the glaciomarine sedimentation to a maximum at about 950 Ma (Pedrosa-Soares *et al.*, 1999c). Moreover, the glacial sedimentation started sometime between 950 and 900 Ma, if the dykes referred to by Almeida-Abreu *et al.* (1997), that are cutting proximal diamictite beds, actually belong to the 900 Ma age swarm.

The Macaúbas Group distal unit is the Salinas Formation (Fig. 4). This formation only records passive margin sedimentation, which represents a transgressive sequence deposited after the end of glaciation. The Salinas Formation is devoid of diamictite, and includes a turbiditic deep-sea sand-mud unit and a volcano-sedimentary unit. The first consists of metamorphosed greywacke pelite and quartzose greywacke, with minor intercalation of carbonaceous pelite, marl (calc-silicate rock), limestone, and clast-supported conglomerate. The distal volcano-sedimentary unit (Ribeirão da Folha facies) includes metamorphosed deep-sea pelite, chert, massive sulphide, banded iron formation units and ocean-floor basalt. The age yielded by this ocean-floor basalt constrains the timing of the passive margin stage at about 800 Ma. In the Salinas Formation, metamorphism grades from low to intermediate amphibolite facies (Pedrosa-Soares *et al.*, 1992, 1996, 1998; Uhlein *et al.*, 1998a).

The Capelinha Formation (Grossi-Sad *et al.*, 1993; Noce *et al.*, 1997) includes a lower greywacke-pelitic unit and an upper sandstone unit, both metamorphosed in the amphibolite facies. This formation may represent sedimentation related to the stage of the basin closure (Pedrosa-Soares, 1995).

The Rio Doce Group consists of a lower carbonate-sandy-pelitic unit and an upper sandstone unit (Pinto *et al.*, 1997). Although the Rio Doce rocks are metamorphosed within greenschist to amphibolite facies, turbidite-like features may be locally recognized in the banded schist (Pedreira *et al.*, 1997). The age of sedimentation of the Rio Doce Group is not available. It was probably deposited in the Neoproterozoic. Its lower unit may be correlated with the Salinas Formation, while the upper unit seems to be a stratigraphic equivalent of the Capelinha Formation.

The Jequitinhonha Complex (Almeida and Litwinski, 1984) comprises a thick kinzigitic pile, consisting of migmatized biotite gneiss with variable contents of garnet, cordierite, sillimanite and graphite, with thick intercalations of graphite-rich gneiss, and minor quartzite, calc-silicate granulite and leptinite (Fig. 4). This rock-assemblage is interpreted as marine arkosic to greywacke pelite, together with carbonaceous-rich pelite (black shale zones), which were deposited as black mud in a confined marine environment under reducing conditions (Faria, 1997).

Samples from the Jequitinhonha Complex yielded Sm/Nd model ages ranging from 1.73 to 1.61 Ga (Celino, 1999) and are in very good agreement with model ages from the Salinas Formation metasediments (1.71 - 1.52 Ga; Pedrosa-Soares *et al.*, in prep.). These Sm/Nd data reinforce the proposal of a stratigraphic correlation between the Salinas Formation and the Jequitinhonha Complex (Siga, 1986; Trompette, 1994). These results also suggest that the protoliths of both the Salinas and the Jequitinhonha pelitic metasediments have a similar mixed origin. They have been supplied from a mixture of Paleoproterozoic and younger sources. Magmatic rocks formed during the Espinhaço and the Araçuaí-West-Congo continental rift stages are good candidates for the younger sources of these sediments. The Jequitinhonha Complex, together with the Salinas Formation are conceived as Neoproterozoic passive margin-type deposits (Pedrosa-Soares *et al.*, 1999b). The probable correlation between the Jequitinhonha and the Paraíba do Sul complexes is discussed below.

The Dom Silvério Group forms a tight refolded syncline of schistose rocks (Brandalise, 1991). It is similar to the distal Salinas Formation. The distinct and continuous magnetic anomaly obtained for these units suggests a stratigraphic correlation between them (CPRM-SGA, 1997). In fact, recent mapping (Klumb and Leite, 1999; Ribeiro, 1999; Silva, 1999) shows the geographic link between these units in the field, as shown in Figure 4. As the Salinas Formation, the Dom Silvério Group also includes a volcano-sedimentary unit metamorphosed in the amphibolite facies, and encloses tectonic slabs of meta-ultramafic rocks (Jordt-Evangelista, 1992; Cunningham *et al.*, 1998). Pelitic schist from the Dom Silvério Group yielded Sm/Nd mineral ages around 549 Ma, indicating Brasiliano metamorphism. On the other hand, whole-rock model age (*c.* 2.2 Ga) obtained for the Dom Silvério schist suggests sedimentary derivation from a Paleoproterozoic source (Brueckner *et al.*, 1998).

The Paraíba do Sul Complex

The Paraíba do Sul Complex (or Group; Ebert 1968) is one of the most widespread units in the internal tectonic domains of both Araçuaí and northern Ribeira belts (Fig. 4). It has been the focus of controversy mainly because of its age, origin and tectonic setting, especially in the geological literature on the Araçuaí Belt.

The joint effort of DNPM (Departamento Nacional da Produção Mineral), CPRM (the Brazilian Geological Survey) and RADAMBRASIL, during the seventies and eighties, brought to light the continuation of the Paraíba do Sul Complex into the Espírito Santo State, to the E and NE of Minas Gerais State up to the SE of Bahia State. This mainly metasedimentary rock-unit was then considered to be exclusively of Paleoproterozoic to Archean age, although geochronological data from eastern and northeastern Minas Gerais and Espírito Santo (Delhal *et al.*, 1969; Lima *et al.*, 1981; Machado *et al.*, 1983; Silva *et al.*, 1987) pointed towards large areas of Neoproterozoic terranes (once called the Embu Complex; *cf.* Schobbenhaus *et al.*, 1981).

The Paraíba do Sul Complex is subdivided into two lithological domains (Fig. 4), both metamorphosed during the Brasiliano Orogeny (Sluitner and Weber-Diefenbach, 1989; Sluitner, 1990; Féboli, 1993; Geiger, 1993; Tuller, 1993;

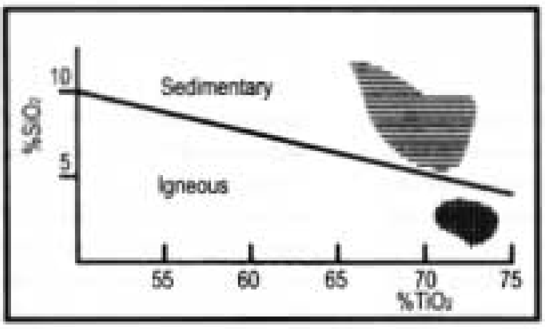


FIGURE 6 - TiO_2 vs SiO_2 diagram showing predominance of sedimentary protoliths in the southeastern region of the Paraíba do Sul Complex (after Sluitner, 1990).

Signorelli, 1993; Silva *et al.*, 1987; Vieira, 1993, 1997; Wiedemann *et al.*, 1997):

- The amphibolite-facies domain largely prevails in the Paraíba do Sul Complex. It consists of a mainly metasedimentary inhomogeneous package of partially migmatized banded gneiss, metamorphosed in the amphibolite to amphibolite-granulite facies transition. It includes biotite gneiss, biotite-garnet gneiss, biotite-garnet-cordierite-sillimanite gneiss, biotite-garnet-cordierite-sillimanite-graphite gneiss (kinzigite *s. s.*), dolomitic and calcitic marbles, sillimanite quartzite, with minor intercalations of quartz-feldspar granulite (leptinite, probably representing felsic volcanics), paraderived calc-silicate granulite and orthoamphibolite.

- The granulite-facies domain is best exposed along the coastal zone of southern Espírito Santo and northern Rio de Janeiro states, and in the Caparaó Ridge (Fig. 5d), but also occurs as indiscriminated inliers in the amphibolite-facies domain. It chiefly consists of charno-enderbitic granulite locally with garnet, enderbitic granulite interlayered with porphyroblastic feldspar gneiss and calc-silicate granulite, amphibole-gneiss, leptinite and garnet-cordierite-sillimanite gneiss.

Two marine environments may be attributed to the Paraíba do Sul Complex. The major contribution of terrigenous siliciclastic material with common sandy terms (greywacke gneiss, quartzite) and thick carbonate layers is assigned to a proximal marine environment, probably a shallow shelf. The presence of graphite and the H_2S-content in the marble reveal reducing conditions during sedimentation and a shallow shelf marine environment (Geiger, 1993). The distal pelite-rich marine environment corresponds to peraluminous gneiss with thin calc-silicate lenses (Sluitner, 1990; Vieira, 1993). Scarce leptinite (probably acid volcanics) and orthoamphibolite interlayered in the kinzigitic gneiss indicate that alumina-rich shale accumulated in deeper marine environments with restrict volcanic contribution. These deposits were metamorphosed to sillimanite-cordierite-garnet-gneiss containing zircon and monazite (Sluitner and Weber-Diefenbach, 1989; Féboli, 1993).

The geochemical interpretation of microcline-, biotite- and biotite-garnet gneiss with calc-silicatic and quartzitic layers from the Paraíba do Sul Complex, in Rio de Janeiro, points towards greywacke and carbonate-cherty chemical sediments, usually with intercalations of metavolcanics. Lenses of dolomitic marble, either siliceous or not, may also interchange with amphibole gneiss and amphibole-pyroxene gneiss. In Espírito Santo, Geiger (1993) studied the continuation of the Paraíba do Sul Complex and confirmed a similar origin. There, quartzite and calc-silicate gneiss are also associated with biotite-K-feldspar- and biotite-amphibole gneiss. Comparable interfingering of carbonate layers in the biotite-amphibole gneiss and migmatite were described. Extensive deposits of dolomitic and calcitic marbles are typical. The presence of igneous intrusions, consisting of metagabbro, metapyroxenite, amphibolite, metadiorite and biotite-andesine gneiss is another similarity with the situation described for Rio de Janeiro. Geochemical data from biotite- and biotite-amphibole gneiss indicate that pelitic sediments and sub-greywacke are the most probable protoliths.

In the southern-central region of Espírito Santo, coeval magmatism is very frequent, consisting of metamorphosed continental basalt (tholeiitic amphibolite) which appears as interlayerings in the paragneiss (Geiger, 1993; Sluitner and Weber-Diefenbach, 1989; Wiedemann *et al.*, 1997). The presence of continental tholeiite intercalated with these metapelite beds is an indication of a thick sialic crust.

Detailed petrological and geochronological surveys on the Paraíba do Sul Complex, in southern Espírito Santo, revealed that the whole lithological association originated from a consistent heterogeneous source (Söllner *et al.*, 1989; Sluitner, 1990; Wiedemann *et al.*, 1997). A greywacke sequence interlayered with clay-rich sediments containing minor amounts of carbonate, as well as volcanogenic deposits (orthoamphibolite and leptinite), formed the original package (Fig. 6). With the increase of regional metamorphism under high H_2O-pressure conditions, part of the sedimentary sequence was melted and crystallized still under high-amphibolite metamorphic conditions. Söllner *et al.* (1991) dated this high-grade metamorphic event in the Espírito Santo region around 590 Ma (U/Pb, zircon). When the main regional deformation ceased, a later change in the fluid content (CO_2-rich fronts) originated charno-enderbitic rocks around 560 Ma (U/Pb, in zircon and monazite).

The acidic to intermediate granulite, graphitic bands, quartzite, feldspar-rich quartzite and kinzigite of the Caparaó Ridge (Fig. 5d) were inferred as having originated from a greywacke sequence associated with a bimodal magmatic suite (Seidensticker and Wiedemann, 1992). Paleoproterozoic (Transamazonian) protoliths have been detected by isotope analysis on inherited zircon rounded cores. The Neoproterozoic granulitic metamorphism was dated at about 586 ± 2 Ma (U/Pb, zircon; Söllner *et al.* 1991). Conversely, Campos Neto and Figueiredo (1995) considered the Caparaó granulite as synorogenic subduction-related calc-alkaline granitoid rocks.

Taking into account the above data, we regard the Paraíba do Sul Complex as representing extensive volcano-sedimentary marine sequences of Neoproterozoic age. This pile of sediments was partially melted and supplied crustal magmas, which contributed to the formation of the AWCO's magmatic arc (see section on Orogenic Magmatism).

Was the Paraíba do Sul Complex a passive margin sequence correlated to the Salinas Formation and the Jequitinhonha Complex, or was it related to an active margin setting?

Grossi Sad and Dutra (1988), and Féboli (1993) were the first to consider the latter hypothesis. Inside the mainly metasedimentary Paraíba do Sul Complex, tonalite to granodiorite gneiss rocks were inferred as corresponding to calc-alkaline island-arc remnants (Grossi Sad and Dutra, 1988). Although those rocks were correlated to the Paraíba do Sul Complex, their radiometric ages are unknown, and they could represent slices of an older crust tectonically emplaced in the Paraíba do Sul Complex, a very common situation in the region under study.

In view of the available data, we regard the Paraíba do Sul Complex as a Neoproterozoic passive margin sequence because: i) along the Paraíba do Sul outcrops there is only evidence of a pre-Mesoproterozoic basement (Söllner *et al.*, 1991; Valadares *et al.*, 1997; Machado *et al.*, 1996; Machado, 1998); ii) this older crust includes calc-alkaline remnants of a Paleoproterozoic magmatic arc (Figueiredo and Teixeira, 1996; Alkmim and Marshak, 1998; Costa, 1998; Fischel, 1998); iii) as highlighted by Söllner *et al.* (in prep.) the lack of 1.85 Ga - 800 Ma age-components in the Paraíba do Sul Complex suggests a long lasting period of quietness, compatible with a passive margin setting; iv) despite the high-grade metamorphic overprint, deformation and degree of partial melting of the Paraíba do Sul rocks, the predominance of pelite-rich gneiss points toward continental source areas; v) the most consistent candidates as source areas for the Paraíba do Sul metasediments are the Paleoproterozoic and Archean complexes (and their equivalents in the West-Congo Belt basement). These older terranes were eroded to form thick successions of supracrustal rocks, which have been metamorphosed to the amphibolite and granulite facies, but preserved high-grade rafts of an older magmatic arc, deformed together with the whole sedimentary pile; vi) indeed, in such a deep crustal level the preservation of a lithological record of volcano-sedimentary deposits from a magmatic arc is quite uncommon. Nevertheless, the roots of this Neoproterozoic arc remain preserved in calc-alkaline plutons (see section on Orogenic Magmatism).

Therefore, the Paraíba do Sul and Jequitinhonha complexes are interpreted as stratigraphic equivalents that, together with the Salinas Formation and correlatives, record the Neoproterozoic passive margin setting of the Araçuaí-West-Congo basin system.

The orogenic event

This section summarizes data and interpretations on deformation, metamorphism and magmatism, which are related to the compressional tectonic regime during the Brasiliano Orogeny. In order to restrict the span of each orogenic stage to its peak moment, only U/Pb (dilution and SHRIMP methods) and Pb/Pb (monozircon evaporation method) data have been taken into account. In spite of the bulky set of Rb/Sr and K/Ar dates, ages yielded by both these methods generally led to biased controversial interpretations. In the region, the reliable zircon U/Pb and Pb/Pb ages for pre, syn and late collisional granitoid plutons indicate that the orogenic event lasted from *c.* 625 to 560 Ma (Pedrosa-Soares *et al.*, 1999b). This event was preceded by the emplacement of alkaline intrusions at about 676 Ma in the cratonic region of southeastern Bahia (Teixeira *et al.*, 1997).

Regional deformation and metamorphism

The collisional stage of the Araçuaí-West-Congo Orogen increased metamorphism, overprinted older structures and caused shortening of *c.* 30 - 40% in cross-sections. The main regional schistosity or gneissic foliation defines the structural trend of each tectonic domain, as previously described. This schistosity or gneissic foliation is associated with frontal to oblique thrust systems or with steeply dipping oblique to strike-slip shear zones.

The following descriptions emphasize the recorded Brasiliano structures and metamorphism. Only mineral assemblages, which are synkinematic to the main regional foliation have been considered in the descriptions of metamorphic zoning. We selected four cross-sections as representative for the orogenic frame in different portions of the Araçuaí Belt (Fig. 5). The selected cross-sections are generalized, regional sections in which details, such as local metamorphic truncations and tectonic intercalations of units, are absent. It is worth mentioning that locally thrusting often inverts metamorphic polarity. For the northern curvature of the Araçuaí Belt, the external and internal tectonic domains will be described separately.

External tectonic domain

Thick fault blocks make up the frontal thrust system that characterizes the northern segment of the external tectonic domain. In this part of the belt (Fig. 5a), Neoproterozoic rocks predominate (Pedrosa-Soares *et al.*, 1992; Crocco-Rodrigues *et al.*, 1993; Alkmim *et al.*, 1996; Pedrosa-Soares and Pedreira 1996; Uhlein *et al.*, 1998a). The main regional schistosity trends N-S to NNE, and tends to turn steeper from W to E. This S-surface contains an E-plunging, downdip stretching lineation. A W-dipping crenulation cleavage cuts the regional schistosity, and may be prominent as it surrounds the eastern branch of the northern Espinhaço Supergroup.

In the Macaúbas Group, the Brasiliano metamorphism increases from W to E, grading from low greenschist to low amphibolite facies, along the northern segment of the external tectonic domain. The W to E succession of metamorphic zones (chlorite, biotite, biotite + garnet, garnet ± staurolite ± kyanite) is typical of an intermediate-pressure (Barrowian-type) metamorphic regime. In the southern segment of the external tectonic domain, the Macaúbas Group is narrow and displays metamorphism of the greenschist facies, chlorite zone (Pedrosa-Soares *et al.*, 1992; Pedrosa-Soares and Pedreira 1996; Grossi-Sad *et al.*, 1997).

Northern subdomain of the Internal tectonic domain

This NE- to NNE-trending subdomain ends southwards around the 19°S parallel (figs. 5b,c). Shallow to moderately-dipping thrusts, oblique NE-plunging stretching lineation imprinted in the regional foliation, and positive flower-structures characterize this subdomain as a SW-vergent oblique thrust system (Pedrosa-Soares *et al.*, 1992, 1993; Cunningham *et al.*, 1996; Pinto *et al.*, 1997; Oliveira *et al.*, 1997; Uhlein *et al.*, 1998a,b).

Two metamorphic regimes were identified in this subdomain. The metapelite of the Salinas and Capelinha formations (Fig. 4) displays a W to E succession of intermediate-pressure metamorphic zones of the amphibolite facies. These zones are marked by the progressive crystallization of staurolite, kyanite and sillimanite, in garnet-biotite schist (Pedrosa-Soares *et al.*, 1992, 1993, 1996; Cunningham *et al.*, 1996; Grossi-Sad *et al.*, 1997; Pinto *et al.*, 1997). These metamorphic assemblages are especially abundant in metapelite of the Salinas metavolcano-sedimentary unit, in which they are synkinematic in relation to thrusts that contain slivers of Neoproterozoic oceanic lithosphere (Fig. 5c).

In the Salinas Formation, the low-pressure metamorphic regime has been only identified eastwards from the town of Araçuaí. There, from W to E, the metamorphic zones in biotite schist are marked by a progressive crystallization of andaluzite, cordierite and sillimanite (Costa, 1989). This sector coincides with the axis and eastern flank of a large, positive flower-structure (Fig. 5b), where the metamorphic PT path reaches the lowest pressure value (*c.* 2 kbar) in relation to temperature, *c.* 600 °C (Pedrosa-Soares *et al.*, 1996). To the E, these schists give rise to foliated garnet-biotite granite of the G2 suite that preserved rafts of sillimanite-cordierite-garnet gneiss.

Although this low-pressure metamorphic zoning is confined to a small sector of the Salinas Formation, it distinguishes the boundary between the Brasiliano intermediate-pressure regime, typical for the whole region W of 42 °W, and the low-pressure/high-temperature regime that prevails in the migmatitic-granulitic core of the orogen (Leonardos and Fyfe, 1974).

The progressive metamorphic assemblages observed in the Jequitinhonha Complex paragneiss consist of biotite + K-feldspar + plagioclase + garnet ± cordierite ± sillimanite ± graphite. Extensive migmatization is a typical characteristic. Calc-silicate granulite lenses show quartz + calcium-rich plagioclase + diopside ± hypersthene. Based on textural relations and mineral reactions, Faria (1997) and Reis (1999) concluded that the Jequitinhonha Complex was metamorphosed in the amphibolite-granulite facies transition, high anatexis zone, in PT conditions ranging from 3 to 6 kbar, and 700° to 800° C. The increase in migmatization turns this segment into the anatetic core of the orogen.

The NW-trending, sinistral, steeply dipping, Vitória Shear Zone is a prominent feature both in the northern and southern subdomains of the internal tectonic domain (Fig. 4). Most of the late and postcollisional plutons with charnockite facies crop out along this lineament, suggesting that this long-lived shear zone was the path for extensive CO_2-rich fluids, which might have affected the crystallization conditions of I-type calc-alkaline intrusions.

Southern subdomain of the Internal tectonic domain

The northern limit of this NNE-trending subdomain runs approximately along the 19°S parallel (Fig. 4). The Brasiliano ductile deformation is imprinted by W-vergent, moderately to steeply-dipping thrusts, which are embedded and/or truncated by important dextral, high-angle dip, oblique- to strike-slip shear zones, outlining a prominent transpressive system as shown in Figure 5d (Lammerer, 1987; Fritzer, 1991; Costa *et al.*, 1993, 1998; Wiedemann, 1993; Cunningham *et al.*, 1996, 1998; Oliveira *et al.*, 1997; Pinto *et al.*, 1997; Wiedemann *et al.*, 1997; Ebert and Hasui, 1998; Fischel, 1998).

The metamorphic banding was tightly to isoclinally folded and refolded to form long-wave folds, with upright to slightly W-verging axial planes and amplitudes up to 10 km (Lammerer, 1987; Fritzer, 1991). Such folding is associated to contemporaneous stretching parallel to the fold axes, indicative of a transpressive regime. Progressive motion and metamorphic recrystallization along the oblique- to strike-slip shear zones were superimposed on the previous fold system. These high-angle dip shear zones were active throughout the orogen, even during the docking stage of the colliding blocks, when accommodation of motion was in progress. The most outstanding shear zone of the region is the Guaçuí Lineament (Fritzer, 1991).

The Bouguer profile shows the limit between the external and internal tectonic domains (Fig. 5d). A large negative anomaly marks the external tectonic domain, separating the gravimetric patterns from the cratonic region and internal tectonic domain. In the western border of the internal tectonic domain, the Dom Silvério Group shows progressive crystallization of staurolite, kyanite and sillimanite, in garnet-biotite schist (Jordt-Evangelista, 1992). In like manner to the Macaúbas Group this group records the intermediate-pressure metamorphic regime that characterizes the western segment of the internal tectonic domain. Gravimetric values are stronger in the Juiz de Fora Complex, where they reach a maximum around - 40 mgal, implying a significant influence of high-density rocks (Fig. 5d). Along the Manhuaçu Shear Zone, the transpressive tectonics has juxtaposed slices of basic granulite of the Juiz de Fora Complex with Neoproterozoic peraluminous granulite and staurolite-garnet gneiss (Costa *et al.*, 1998). Geothermo-barometric studies of granulite and high-amphibolite gneiss yielded metamorphic temperatures greater than 800 °C and pressures around 8 - 10 kbar, confirming that this region exposes the deepest crustal levels of the orogen (Fritzer, 1991; Seidensticker and Wiedemann 1992; Costa *et al.*, 1993; Costa, 1998).

To the E of the Manhuaçu Sear Zone, the up and down gravimetric pattern decreases to a minimum of about - 70 mgal close to the Guaçuí shear zone, recording the thick metasedimentary packages of the Rio Doce Group and Paraíba do Sul Complex. In this region, the P-T metamorphic conditions reached temperatures greater than 650 °C and pressures around 6.5 - 7.5 kbar, at about

590 Ma (Geiger, 1993). From the Guaçuí Shear Zone towards the town of Vitória, the gravimetric values increase again (Fig. 5d). Although most of this increasing is due to the present Atlantic oceanic crust, some influence may be attributed to the Paraíba do Sul granulitic rocks, exposed along the coast, and to the widespread intrusion of granitoid plutons with mafic cores. These granulitic rocks crystallized under temperatures greater than 800 °C, and pressures around 6.5 - 7.5 kbar.

Northern curvature

The northern curvature of the Araçuaí Belt is given by a bent structural trend with predominant steeply dipping schistosity or gneissic foliation (figs. 3 and 5b). The oblique to strike stretching lineation imprinted in the regional foliation usually plunges to NE, E, and SE, respectively in the western, central and eastern segments of the curvature. Axes of tight to isoclinal folds are usually roughly parallel either to the stretching lineation or to the L-type tectonites. From N to S, deformation grades from brittle, in the cratonic border, to ductile along the high-angle dip shear zones. The northern curvature is understood as an oblique to transcurrent system that seems to have accommodated, along a prior basin margin, the Brasiliano orogenic closure (Pedrosa-Soares *et al.*, 1992; D'El-Rey-Silva *et al.*, 1995; Faria, 1997; Reis, 1999; Uhlein *et al.*, 1999).

The Brasiliano metamorphic zoning is marked, from N to S, by the progressive appearance of biotite, garnet, staurolite, kyanite and sillimanite in pelitic schist and gneiss of the Macaúbas Group. An intermediate-pressure regime was outlined by Almeida *et al.* (1978).

The same progressive metamorphic assemblages, observed in the Jequitinhonha Complex from the northern internal tectonic domain, continue in this sector.

Evidence for a Suture Zone

The absence of Neoproterozoic oceanic slivers in the African side of the Araçuaí-West-Congo Orogen has been asserted (Trompette, 1994, 1997; Vicat and Pouclet, 1995; Tack and Fernandez-Alonso, 1998). In the Araçuaí Belt Neoproterozoic oceanic remnants were several times described in the last ten years (Pedrosa-Soares *et al.*, 1992, 1998, 1999b; Grossi-Sad *et al.*, 1997; Pinto *et al.*, 1997; Cunningham *et al.*, 1998; Uhlein *et al.*, 1998a). The main field evidence is a volcano-sedimentary rock-assemblage association of deep-sea pelite beds and mafic rocks, consisting of metamorphosed banded iron formation units (oxide, silicate and sulfide-type), metachert (with variable sulfide content, Al-silicates, graphite and/or Mn-rich minerals), diopsidite (a singular Mg- and sulfide-rich rock, probably derived from a volcanic-exhalative sediment), bodies of massive sulfide, hyperaluminous schist (extremely rich in staurolite and/or kyanite and/or sillimanite, with some graphite), Mn-rich rocks (gondite), graphite schist, and orthoamphibolite concordant intercalations, associated with mica schist and quartz-mica schist (with variable contents of accessory garnet, staurolite, kyanite and/or sillimanite), and slices of meta-ultramafic rocks. In general, these rock-assemblages are only preserved in disrupted sections (figs. 4, 5c - d).

The amphibolite from the Ribeirão da Folha area (close to the northernmost ultramafic body shown in Figure 4) represents orthoderived mafic rocks with an ocean-floor geochemical signature, comparable to both modern ocean-floor basalt and Neoproterozoic ophiolite (Fig. 7). Their magmatic protoliths were extracted from a depleted mantle and crystallized at about 816 ± 72 Ma (Sm/Nd whole-rock isochron, five samples, $\varepsilon_{Nd(T)} = +3.4$ to $+4.6$; Pedrosa-Soares *et al.*, 1998). The meta-ultramafic rocks constitute up to 1 km thick syntectonic slabs, that were thrusted into the volcano-sedimentary sequence (figs. 4 and 5c). The meta-ultramafic rocks are generally tremolite- and/or talc-rich schists but peridotite and pyroxenite relicts are also present (Pedrosa-Soares *et al.*, 1992, 1998). The schistosity and stretching lineation of both the ultramafic schists and the country rocks exhibit the same orientation.

The present position of these remnants of oceanic lithosphere is approximately along the 42°W meridian, between the parallels 17°S and 19°S, making up a segment of a suture zone marked by major thrusts and positive flower-structures (figs. 4 and 5c). If the Dom Silvério Group actually includes Neoproterozoic oceanic remnants (Cunningham *et al.*, 1998), this suture could be extended to around 20°40'S. This line matches well with regional magnetic and gravity linear anomalies (Haralyi and Hasui, 1982; CPRM-SGA, 1997).

Orogenic Magmatism

Contrasting with the West Congo Belt, where no orogenic calc-alkaline granitoid plutons of Neoproterozoic age are known (Trompette 1994, 1997; Tack and Fernandez-Alonso, 1998), a profusion of late Neoproterozoic to Cambro-Ordovician calc-alkaline, I-type granitoid plutons occur along the internal tectonic domains of the Araçuaí Belt and northern Ribeira Belt (Wiedemann, 1993; Campos-Neto and Figueiredo, 1995; Nalini, 1997; Pinto *et al.*, 1997; Wiedemann *et al.*, 1997; Tupinambá *et al.*, 1998; Uhlein *et al.*, 1998b; Celino, 1999; Noce *et al.*, 1999; Paes, 1999; Pedrosa-Soares *et al.*, 1999a,b; Silva, 1999).

Taking into account high-quality zircon U/Pb and Pb/Pb geochronological data, the subdivision of the magmatism into five suites is given as proposed by Pedrosa-Soares *et al.* (1999b). The present G1, G2, G4 and G5 suites correspond to the G2, G1, G5 and G4 suites of Pedrosa-Soares *et al.* (1999a), respectively.

The S-type suites (peraluminous) are those melted from a largely predominant (meta)sedimentary source with minor contribution from the partial melting of an oceanic crust and/or a mantelic source. The I-types (metaluminous) are melts derived from mixed sources, with an important contribution of an oceanic crust and/or a mantelic source mixed or mingled with partial melts from a prevailing (meta)igneous and minor (meta)sedimentary crust.

The G1 and G2 suites are related to the pre to syncollisional stage. Two distinct contemporaneous G3 suites, the S-type G3-S and I-type G3-I (Pedrosa-Soares *et al.*, 1999b) are associated with a late to postcollisional stage. The G4 and G5 suites are related to the final collapse of the orogen.

Pre to syncollisional magmatism

The G1 suite is characterized by extensive multi-diapiric granitoid bodies. They consist of tectonically foliated tonalite and granodiorite, and minor granite (Figs. 4 and 8). This suite is also known by local names, such as Brasilândia, Estrela-Muniz Freire, Galiléia, Guarataia, and São Vitor.

The G1 granitoid plutons usually show deformed eye-shaped K-feldspar megacrystals in a foliated biotite-rich matrix. Metamorphic and mylonitic textures predominate, but magmatic textures tend to be preserved within pluton cores. Local signs of migmatization have been observed. Mafic to intermediate microgranular enclaves, usually stretched along the gneissic foliation, are widespread. A late retrograde metamorphism is portrayed by the growth of biotite and the development of symplectic textures in granulitic/charnockitic rocks (Lammerer, 1987; Fritzer, 1991; Geiger, 1993; Nalini, 1997; Pinto *et al.*, 1997; Wiedemann *et al.*, 1997; Bilal *et al.*, 1998; Aracema *et al.*, 1999; Paes, 1999; Pedrosa-Soares *et al.*, 1999b).

Geochemical data from several G1 plutons point towards calc-alkaline, metaluminous to slightly peraluminous magmas, formed in a volcanic arc setting (Fig. 8). These batholitic intrusions may therefore represent the roots of a pre to syncollisional magmatic arc, related to subduction of oceanic lithosphere (Geiger, 1993; Wiedemann, 1993; Campos-Neto and Figueiredo, 1995; Wiedemann *et al.*, 1997; Aracema *et al.*, 1999; Pedrosa-Soares *et al.*, 1998, 1999a,b).

In the Galiléia Batholith (SE of Governador Valadares, Figure 4), microgranular enclaves with grossular-rich garnet ($Al_{40-64}Gr_{20-42}Sp_{7-15}Py_{3-9}$) and anorthite-rich plagioclase (An_{70-80}) may indicate contribution of a mantle source, and magmatic crystallization under pressures greater than 10 kbar (Nalini *et al.*, 1998; Bilal *et al.*, 1998). Nonetheless, Rb/Sr and Sm/Nd isotope data suggest significant involvement of continental material, resulting in granitoid crystallized from hybrid magmas formed by crustal and mantelic components (Nalini, 1997).

Söllner *et al.* (1991), dating the Estrela-Muniz Freire Batholith, in the State of Espírito Santo, obtained the age of 580 +20/-6 Ma (U/Pb in zircon). According to Nalini (1997), the magmatic crystallization of the Galiléia Batholith took place at about 594 ± 6 Ma (U/Pb in zircon). Orthogneiss of the Brasilândia, São Vitor and Guarataia plutons, located in the region of Teófilo Otoni (Fig. 4), yielded the ages of 595 ± 3 Ma, 576 ± 5 Ma and 575 ± 2 Ma (Pb/Pb in zircon), respectively (Noce *et al.*, 1999). The zircon from a tonalitic orthogneiss yielded a Pb/Pb age of 625 ± 11 Ma (Paes *et al.*, 1999), indicating the magmatic crystallization age of a G1 pluton located near Alvarenga (southeast of Governador Valadares). The referred geochronological data constrain the evolution of the G1 suite from *c.* 625 to 575 Ma. This time interval also limits the age of regional metamorphism (Pedrosa-Soares *et al.*, 1999b).

Despite their different ages, the São Vitor Batholith and Brasilândia Pluton (Teófilo Otoni region, Figure 4) show a consistent geochemical signature of precollisional, volcanic arc granitoid rocks (Aracema *et al.*, 1999; Noce *et al.*, in prep.). The same signature is displayed by the Galiléia Batholith (Nalini, 1997). In addition, the ages obtained for the Galiléia Batholith (594 ± 6 Ma) and Brasilândia Pluton (595 ± 3 Ma) are equal. This suggests a time constraint for the precollisional stage of the orogen, *i. e.*, the minimal age of this stage could be around 595 Ma. The volcanic arc signature of the younger São Vitor and Estrela-Muniz Freire batholiths may indicate that the subducted oceanic crust affected the I-type granitogenesis, even during the collisional stage, until *c.* 575 Ma.

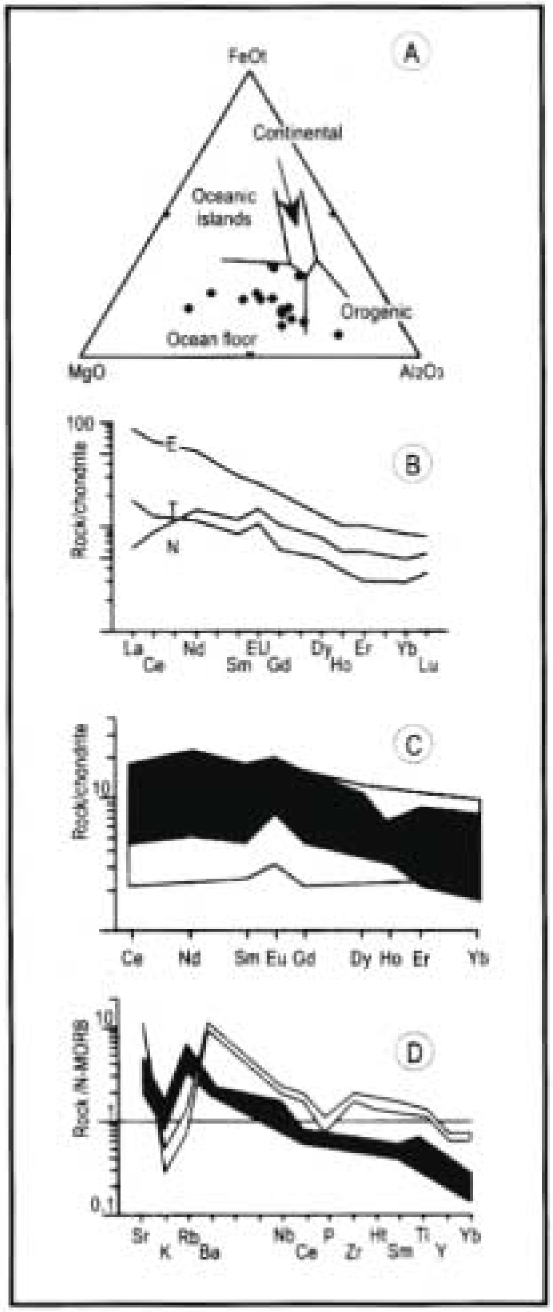
In Espírito Santo, the intrusion of granodioritic, tonalitic and granitic magma, under wet amphibolite facies conditions, was slightly later (580 Ma) than the main metamorphic phase (590 Ma; Wiedemann *et al.*, 1997). Metamorphic temperatures and pressures, deduced from mineral paragenesis, range from 640° to 685°C and 5.6 to 7.8 kbar (Geiger, 1993), clearly crossing the granite and tonalite *solidus* curves.

Part of the syncollisional G1 melts crystallized under drier conditions and high CO_2 pressures, originating magmatic charnockitoid with noritic cores (Fritzer, 1991; Wiedemann *et al.*, 1997; Mendes *et al.*, 1997). The hypersthene-bearing orthogneiss from the Serra do Valentim (W of the Guaçuí Shear Zone) was formed under temperatures of at least 800°C and pressures of 8 to 9 kbar. The very low H_2O pressure values calculated for this unit were presumably controlled by a combination of dehydration and melting reactions (Fritzer, 1991).

Small bodies of granular mantle-derived phlogopite peridotite have been mapped and connected with antiformal structures (Tuller, 1993; Wiedemann *et al.*, 1997). The presence of metamorphosed basic magmatites together with Rb/Sr and Sm/Nd isotope data are strong evidence that during the main orogenic stage, the supply of crustal magma was contemporaneous with that of mantle-derived melts.

The G2 suite (Fig. 4) is also called by local names, such as Buranhém, Iriri, Montanha, Nanuque, São Paulinho and Urucum. Although some large intrusions, like the Urucum plutons, are clearly intrusive and discordant, gradational contacts with the Jequitinhonha and Paraíba do Sul complexes prevail. Ghost migmatitic structures and ubiquitous rafts of migmatite and paragneiss indicate a predominant autochthonous to para-autochthonous nature. A widespread process of *in situ* migmatization originated large volumes of diatexite (Costa, 1989; Uhlein, 1991; Trompette, 1994; Pedrosa-Soares and Pedreira, 1996; Faria, 1997; Pinto *et al.*, 1997; Wiedemann *et al.*, 1997; Celino, 1999; Pedrosa-Soares *et al.*, 1999a,b). The G2 suite makes up the anatetic core of the Araçuaí-West-Congo Orogen.

This suite mainly consists of batholitic bodies of tectonically foliated, S-type, subalkalic calc-alkaline, peraluminous garnet-biotite granite. Cordierite and/or sillimanite are accessory minerals, but they may be absent in garnet-poor two mica granite intrusions. The gneissic foliation follows the structural trends of the Araçuaí Belt, indicating metamorphic recrystallization during the main deformational phase (Faria, 1997; Nalini, 1997; Pinto *et al.*, 1997; Celino, 1999; Pedrosa-Soares *et al.*, 1999a,b). The Urucum intrusive leucogranite crystallized under P-T conditions around 4 kbar and 600° - 750°C, within a fluid-rich environment (Nalini *et al.*, 1998). U/Pb (Nalini, 1997) and Pb/Pb (Noce *et al.*, 1999) zircon ages allow us to



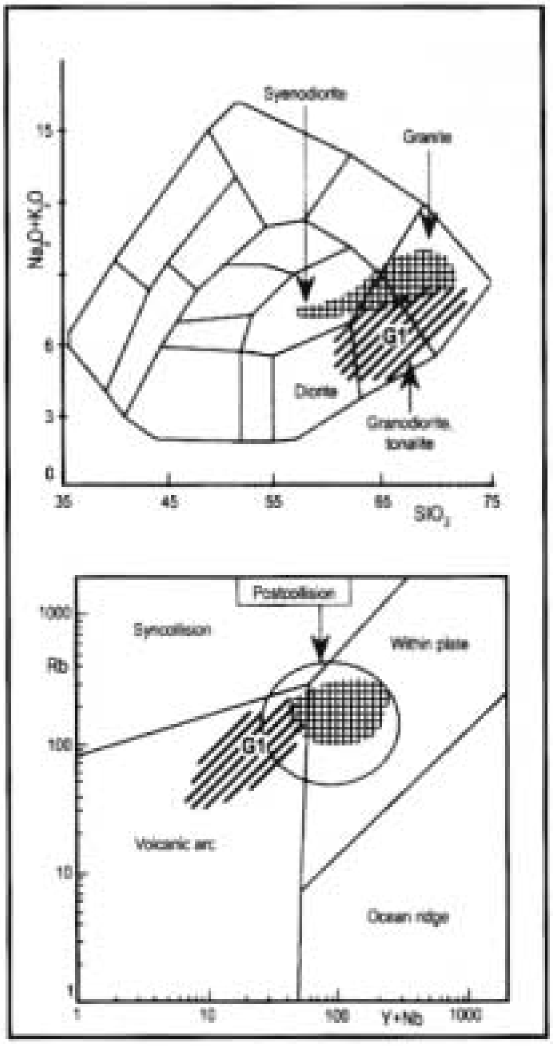


FIGURE 7 - Geochemical signature of ortho-amphibolite of the Ribeirão da Folha facies (distal unit of the Salinas Formation), in relation (A) to basic volcanic rocks from modern geotectonic settings (diagram fields from Pearce et al., *1977), correlated (B) to normal (N), transitional (T) and enriched (E) mid-ocean-ridge basalt (MORB) types of Saunders (1984), and compared (dark shaded fields) to metamafic rocks of early Neoproterozoic ophiolite from (C) Arabian-Nubian Shield (Bokhari and Kramers, 1981; Claesson* et al., *1984) and (D) Taymyr Belt, northern Siberia (Vernikovskiy* et al., *1998) (after Pedrosa-Soares* et al., *1998, 1999b).*

FIGURE 8 - Geochemical signature of the G1 suite, contrasting to G3-I and G5 suites. Data of the G1 suite are from Nalini (1997) for the Galiléia Batholith; and from Aracema et al. *(1999) for the São Vitor Batholith and Brasilândia Pluton. Data of the G3-I and G-5 suites are from Fernandes (1991), Wiedemann (1993), Faria (1997), Celino (1999), and Achtschin (in prep.). Diagram fields are from Wilson (1989) and Pearce (1996) for Y+Nb x Rb.*

constrain the evolution of G2 suite from *c.* 591 Ma to 575 Ma, in the region from 19° 30'S to 16 °S (Fig. 4).

In the southern coastal region (Costeiro Complex), a whole sequence of sediment-derived gneiss, which underwent partial melting, was deformed and recrystallized around 575 ± 10 Ma (U/Pb in zircon and monazite). Folded garnet-bearing leucosomes indicate that the high-grade metamorphism was followed by dehydration melting, but still during the main deformational phase. Around 560 Ma (U/Pb in zircon and monazite) these previously deformed gneiss (mostly metadiatexite) underwent local charno-enderbitization to originate a package of green hypersthene-bearing augen gneiss, granitoid gneiss and enderbitic gneiss, which was called the Iriri Series (Sluitner and Weber-Diefenbach, 1989; Söllner *et al.*, 1989). Inhomogeneous metamorphic banding, calc-silicate interlayering and geochemical signatures of the Iriri Series indicate a primary supracrustal origin. Although showing contrasting geochemical characteristics, this granitoid-generating episode is coeval with the I-type 580 Ma old magmatism (G1 suite). The Iriri Series represents a later recrystallization of S-type metagranitoid bodies, still associated with the same orogenic event. To explain this younger age, Campos Neto and Figueiredo (1995) proposed a second orogenic event, the Rio Doce Orogeny, with a 560 Ma collisional episode. However, the 560 Ma metamorphic episode most probably resulted from the progressive late collisional reheating at low-P (3 - 4 kbar) and high-T (*c.* 800°C) conditions (Sluitner and Weber-Diefenbach, 1989; Wiedemann *et al.*, 1997).

Late to postcollisional magmatism

Two coeval magmatic suites were generated in the Araçuaí Belt internal domain, during the late orogenic stage. The G3-I suite consists of calc-alkaline granitoid intruded along oblique to strike-slip shear zones. The subalkaline G3-S suite corresponds to the remelting of the peraluminous G2 suite (Fig. 4).

The G3-I suite is also called by local names, such as Itagimirim, Guaratinga, Lagoa Preta, Rubim, Salomão and Santo Antônio do Jacinto. G3-I plutons are roughly elliptical bodies, generally emplaced along steeply dipping oblique- to strike-slip shear zones. Aureoles of recrystallization and incipient partial melting of host rocks surround the intrusions. Remarkable magmatic flow structures are roughly parallel to the gneissic foliation of the country rocks. Locally, tectonic foliation is restricted to the regions along pluton borders (Faria, 1997; Uhlein *et al.*, 1998b; Celino, 1999).

G3-I granitoid plutons mainly consist of granite to granodiorite, with frequent microgranular enclaves of intermediate to mafic rocks. Some G3-I bodies are complexly zoned, comprising gabbroic to granitic rocks and minor charno-enderbitic facies (Vieira, 1997). They are I-type, high-K calc-alkaline granitoids. Geochemical signatures (Fig. 8) together with structural features indicate they were emplaced and crystallized during the late to postcollisional phase (Faria, 1997; Celino, 1999). The magmatic crystallization is constrained by a Pb/Pb zircon age (Dussin *et al.*, 1998) and Rb/Sr ages (Siga, 1986; Faria, 1997) to 585 - 570 Ma.

The G3-S suite consists of a series of small coalescent and isolated leucogranitic sillimanite-cordierite-garnet granitic bodies. Almenara is the local name for the G3-S suite. This peraluminous subalkaline granite is only present where the G2 suite predominates (Pedrosa-Soares *et al.*, 1999a). Widespread migmatitic ghost structures and gradational contacts with the deformed host migmatites reveal the anatetic, autochthonous to para-autochthonous nature of the G3-S suite. These granitoid plutons were formed under wet crustal melting conditions, which lasted for the period that succeeded the peak of metamorphism and deformation. No reliable age is available for these granites. They were not found in the southern region of the belt.

The late orogenic stage: final collapse of the orogen

A quiescence in the magmatic activity seems to have taken place from 560 Ma to 535 Ma along the entire belt. A new magmatic episode started only around 535 and lasted until 490 Ma. The G4 and G5 suites were emplaced and crystallized during this late stage (Fig. 4). They comprise, respectively, S and I-type plutons, generally intruded along high-angle shear zones and antiformal fold hinges, which belong to the previous deformations (Figs. 5b, c, d).

The G4 suite consists of S-type, peraluminous to slightly metaluminous, balloon-like granitic plutons (Fig. 4). This suite is locally named as Coronel Murta, Itaporé, Mangabeiras and Santa Rosa. These intrusions are the source of a myriad of lithium and tourmaline-rich pegmatites. Mineral associations of contact metamorphism and petalite mineralization, instead of spodumene, in some pegmatites, indicate depths of emplacement between 12 to 6 km (Correia-Neves *et al.*, 1986; Pedrosa-Soares *et al.*, 1987; Grossi-Sad *et al.*, 1997; Pinto *et al.*, 1997).

Zoned plutons contain biotite-granite centres grading into two-mica or muscovite-garnet leucogranite towards the upper borders. The cupolas of the intrusions consist of residual pegmatoid granite. Xenoliths of the host rocks are commonly found in the outer parts of the intrusions. The emplacement mechanism forced the regional schistosity to accommodate around the intrusive bodies, forming post-tectonic curvilinear structures clearly detected in aerial photographs and satellite images (Pedrosa-Soares *et al.*, 1987; Monteiro *et al.*, 1990).

Zircon from one of the G4 granite intrusions yielded a Pb/Pb crystallization age of 530 ± 8 Ma (Basílio *et al.*, 1998). Inherited rounded zircon of late Archean age and high $^{87}Sr/^{86}Sr$ ratios testify the melting of ancient metasedimentary rocks to generate the G4 suite. This suite is restricted to the western limit of the internal tectonic domain. In the southernmost region the G4 suite is not found, probably due to the exposure of much lower crustal levels.

The G5 suite records the youngest postcollisional magmatic episode (Figs. 4 and 8). It is characterized by several diapirs with compositions varying from gabbro to granite. Although considerably younger, the G5 suite resembles the G3-I suite in most field and petrographic aspects. Local names for G5 suite are Aimorés, Caladão, Mimoso do Sul, Padre Paraíso, Pedra Azul, Santa Angélica, Várzea Alegre, and Venda Nova.

The Pedra Azul, Caladão, and Padre Paraíso batholiths represent the northern granitoid bodies of the G5 suite (Aimorés Suite of Pinto *et al.*, 1997). Granitic batholiths with charnockitic (Padre Paraíso Charnockite and Caladão Granite) or enderbitic (Mangalô Enderbite) cores or borders (Carvalho and Pereira 1997b; Pinto *et al.*, 1997) are common in this region. Coalescent G5 bodies of megaporphyritic granite grading into granodiorite form large polydiapiric structures, with the metamorphic foliation of the host rocks wrapped around them (Figs. 5b, c, d). They occasionally show border foliation that becomes stronger according to the depth of intrusion. Finer-grained biotite granodiorite, biotite monzo to syenogranite are late facies intruded as stocks, dykes and, in some places, form the uppermost portion of the plutons. These intrusions are the source of tourmaline-poor, beryl-rich pegmatite (Achtschin *et al.*, 1998).

In the region of the State of Espírito Santo, the deep erosional level reveals the roots of these diapirs, the inverse zoning of which is usually formed by the interfingering of basic to intermediate magma, in the core, and sienomonzonitic to granitic magma in the borders, showing widespread magma mingling structures (Bayer *et al.*, 1986; Wiedemann, 1993; Wiedemann *et al.*, 1997). Whole-rock geochemical analysis point towards the existence of three magma series: a tholeiitic, a calc-alkaline and an alkaline series (Wiedemann, 1993). The calc-alkaline series is the most outstanding and comprises about 90% of all G5 plutons in the southern region.

The G5 suite comprise metaluminous, high-K calc-alkaline, I-type granitoid plutons originated in the lowermost continental crust with important mantle contribution (Bayer *et al.*, 1986; Fernandes, 1991; Wiedemann, 1993; Faria, 1997; Ludka *et al.*, 1998; Bilal *et al.*, 1998; Mendes *et al.*, 1997, 1999; Achtschin, in prep.).

Rb/Sr and Sm/Nd isotope data (Medeiros, 1999) and the geochemical signatures (Ludka *et al.*, 1998) indicate an enriched mantle reservoir for the basic and intermediary rocks from this suite. The calculated CHUR model age (related to the time when the basic magma was extracted from the mantle) is around 1.0 Ga, and may reflect an episode of mantle enrichment.

Some plutons with charno-enderbitic outer rims testify crystallization under high CO_2 fluid pressure. Bilal *et al.*, (1998) studied the ring-like structure of Aimorés and Ibituba-Itapina (537 - 520 Ma; U/Pb in zircon), in the State of Espírito Santo. They report both hypersthene-rich and hypersthene-free granitoid rocks with ilmenite and magnetite. Calculated T-conditions range from 760° to 820°C, while P-conditions were calculated from 6.5 to 7.0 kbar for the charnockite, 5.9 to 6.2 kbar for the granite, and nearly 8 kbar for the diorite. Comparable P-T values were obtained by Mendes *et al.* (1999) for the structures of Várzea Alegre and Venda Nova. Detailed geochemical studies on both charnockitoids and innermost domains of G5, indicate that the older intrusions are predominantly lower to medium depth crustal melts (> 20 km), whereas the younger intrusions have an important mantle component. From around 530 to 490 Ma a progressive increase in the amount of mantelic melts mixed with crustal melts suggests an important mechanism of underplating (Mendes *et al.*, 1999; Medeiros and Wiedemann, 1999). Zircon ages constrain the magmatic crystallization of G5 plutons from *c.* 530 Ma to 490 Ma (Söllner *et al.*, 1991; Noce *et al.*, 1999), confirming their postcollisional signature (Fig. 8).

An evolutionary model

The evolution of the Araçuaí Belt and of the northernmost portion of the Ribeira Belt, in a broad geotectonic scenario, is summarized here, under the afore-mentioned global tectonics paradigm. Our evolutionary model used as keystones the following evidence: i) remnants of Neoproterozoic oceanic lithosphere; ii) late Neoproterozoic orogenic magmatism, marking precollisional, syncollisional and late collisional stages; and iii) late Cambrian to early Ordovician, late orogenic bimodal magmatism.

Our summarized model is portrayed in the cartoon (Fig. 9). Seven evolutionary steps illustrate the geological history of the region (references on regional data and interpretations were cited in the previous sections of this chapter):

- At about two billion years ago, the Atlantica Paleocontinent was formed by a Paleoproterozoic (Transamazonian-Eburnean) collage (Rogers, 1996). At that time, ancient cratons were welded together along the Mantiqueira-Itabuna Orogen (*cf.* Barbosa and Dominguez, 1996; Alkmim and Marshak, 1998). This Paleoproterozoic paleocontinent became a stable platform around 2.0 - 1.9 Ga (Brito-Neves and Sato, 1998).

- A late Paleoproterozoic taphrogenesis gave rise to the Espinhaço Basin (Brito-Neves *et al.*, 1995). Sediments and bimodal volcanics were deposited in this long-lived continental rift-sag basin from *c.* 1.77 to 1.25 Ga. The Borrachudos Granite represents the plutonic magmatism associated with this rifting process. The Espinhaço Basin did not evolved to an ocean basin, and only underwent orogenic inversion during the Brasiliano Orogeny. Consequently, in late Mesoproterozoic time, the South American sector of the Atlantica Paleocontinent consolidated as a platform as the Rodinia Supercontinent was assembled (Dalziel, 1997).

- Ascent of a mantle plume, beneath this region of the Rodinia Supercontinent, generated mafic dyke swarms and volcanic rocks, and induced anorogenic felsic magmatism and a rifting process. Sediments and volcanics of the proximal unit of the Macaúbas Group and of the West-Congo Supergroup filled the Neoproterozoic rift. This stage is partly coeval with a glaciation that is explained in different ways: i) the region was in high latitudes during that time (D'Agrella *et al.*, 1990), ii) the elastic rebound of the rift shoulders induced a mountain glaciation (Martins-Neto, 1998b), iii) the glaciation was of global amplitude (Hoffman *et al.*, 1998).

- The Neoproterozoic rift went through a complete Wilson Cycle, evolving to an ocean basin, although its northernmost portion (northwards from the zone of oceanic slivers in Figure 3) remained ensialic. The Salinas Formation, Dom Silvério and Rio Doce groups, Paraíba do Sul and Jequitinhonha complexes record passive margin

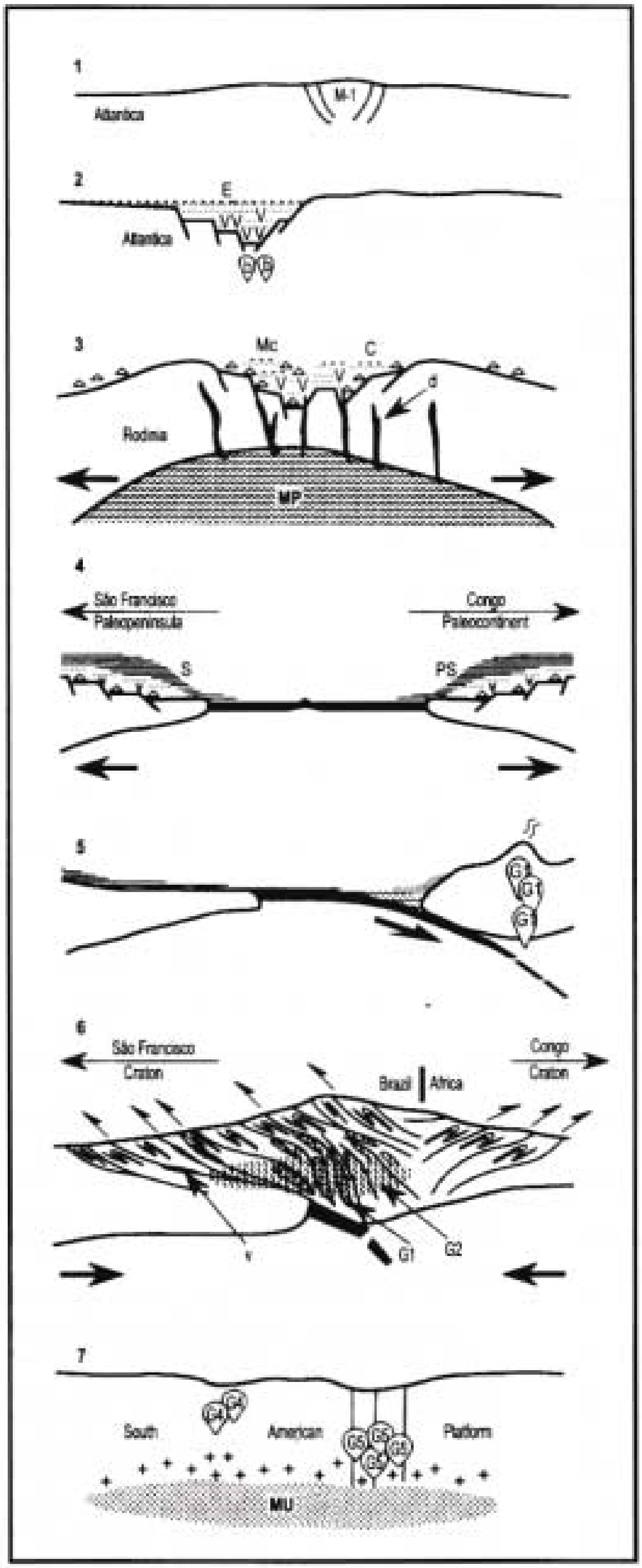


FIGURE 9 - *Cartoon illustrating the geological evolution of eastern Brazil from late Paleoproterozoic to the Cambro-Ordovician (not to scale; descriptions and references in text). 1 - Mantiqueira-Itabuna Belt (M-I); 2 - Espinhaço Supergroup (E) and Borrachudos Suite (b); 3 - mantle plume (MP), mafic dyke swarms (d), and sequences of the Macaúbas Group (Mc) and West-Congo Supergroup (C); 4 - Salinas Formation (S) and Paraíba do Sul Complex (PS); 5 - precollisional calc-alkaline plutons (G1); 6 - slivers of oceanic lithosphere (v), syncollisional calc-alkaline plutons (G1), and the zone of crustal anatexis (G2 suite); 7 - magma underplating (MU), S-type granitic plutons (G4), and high-K calc-alkaline I-type plutons (G5).*

sedimentation and magmatism. Slivers of oceanic lithosphere yielded ages around 816 Ma.

- During the B-subduction-related (accretionary) orogenic stage (*c.* 625 - 595 Ma), a precollisional, calc-alkaline magmatic arc, recorded by G1 plutons, developed in a continental active margin. This arc was entirely inherited by the Araçuaí Belt after South Atlantic opening in Mesozoic time.

- During the collisional stage (*c.* 595 - 575 Ma), slivers of the Neoproterozoic oceanic lithosphere were thrusted upon the Salinas-Dom Silvério passive margin sequence. Following this, G1 plutons were emplaced and deformed, and crustal thickening triggered widespread anatexis (G2 suite). The late collisional to postcollisional G3-I and G3-S suites (*c.* 575 - 560 Ma) are not represented in the cartoon. They probably portray a transcurrent (docking) stage at the end of the orogeny. The time span from the late collisional to the late orogenic stage lasted for about 90 Ma. Progressive younger ages are given by the orogenic polarity, from N to S and from W to E. The gradual change from a compressive to an extensional tectonic regime and the onset of underplating of basic magma lasted longer in the southern than in the northern region.

- *Some 40 Ma after the end of the collisional stage*, during Cambro-Ordovician times (*c.* 535 - 490 Ma), the G4 and G5 plutonic suites crystallized. During this period, resetting of isotope systems was widespread in the Araçuaí and northern Ribeira belts. Local deformation was restricted to shear zones in the southernmost domain of the orogen. The regional reheating associated with new partial melting of the crust and coeval intrusion of basic magma was probably induced by mantle magma underplating beneath the South American Platform, reflecting the orogenic collapse that preceded the onset of large Paleozoic basins.

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REFERENCES

Achtschim, A.B., Pedrosa-mSoares, A.C., Marciano, V.R.P.R.O. and Daconti, B.C. (1998). Depósitos de água-marinha de Padre Paraíso, MG: Exemplo de pegmatitos derivados de granitóide cálcio-alcalino de alto-K. *SBG, Congresso Brasileiro Geologia, 40, Belo Horizonte, Anais*, 275.

Alkmim, F.F., Brito-Neves, B.B. and Castro-Alves, J.A. (1993). Arcabouço tectônico do Cráton do São Francisco. In: *O Cráton do São Francisco*. Salvador, Sociedade Brasileira de Geologia, eds. Dominguez, J.M.L. and Misi, A., pp. 45-62. Núcleo Bahia.

Alkmim, F.F., Chemale Jr., F. and Endo, I. (1996). A deformação das coberturas proterozóicas do Cráton do São Francisco e o seu significado tectônico. *Revista Escola de Minas*, 49(1), 22-38

Alkmim, F.F. and Marshak, S. (1998). Transamazonian orogeny in the southern São Francisco Craton region, Minas Gerais, Brazil: evidence for Paleoproterozoic collision and colapse in the Quadrilátero Ferrífero. *Precambrian Research*, 90, 29-58

Almeida, F.F.M. de (1977). O Cráton do São Francisco. *Revista Brasileira de Geociências*, 7, 349-364

Almeida, F.F.M. de, Amaral, G., Cordani, U.G. and Kawashita, K. (1973). The Precambrian evolution of the South American cratonic margin south of the Amazon River. In: *The Ocean Basin and Margins*, eds. Nairn, A.E.M. and Stehli, F.G., pp. 411-446, Plenum Publ. 1.

Almeida, F.F.M. de, Hasui, Y., Rodrigues, E.P. and Yamamoto, J. (1978). A Faixa de Dobramentos Araçuaí na região do Rio Pardo. *SBG, Congresso Brasileiro Geologia, 30, Recife, Anais*, 1, 270-283

Almeida, F.F.M., Litwinski, N. (1984). Província Mantiqueira: setor setentrional. In: *O Pré-Cambriano do Brasil*, eds. Almeida, F.F.M. and Hasui, Y., pp. 282-307. São Paulo, Editora Edgar Blücher.

Almeida-Abreu, P.A. (1993). *A Evolução Geodinâmica da Serra do Espinhaço Meridional, Minas Gerais, Brasil. Universitat Freiburg, Germany*. Dr. Thesis, 150 p.

Almeida-Abreu, P.A., Renger, F.E., Lindenberg, S.F., Silva, M.F.L. and Góes, H. (1997). A Bacia Macaúbas no nordeste da Serra do Espinhaço Meridional, Minas Gerais. *Boletim da Sociedade Brasileira Geologia-Núcleo Minas Gerais*, 14, 12-13

Aracema, L.W., França, A.V.M., Pedrosa-Soares, A.C., Noce, C.M., Ferreira, D.C. (1999). Granitóides cálcio-alcalinos do arco magmático neoproterozóico da Faixa Araçuaí: Evidências petrográficas e geoquímicas da região de Teófilo Otoni, MG, Brasil. *SBGq, Congresso de Geoquímica dos Países de Língua Portuguesa, Porto Seguro, 5, Anais*, 435-437

Barbosa, J.S.F. and Dominguez, J.M.L., eds. (1996). *Geologia da Bahia: Texto explicativo para o mapa geológico ao milionésimo*. Salvador, Brazil, Secretaria da Indústria Comércio e Mineração, Superintendência de Geologia e Recursos Minerais, 400 p.

Basílio, M.S., Dussin, I.A., Pedrosa-Soares, A.C., Jordt-Evangelista, H., Macambira, M., Voll, E. and Pimenta, V.B. (1998). Gênese dos depósitos de alexandrita de Malacacheta, Minas Gerais: evidências geológicas e geocronológicas. *SBG, Congresso Brasileiro Geologia, 40, Belo Horizonte, Anais*, 281.

Bayer, P., Horn, H., Lammerer, B., Schmidt-Thomé, R., Weber-Diefenbach, K. and Wiedemann, C. (1986). The Brasiliano Mobile Belt in southern Espírito Santo (Brazil) and its igneous intrusions. *Zentralblatt für Geologie und Palæontologie*, 1(9/10), 1429-1440.

Bilal, E., Nalini, H.A., Horn, H., Correia-Neves, J.M., Giret, A., Fuzikawa, K., Fernandes, M.L., Mello, F. and Moutte, J. (1998). Neoproterozoic granitoid suites of the Rio Doce region, Brazil. *UFOP, International Conference on Basement Tectonics, 14, Ouro Preto, Abstracts*, 41-43.

Bokhari, F.Y. and Kramers, J.D. (1981). Island arc character and late Precambrian age of volcanics at Wadi Shwas, Hijaz, Saudi Arabia: Geochemical and Sr and Nd isotope evidence. *Earth and Planetary Science Letters*, 54, 409-422.

Brandalise, L.A. (1991). *Programa Levantamentos Geológicos Básicos do Brasil. Ponte Nova. Folha SF 23-X-B-II. Estado do Espírito Santo*. Brasília, DNPM/CPRM, 122 p.

Brito-Neves, B.B. and Cordani, U.G. (1991). Tectonic evolution of

South America during the Late Proterozoic. *Precambrian Research*, **53**, 23-40.

Brito-Neves, B.B. and Alkmim, F.F. (1993). Cráton: evolução de um conceito. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and Misi, A. pp. 1-10. Salvador, Sociedade Brasileira de Geologia, Núcleo Bahia.

Brito-Neves, B.B., Sá, J.M., Nilson, A.A. and Botelho, N.F. (1995). A tafrogênese estateriana nos blocos paleoproterozóicos da América do Sul e processos subsequentes. *Geonomos*, **3**(2), 1-21

Brito-Neves, B.B. and Sato, K. (1998). Chronological milestones in the evolution of the South American Continent. *UFOP, International Conference on Basement Tectonics, 14 Ouro Preto, Abstracts*, 2-4

Brueckner, H., Cunningham, W.D., Alkmim, F.F. and Marshak, S. (1998). Implications of new Precambrian Sm-Nd dates from the Quadril·tero FerrÌfero and adjacent northern Mantiqueira Province. *UFOP, International Conference on Basement Tectonics, 14 Ouro Preto, Abstracts*, 24-25

Campos-Neto, M.C. and Figueiredo, M.C.H. (1995). The Rio Doce Orogeny, Southeastern Brazil. *Journal of South American Earth Sciences*, **8**, 143-162

Campos-Neto, M.C., Caby, R., Janasi, V.A., Basei, M.A.S., Garcia, M.G.M. and Siga Jr., O. (1999). Transition from continental to intracontinental subduction, south of São Francisco Craton: terrane accretion, metamorphic paths and nappe kinematics. *SBG, Simpósio Nacional de Estudos Tectônicos, Lençóis, Anais*, 7, Sessão 2, 66-70.

Carvalho, J.B. and Pereira, L.M.M. (1997a). Metamorfismo das rochas supracrustais da porção leste do Estado de Minas Gerais. *Boletim da Sociedade Brasileira Geologia-Núcleo Minas Gerais*, **14**, 101-102

Carvalho, J.B. and Pereira, L.M.M. (1997b). Caracterização petrológica das rochas graníticas e charnockíticas da porção leste do Estado de Minas Gerais, baseada em dados petrográficos. *Boletim da Sociedade Brasileira Geologia-Núcleo Minas Gerais*, **14**, 98-100

Celino, J.J. (1999). *Variação composicional em suites de granitóides neoproterozóicos e sua implicação na evolução do Orógeno Araçuaí (Brasil)-Oeste Congolês (África)*. Instituto de Geociências, Universidade de Brasília, Brasil, Dr. thesis, 266 p.

Claesson, S., Pallister, J.S. and Tatsumoto, M. (1984). Samaryum-neodymium data on two Late Proterozoic ophiolites of Saudi Arabia and implications for crustal and mantle evolution. *Contributions to Mineralogy and Petrology*, **85**, 244-252.

Condie, K.C. (1993). *Plate Tectonics and crustal evolution*. 3rd ed. Oxford, Pergamon Press, 492 p.

Correa-Gomes, L.C. and Oliveira, E.P. (1997). O enxame radial gigante de diques máficos da província Bahia-Congo: Implicações reológicas e tectônicas da presença de uma pluma mantélica na interface América do Sul-África, 1.0 Ga atrás. *SBG, Simpósio Nacional de Estudos Tectônicos, Pirenópolis, Anais*, 5, 52-55.

Correia-Neves, J.M., Pedrosa-Soares, A.C. and Marciano, V.R.P.R.O. (1986). A Província PegmatÌtica Oriental do Brasil à luz dos conhecimentos atuais. *Revista Brasileira de Geociências*, **16**, 106-118.

Costa, A.G. (1989). Evolução petrológica para uma sequência de rochas metamórficas regionais do tipo baixa pressão na região de Itinga, NE de Minas Gerais. *Revista Brasileira de Geociências*, **19**, 440-448.

Costa, A.G. (1998). The granulite-facies rocks of the northern segment of the Ribeira Belt, eastern Minas Gerais, SE Brazil. *Gondwana Research*, **1**, 367-372.

Costa, A.G., Rosière, C.A., Lobato, L.M. and Laureano, F.V. (1993). Evolução petrológica e estrutural da porção oriental do Estado de Minas Gerais e suas implicações geotectônicas. *Geonomos*, **1**, 39-50.

Costa, A.G., Rosière, C.A., Ebert, H.D., Fischel, D.P., Fuck, R.A. and Pimentel, M.M. (1998). The relation between frontal and strike-slip shear zones and the regional metamorphism and magmatism in a deep-crustal segment of the northern Ribeira Belt, SE Brazil. *UFOP, International Conference on Basement Tectonics, 14, Ouro Preto, Abstracts*, 26-28.

CPRM-SGA (1997). *Mapa magnetométrico da região entre 14°-22°30' S e 40°-44° W*. Serviço de Geofísica Aplicada, Companhia de Pesquisa de Recursos Minerais, Belo Horizonte.

Crocco-Rodrigues, F.A., Guimarães, M.L.V., Abreu, F.R., Oliveira, O.A.B. and Greco, F.M. (1993). Evolução tectônica do Bloco Itacambira-Monte Azul entre Barrocão e Porteirinha, Minas Gerais. *SBG, Simpósio Nacional de Estudos Tectônicos, Belo Horizonte, Anais*, 4, 212-216.

Cunningham, W.D., Marshak, S. and Alkmim, F.F. (1996). Structural style of basin inversion at mid-crustal levels: two transects in the internal zone of the Brasiliano Araçuaí Belt, Minas Gerais, Brazil. *Precambrian Research*, **77**, 1-15

Cunningham, W.D., Alkmim, F.F. and Marshak, S. (1998). A structural transect across the Coastal Mobile belt in the Brazilian Highlands (latitute 20× S): the roots of a Precambrian transpressional orogen. *Precambrian Research*, **92**, 251-275

D'Agrella-Filho, M.S., Pacca, I.G., Teixeira, W., Onstott, T.C. and Renne, P.R. (1990). Paleomagnetic evidence for the evolution of Meso to Neoproterozoic glaciogenic rocks in central-eastern Brazil. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **80**, 255-265.

D'El-Rey-Silva, L.J.H., Pedrosa-Soares, A.C. and Fuck, A.R. (1995). Kinematic analysis in the Neoproterozoic Araçuaí Belt, SE Brazil: tectonic implications for Gondwana Assembly. *Boletim da Sociedade Brasileira Geologia-Núcleo Minas Gerais*, **13**, 61-63.

Dalziel, I.W.D. (1997). Neoproterozoic-Paleozoic geography and tectonics: review, hypothesis, environmental speculation. *Geological Society of America Bulletin*, **109**, 16-42.

Delhal, J., Ledent, D. and Cordani, U.G. (1969). Ages Pb/U, Sr/Rb et Ar/K de formations mÈtamorphiques et granitiques du Sud-Est du Brésil (États de Rio de Janeiro et de Minas Gerais). *Annales de la Societé Géologique de Belgique*, **92**, 271-283.

Djama, L.M., Leterrier, J. and Michard, A. (1992). Pb, Sr and Nd isotope study of the basement of the Mayumbian Belt (Guena gneisses and Mfoubou granite, Congo): Implications for crustal evolution in Central Africa. *Journal of African Earth Sciences*, **14**, 227-237.

Dossin, I.A. (1994). *Evolution structurale de la region de L'Espinhaço Meridional, bordure sud-est du Craton São Francisco, Brésil: tectoniques superposées au Protérozoïque*. Université d'Orléans, Orléans, France, Dr. thesis, 191 p.

Dossin, I.A., Dossin, T.M., Charvet, J., Cocherie, A. and Rossi, P. (1993). Single-zircon dating by step-wise Pb-evaporation of Middle Proterozoic magmatism in the Espinhaço Range, southeastern São Francisco Craton (Minas Gerais, Brazil). *SBG-Núcleo Bahia, Simpósio sobre o Cráton São Francisco, 2, Salvador, Anais*, 39-42

Dussin, I.A. and Dussin, T.M. (1995). Supergrupo Espinhaço: modelo de evolução geodinâmica. *Geonomos*, **3**, 19-26.

Dussin, I.A., Dussin, T.M., Vieira, V.S. and Macambira, M.J.B. (1998). Idade do Maciço Lagoa Preta (Suíte Aimorés, MG-ES): determinação de $^{207}Pb/^{206}Pb$ sobre zircões de hiperstênio granito. *SBG, Congresso Brasileiro Geologia, 40, Belo Horizonte, Anais*, 34.

Ebert, H. (1968). Ocorrências da fácies granulÌtica no sul de Minas Gerais e em áreas adjacentes, em dependência da estrutura orogênica: hipóteses sobre sua origem. *Anais da Academia Brasileira de Ciências*, **40** (Suplemento), 215-229.

Ebert, H. and Hasui, Y. (1998). Transpressional tectonics and strain partitioning during oblique collision between three plates in the Precambrian of southeast Brazil. In: *Continental Transpressional and Transtensional Tectonics*, eds. R.E. Holdsworth, R.A. Stranchan and J.F. Dewey. *Geological Society London, Spec. Public.*, **135**, 231-252

Faria, L.F. de (1997). *Controle e tipologia de mineralizações de grafita* flake *do nordeste de Minas Gerais e sul da Bahia: uma abordagem regional.* Instituto de Geociências, Universidade Federal de Minas Gerais, Belo Horizonte, Brasil, M.Sc. thesis, 102 p.

Féboli, W.L. (org.) (1993). *Programa Levantamentos Geológicos Básicos do Brasil. Piuma. Folha SF 24-V-A-VI. Estado do Espírito Santo.* Brasília. DNPM/CPRM. 140 p.

Fernandes, M.L.S. (1991). *Geologia, petrografia e geoquímica de rochas granitóides da região de Pedra Azul, MG.* Instituto de Geociências, Universidade Federal do Rio de Janeiro, Brasil, Ms thesis, 191 p.

Fernandes, M.L.S., Bilal, E., Wiedemann, C. and Correia-Neves, J.M. (1999). O registro da deformação brasiliana em um corpo da Suíte Borrachudos aflorante na região de Dores de Guanhães, MG: evidência geocronológica. *SBG, Simpósio Nacional de Estudos Tectônicos, Lençóis, Anais, 7, Sessão* 2, 34-37.

Fernandes, M.L.S., Pedrosa-Soares, A.C., Noce, C.M., Wiedemann, C. and Correia-Neves, J.M. (1999). U-Pb geochronology of the Borrachudos Suite: Evidence of Brasiliano tectonism recorded by late Paleoproterozoic anorogenic granites (Araçuaí Belt, Minas Gerais). *31 International Geologic Congress.* (Submitted to).

Figueiredo, M.C.H. and Teixeira, W. (1996). The Mantiqueira Metamorphic Complex, eastern Minas Gerais State: preliminary geochronological and geochemical results. *Anais da Academia Brasileira de Ciências*, **68**, 1-24

Fischel, D.P. (1998). *Geologia e dados isotópicos Sm-Nd do Complexo Mantiqueira e do Cinturão Ribeira na região de Abre Campo, Minas Gerais.* Instituto de Geociências, Universidade de Brasília, Brasil, M.Sc. thesis, 99 p.

Fonseca, M.J.G. (1998). *Mapa Geológico do Estado do Rio de Janeiro.* Rio de Janeiro, Departamento Nacional da Produção Mineral (DNPM), 141 p.

Fritzer, T. (1991). *Das Guaçuí Lineament und die orogene Entwicklung des Zentralen Ribeira Belt (Espírito Santo, Brasilien).* Münchener Geowissenschaftliche Abhandlungen, 2, Munich, Germany, Friedrich Pfeil Verlag, 196 p.

Geiger, A. (1993). *Die Geologie des Steinbruchreviers von Cachoeiro de Itapemirim Südliches (Espírito Santo, Brasilien).* Münchener Geologische Hefte, 11, Munich, Germany, Friedrich Pfeil Verlag, 217 p.

Grossi-Sad, J.H. and Dutra, C.V. (1988). Chemical composition of supracrustal rocks from Paraíba do Sul Group, Rio de Janeiro State, Brazil. *Geochimica Brasiliensis*, **2**, 143-165

Grossi-Sad, J.H., Motta, E. and Baars, F.J. (1993). Formação Capelinha: uma nova entidade litoestratigráfica do Grupo Macaúbas Superior. *SBG-MG, Simpósio de Geologia de Minas Gerais, 7, Belo Horizonte, Anais*, 30-33

Grossi-Sad, J.H., Lobato, L.M., Pedrosa-Soares, A.C. and Soares-Filho, B.S., eds. (1997). *Projeto Espinhaço em CD-ROM.* Belo Horizonte, Brazil, Companhia Mineradora de Minas Gerais, 2693 p.

Haralyi, N.L.E. and Hasui, Y. (1982). The gravimetric information and the Archean-Proterozoic structural framework of Eastern Brazil. *Revista Brasileira de Geociências*, **12**, 160-166.

Hasui, Y., Carneiro, C.D.R. and Coimbra, A.M. (1975). The Ribeira Fold Belt. *Revista Brasileira de Geociências*, **5**, 257-266.

Hasui, Y. and Oliveira, M.A.F. (1984). Província Mantiqueira: setor central. In: *O Pré-Cambriano do Brasil*, eds. Almeida, F.F.M. and Hasui, Y., pp. 308-344. São Paulo, Editora Edgar Blücher.

Heilbron, M., Valeriano, C.M., Valladares, C.S. and Machado, N. (1995). A orogênese brasiliana no segmento central da Faixa Ribeira, Brasil. *Revista Brasileira de Geociências*, **25**, 249-266.

Heilbron, M., Valeriano, C.M., Tupinambá, M. and Almeida, J.C.H. (1999). An evolutionary tectonic model for the central segment of Ribeira Belt: From the Transamazonian collage to Gondwana amalgamation. *SBG, Simpósio Nacional de Estudos Tectônicos, Lençóis, Anais, 7, Sessão* 2, 58-61.

Hoffman, P.F., Kaufman, A.J., Halverson, G.P. and Schrag, D.P. (1998). A Neoproterozoic snowball Earth. *Science*, **281**, 1342-1346.

Jordt-Evangelista, H. (1992). O Grupo Dom Silvério a SE de Minas Gerais: petrografia, metamorfismo e geologia econômica. *Revista Escola de Minas de Ouro Preto*, **45**(1/2), 140-142.

Karfunkel, J. and Hoppe, A. (1988). Late Proterozoic glaciation in central-eastern Brazil: Synthesis and model. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **65**, 1-21.

Klumb, A. and Leite, C.A. (1999). *Folha Ipatinga.* Projeto Leste, CPRM, Belo Horizonte.

Lammerer, B. (1987). Short notes on a structural section through the Ribeira Mobile Belt (Minas Gerais and Espírito Santo, Brazil). *Zentralblatt für Geologie und Palæontologie*, **7/8**, 719-728.

Leonardos, O.H. and Fyfe, W.S. (1974). Ultrametamorphism and melting of a continental margin: the Rio de Janeiro region, Brasil. *Contributions to Mineralogy and Petrology*, **46**, 201-214.

Lima, M.I.C., Silva, J.M.R. and Siga Jr., O. (1981). Levantamento de Recursos Naturais, Folha SD 24, Salvador: Geologia. *Projeto RADAMBRASIL, Rio de Janeiro, IBGE*, **24**, 25-192.

Ludka, I.P., Wiedemann, C., Lammerer, B. and Toepffer, C. (1998). On the origin of incompatible element enrichment in the Venda Nova pluton, State of Espírito Santo, SE Brazil. *Journal of South American Earth Sciences*, **11**, 473-486.

Machado, L, Ribeiro, M.W., Gonzalez, S.R., Schenini, C.A, Santos-Neto, A., Palmeira, R.C.B., Pires, J.L., Teixeira, W. and Castro, H.E.F. (1983). Levantamento de Recursos Naturais, Folha SF 23/24, Rio de Janeiro/Vitória: Geologia. *Projeto RADAMBRASIL, Rio de Janeiro, IBGE*, **32**, 27-304.

Machado, N. (1998). The Ribeira Belt as a continental rift in the Congo-Franciscan plate. *UFOP, International Conference on Basement Tectonics, 14, Ouro Preto, Abstracts*, 29-31.

Machado, N., Schrank, A., Abreu, F.R., Knauer, L.G. and Almeida-Abreu, P.A. (1989). Resultados preliminares da geocronologia U-Pb na Serra do Espinhaço Meridional. *Boletim da Sociedade Brasileira Geologia-Núcleo Minas Gerais*, **10**, 171-174.

Machado, N., Valladares, C., Heilbron, M. and Valeriano, C. (1996). U-Pb geocronology of the central Ribeira Belt (Brazil) and implications for the evolution of the Brasiliano Orogeny. *Precambrian Research*, **79**, 347-361.

Marshak, S. and Alkmim, F.F. (1989). Proterozoic contraction/extension tectonics of the southern São Francisco region, Minas Gerais, Brazil. *Tectonics*, **8**, 555-571.

Martins-Neto, M.A. (1998a). O Supergrupo Espinhaço em Minas Gerais: Registro de uma bacia rift-sag do Paleo/Mesoproterozóico. *Revista Brasileira de Geociências*, **28**, 151-168.

Martins-Neto, M.A. (1998b). Mantle plume, rifting and the early Neoproterozoic glaciation in the São Francisco Craton and Araçuaí Belt, southeastern Brazil. *UFOP, International Conference on Basement Tectonics, 14, Ouro Preto, Abstracts*, 32-34.

Medeiros, S.R. (1999). *Estudo mineralógico, petrológico, geoquímico e isotópico do Complexo Intrusivo de Várzea Alegre (ES).* Instituto de Geociências, Universidade Federal do Rio de Janeiro, Dr. thesis, 174 p.

Medeiros, S.R. and Wiedemann, C. (1999). Condições de cristalização (P e T) de magmas cálcio-alcalinos de médio e alto K no Cinturão Ribeira: Estudo do complexo inrusivo de Várzea Alegre (ES). *SBGq, Congresso de Geoquímica dos Países de Língua Portuguesa, Porto Seguro, 5, Anais*, 552-554.

Mendes, J.C, Wiedemann, C. and McReath, I. (1999). Charnockitic magmatic rocks from the Várzea Alegre Massif, Espírito Santo, Southeast Brazil: Conditions of Formation. *Revista Brasileira de Geociências*, **29** (in press).

Mendes, J.C., McReath, I., Wiedemann, C.M. and Figueiredo, M.C.H. (1997). Charnockitóides do Maciço de Várzea Alegre: um novo exemplo do magmatismo cálcio-alcalino de alto-K no arco magmático do Espírito Santo. *Revista Brasileira de Geociências*, **27**, 13-24.

Monteiro, R.L.B.P., Leonardos, O.H. and Correia-Neves, J.M. (1990). An epigenetic origin for the new scheelite and wolframite occurrences in the Middle Jequitinhonha Valley, Minas Gerais, Brazil. *Revista Brasileira de Geociências*, **20**, 68-74.

Nalini, H. A. (1997). *Caractérization des suites magmatiques néoprotérozoïques de la région de Conselheiro Pena et Galiléia, Minas Gerais, Brésil.* École Nationale Superière des Mines, Saint Etienne, France, Dr. thesis, 237 p.

Nalini, H.A., Bilal, E. and Correia-Neves, J.M. (1998). Mineralogical, geochemical and isotope constrains of Neoproterozoic granitoids (Urucum and Galiléia suites), eastern Minas Gerais State, Brazil. *UFOP, International Conference on Basement Tectonics, 14, Ouro Preto, Minas Gerais. Abstracts*, 44-46.

Noce, C.M., Pedrosa-Soares, A.C., Grossi-Sad, J.H., Baars, F.J., Guimarães, M.L.V., Mourão, M.A.A., Oliveira, M.J.R. and Roque, N.C. (1997). Nova subdiviSão estratigráfica regional do Grupo Macaúbas na Faixa Araçuaí. *Boletim da Sociedade Brasileira Geologia-Núcleo Minas Gerais*, **14**, 29-31.

Noce, C.M., Machado, N. and Teixeira, W. (1998). U-Pb geochronology of gneisses and granitoids in the Quadrilátero Ferrífero (southern São Francisco Craton): age constraints for Archean and Paleoproterozic magmatism and metamorphism. *Revista Brasileira de Geociências*, **28**, 95-102.

Noce, C.M., Macambira, M.J.B., Pedrosa-Soares, A.C., Martins, V.T.S. and Ferreira, D.C. (1999). Chronology of Late Proterozoic-Cambrian granitic magmatism in the Araçuaí belt, Eastern Brazil, based on dating by single zircon evaporation. *Servicio Geologico Minero Argentino, South American Symposium on Isotope Geology, 2, Vila Carlos Paz, Argentina, Anaes*, 86-89.

Oliveira, M.J.R., Pinto, C.P. and Féboli, W.L. (1997). Integração de dados estruturais do leste mineiro. *Boletim da Sociedade Brasileira Geologia-Núcleo Minas Gerais*, **14**, 72-73.

Paes, V.J.C. (1999). *Geologia e geoquímica da região de Alvarenga, Minas Gerais: Implicacões geotectônicas e metalogenéticas.* Instituto de Geociências, Universidade Federal de Minas Gerais, Belo Horizonte, Brasil, M.Sc. thesis, 149 p.

Paes, V.J.C., Lobato, L.M., Dussin, I.A., Baars, F.J., Pimentel, M.M. and Noce, C.M. (1999). 450 m.y. of Neoproterozoic orogenic lithospheric evolution of southern Araçuaí Belt within ancient continental crust: geological, geochronological and geochemical constraints, Alvarenga region, Minas Gerais State, Brazil. Submitted to *31 International Geologic Congress.*

Park, R.G. (1993). *Geological structures and moving plates.* Glasgow, Blackie Academic and Professional Publ., 337 p.

Pearce, J. (1996). Sources and settings of granitic rocks. *Episodes*, **19**, 120-125.

Pearce, J.A., Gorman, B.E. and Birkett, T.C. (1977). The relationship between major element chemistry and tectonic environment of basic and intermediate volcanic rocks. *Earth and Planetary Science Letters*, **36**, 121-132

Pedreira, A.J., Vieira, V.S. and Signorelli, N. (1997). Turbiditos do setor oriental da Faixa Araçuaí, Minas Gerais. *A Terra em Revista*, **3**, 28-33

Pedrosa-Soares, A.C. (1995). *Potencial aurífero do Vale do Araçuaí, Minas Gerais: história da exploração, geologia e controle tectono-metamórfico.* Instituto de Geociências, Universidade de Brasília, Brasil, Dr. thesis, 177 p.

Pedrosa-Soares, A.C., Monteiro, R.L.B.P., Correia-Neves, J.M., Leonardos, O.H. and Fuzikawa, K. (1987). Metasomatic evolution of granites, northeast Minas Gerais, Brazil. *Revista Brasileira de Geociências*, **17**, 512-518 (Errata for geochemical data tables in v. 18, p. 319. 1988).

Pedrosa-Soares, A.C., Noce, C.M., Vidal, Ph., Monteiro, R.L.B.P. and Leonardos, O.H. (1992). Toward a new tectonic model for the Late Proterozoic Araçuaí (SE Brazil)-West Congolian (SW Africa) Belt. *Journal of South American Earth Sciences*, **6**, 33-47.

Pedrosa-Soares, A.C., Baars, F.J., Lobato, L.M., Magni, M.C.V. and Faria, L.F. (1993). Arquitetura tectono-metamórfica do setor central da Faixa Araçuaí e suas relações com o Complexo Guanhães. *SBG, Simpósio Nacional de Estudos Tectônicos, Belo Horizonte, Anais*, **4**, 176-182.

Pedrosa-Soares, A.C. and Pedreira, A.J. (1996). O setor setentrional da Faixa Araçuaí. *SBG, Congresso Brasileiro Geologia, 39, Salvador, Roteiro de Excursão*, E12, 58 p.

Pedrosa-Soares, A.C., Leonardos, O.H., Ferreira, J.C.H. and Reis, L.B. (1996). Duplo regime metamórfico regional na Faixa Araçuaí (?): uma reavaliação à luz de novos dados. *SBG, Congresso Brasileiro de Geologia, 39, Salvador, Anais*, **6**, 5-8.

Pedrosa-Soares, A.C. and Noce, C.M. (1998). Where is the suture zone of the Neoproterozoic Araçuaí-West-Congo orogen? *UFOP, International Conference on Basement Tectonics, 14, Ouro Preto, Abstracts*, 35-37.

Pedrosa-Soares, A.C., Vidal, Ph., Leonardos, O.H. and Brito-Neves, B.B. (1998). Neoproterozoic oceanic remnants in eastern Brazil: Further evidence and refutation of an exclusively ensialic evolution for the Araçuaí-West Congo Orogen. *Geology*, **26**, 519-522.

Pedrosa-Soares, A.C. and Noce, C.M. (1999). The Araçuaí-West-Congo Orogen: Alpine-, Himalayan-type, or neither? *SBG, Simpósio Nacional de Estudos Tectônicos, Lençóis, Anais*, **7**, Sessão 2, 30-33.

Pedrosa-Soares, A.C., Wiedemann, C.M., Fernandes, M.L.S., Faria, L.F. and Ferreira, J.C.H. (1999a). Geotectonic significance of the Neoproterozoic granitic magmatism in the Araçuaí belt, Eastern Brazil: a model and pertinent questions. *Revista Brasileira de Geociências*, **29**, 57-64.

Pedrosa-Soares, A.C., Noce, C.M., Wiedemann, C.M. and Pinto, C.P. (1999b). The Neoproterozoic Araçuaí-West-Congo Orogen in Brazil: Did it start as B-subduction-controlled magmatic arc? *Precambrian Research* (submitted to).

Pedrosa Soares, A.C., Cordani, U.G. and Nutman, A. (1999c). Constraining the age of Neoproterozoic glaciation in eastern Brazil: First U-Pb data from detrital zircons. *Terra Nova.* (submitted to).

Pinto, C.P., Drumond, J.B.V. and Féboli, W.L., coords. (1997). *Projeto Leste: mapeamento geológico, cadastramento de recursos minerais da região leste de Minas Gerais - Etapa 1.* Belo Horizonte, Brazil, Companhia de Pesquisa de Recursos Minerais, 161 p.

Pinto, C.P., Pedrosa-Soares, A.C. and Wiedemann, C. (1998). Mapa geológico da porção brasileira do Orógeno Araçuaí-Oeste-Congo. *SBG, Congresso Brasileiro de Geologia, 40, Belo Horizonte, Abstracts*, 37.

Reis, L.B. (1999). *Estudos de mineralizações de grafita no extremo nordeste de Minas Gerais.* Instituto de Geociências, Universidade Federal de Minas Gerais, Belo Horizonte, Brasil, M.Sc. thesis, 87 p.

Ribeiro, J.H. (1999). *Folha Dom Cavati.* Projeto Leste, CPRM, Belo Horizonte.

Rogers, J.J.W. (1996). A history of continents in the past three billion years. *Journal of Geology*, **104**, 91-107.

Saunders, A.D. (1984). The rare earth element characteristics of igneous rocks from the ocean basins. In: *Rare Earth Element*

Geochemistry, ed. Henderson, P., pp. 205-236. Amsterdam, Elsevier.

Seidensticker, U. and Wiedemann, C.M. (1992). Geochemistry and origin of lower crustal granulite facies rocks in the Serra do Caparaó region, Espírito Santo/Minas Gerais, Brazil. *Journal of South American Earth Sciences*, **6**, 289-298.

Sengor, A.M.C. (1990). Plate Tectonics and orogenic research after 25 years: a Tethyan perspective. *Earth-Science Reviews*, **27**, 1-201.

Schobbenhaus, C. 1996. As tafrogêneses superpostas Espinhaço e Santo Onofre, Estado da Bahia. *Revista Brasileira de Geociências*, **26**, 265-276.

Schobbenhaus, C., Campos, D.A., Derze, G.R. and Asmus, H.E., coords. (1981). *Mapa Geológico do Brasil e da Área Oceânica adjacente, incluindo depósitos minerais, escala 1:2.500.000*. Brasília, Departamento Nacional da Produção Mineral (DNPM).

Siga, O. (1986). *A evolução geocronológica da porção nordeste de Minas Gerais, com base em interpretações geocronológicas*. Instituto de Geociências, Universidade de São Paulo, São Paulo, Brasil, M.Sc. thesis, 140 p.

Signorelli, N., org. (1993). *Programa Levantamentos Geológicos Básicos do Brasil. Afonso Cláudio. Folha SF. 24-V-A-II*. Brasília, DNPM/CPRM, 153 p.

Silva, J.M.R., Lima, M.I.C., Veronese, V.F., Ribeiro JR, R.N., Rocha, R.M. and Siga Jr., O. (1987). Levantamento de Recursos Naturais, Folha SE 24, Rio Doce: Geologia. *Projeto RADAMBRASIL, Rio de Janeiro, IBGE*, **34**, 23-172.

Silva, L.C. (1999). *Geocronologia U-Pb SHRIMP e Sm-Nd na Província Mantiqueira Meridional, no Cinturão Saldania (África do Sul) e a evolução do Ciclo Brasiliano-Pan Africano*. Instituto de Geociências, Universidade Federal do Rio Grande do Sul, Porto Alegre, Brasil, Dr. thesis, 243 p.

Silva, S.L. (1999). *Folha Coronel Fabriciano*. Projeto Leste, CPRM, Belo Horizonte.

Sluitner, Z. (1990). *Petrographie und Geochemie von hochgradig metamorphen suprakrustalen Gesteinsserien des Brasiliano im südlichen Espírito Santo / Brasilien unter besonderer Berücksichtigung der enderbitischen Gesteinen*. Institut für Allgemeine und Angewandte Geologie, Universität München, Germany, *Dr. thesis, 191 p.*

Sluitner, Z. and Weber-Diefenbach, K. (1989). Geochemistry of charnoenderbitic granulites and associated amphibolitic gneisses in the coast region of Espírito Santo, Brazil. *Zentralblatt für Geologie und Palæontologie*, **1**(5/6), 917-931.

Söllner, H.S., Lammerer, B. and Weber-Diefenbach, K. (1989). *Brasiliano age of a charnoenderbitic rock suite in the* Complexo Costeiro (Ribeira Mobile Belt), Espírito Santo, Brazil: Evidence from U-Pb geochronology on zircons. *Zentralblatt für Geologie und Palæontologie*, **1**(5/6), 933-945.

Söllner, H.S., Lammerer, B. and Weber-Diefenbach, K. (1991). Die Krustenentwicklung in der Küstenregion nördlich von Rio de Janeiro/Brasilien. *Münchner Geologische Hefte*, **4**, *100 p.*

Söllner, H.S., Lammerer, B. and Wiedemann, C. From Coastal area to the São Francisco craton - age data on a profile covering the Ribeira mobile belt north of Rio de Janeiro/Brazil (in preparation.).

Söllner, F., Lammerer, B. and Wiedemann-Leonardos, C.M. Dating the Ribeira Mobile Belt of Brazil. In: *Sonderheft der Zeitschrift für Angewandte Geologie 2000: Technical Cooperation with Latin-America* (in press).

Tack, L. and Fernandez-Alonso M.J. (1998). The West Congolian Belt: a critical assessment of available time constraints during the Neoproterozic and Paleozoic amalgamation of Gondwana. *Journal of African Earth Sciences*, **27**, 193.

Teixeira, W., Kamo, S.L. and Arcanjo, J.B.A. (1997). U-Pb zircon and baddeleyite age and tectonic Interpretation of the Itabuna Alkaline Suite, São Francisco Craton, Brazil. *Journal of South American Earth Sciences*, **10**, 91-98.

Trompette, R. (1994). *Geology of Western Gondwana (2000-500 Ma)*. Amsterdam, Balkema, 350 p.

Trompette, R. (1997). Neoproterozoic (~600 Ma) aggregation of Western Gondwana: A tentative scenario. *Precambrian Research*, **82**, 101-112.

Tupinambá, M., Teixeira, W., Heilbron, M. and Basei, M. (1998). The Pan-African/Brasiliano arc-related magmatism at the Costeiro domain of the Ribeira Belt, southeasthern Brazil: new geochronological and geochemical data. *UFOP, International Conference on Basement Tectonics, 14, Ouro Preto, Abstracts*, 12-14

Tuller, M.P., org. (1993). *Programa Levantamentos Geológicos Básicos do Brasil. Colatina. Folha SE. 24-Y-C-VI*. Brasília, DNPM/CPRM, 163 p.

Uhlein, A. (1991). *Transição craton-faixa dobrada: Exemplo do Craton do São Francisco e da Faixa Araçuaí (Ciclo Brasiliano) no Estado de Minas Gerais (aspectos estratigráficos e estruturais)*. Instituto de Geociências, Universidade de São Paulo, São Paulo, Brasil, Dr. thesis, 295 p.

Uhlein, A., Trompette, R. and Egydio-Silva, M. (1998a). Proterozoic rifting and closure, SE border of the São Francisco Craton, Brazil. *Journal South American Earth Science*, **11**, 191-203.

Uhlein, A., Egydio-Silva, M., Bouchez, J.L. and Vauchez, A. (1998b). The Rubim pluton (Minas Gerais, Brazil): a petrostructural and magnetic fabric study. *Journal of South American Earth Science*, **11**, 179-189.

Uhlein, A., Egydio-Silva, M., Trompette, R., Vauchez, A. and Kattah, C. (1999). Paleogeografia e inversão tectônica da Faixa Araçuaí. *SBG, Simpósio Nacional de Estudos Tectônicos, Lençóis, Anais*, **7**, *Sessão* **2**, 26-29.

Ussami, N., Sá, N.C. and Molina, E.C. (1993). Gravity map of Brazil. 2. Regional and residual isostatic anomalies and their correlation with major tectonic provinces. *Journal of Geophysical Research*, **98**, 2199-2208.

Valladares, C.S., Heilbron, M., Figueiredo, M.C.H. and Teixeira, W. (1997). Geochemistry and geochronology of Paleoproterozoic gneissic rocks of the Paraíba do Sul Complex (Quirino Unit), Barra Mansa region, Rio de Janeiro, Brazil. *Revista Brasileira de Geociências*, **27**, *111-120.*

Vernikovskiy, V.A., Vernikovskaya, A.E. and Chernykh, A.I. (1998). Neoproterozoic Taymyr ophiolitic belts and opening of the Paleo-Pacific Ocean. *International Geology Review*, **40**, 528-538.

Vicat, J.P. and Pouclet, A. (1995). Nature du magmatisme lié à une extension pré-panafricaine: les dolérites des bassins de Comba et de Sembé-Ouesso (Congo). *Bulletin de la Société Géologique de France*, **166**, 355-364.

Vieira, V.S., org. (1993). *Programa Levantamentos Geológicos Básicos do Brasil. Baixo Guandu. Folha SE. 24-Y-C-V.* Brasília, DNPM/CPRM, 175 p.

Vieira, V.S., org. (1997). *Programa Levantamentos Geológicos Básicos do Brasil. Cachoeiro do Itapemirim. Folha SF. 24-V-A.* Brasília, DNPM/CPRM, 99 p.

Wiedemann, C. (1993). The evolution of the Early Paleozoic, late to postcollisional magmatic arc of the Coastal Mobile Belt, in the State of Espírito Santo, eastern Brazil. *Anais Academia Brasileira de Ciências*, **65**, 163-181

Wiedemann, C., Mendes, J.C., Moura, J.C., Costa-Nascimento, R.C. and Ludka, I.P. (1997). Granitoids of the Espírito Santo Magmatic Arc. *SBG, International Symposium on Granites and Associated Mineralizations, 2, Salvador. Excursions Guide*, 57-76.

Wiedemann, C., Lammerer, B., Ludka, I.P., Mendes, J.C. and Medeiros, S.R. Evolution of the Ribeira Belt in Espírito Santo, Brazil: the roots of a Neoproterozoic (Brasiliano) mountain range (in preparation).

Wilson, M. (1989). *Igneous Petrogenesis*. London, Unwin Hyman, 466 p.

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THE CENTRAL SEGMENT OF THE RIBEIRA BELT

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The Ribeira Belt, of Neoproterozoic to Early Paleozoic age, extends for 1400 km approximately NE-SW along the southeastern coast of Brazil (Fig 1). A Gondwana reconstruction shows that the Ribeira Belt is part of a larger orogenic system developed in response to the convergence of the São Francisco, Congo and an inferred, third cratonic block presently hidden under the Paleozoic sequences of the Paraná Basin (Brito Neves and Cordani, 1991; Campos Neto and Figueiredo, 1995; Heilbron *et al.*, 2000). To the S, the Ribeira Belt is limited by the Luiz Alves Craton and both are covered by the Phanerozoic successions of the Paraná Basin. To the N there is a lateral transition to the Neoproterozoic Araçuai Belt where the orogen assumes a predominant N-S trend. To the NW, in southern Minas Gerais State, the NE structural trend of the Ribeira Belt overprints the previously developed NNW trend of the Neoproterozoic Brasília Belt, resulting in a complex interference zone between the two belts (Trouw *et al.*, 1994).

The name Ribeira Belt comes from a geotectonic unit that crops out in the Ribeira River valley, in São Paulo State (Almeida *et al.*, 1973). The name was later extended to comprise the presently known limits of the belt. The Ribeira Belt has also been called the Atlantic Mobile or Fold Belt (Leonardos and Fyfe, 1974), the Southeastern Fold Region (Almeida *et al.*, 1976), the Ribeira-Mantiqueira Fold Belt (Trompette, 1994) and, in its northern part, the Coastal Mobile Belt (Wiedemann, 1993; Cunningham *et al.*, 1998). The belt coincides to a large extent with the Mantiqueira Province of Almeida and Hasui (1984).

The Ribeira Belt is part of a continuous network of Neoproterozoic to Early Paleozoic mobile belts generated between 700 and 450 Ma, during the amalgamation of the Gondwana Supercontinent (Trompette, 1994; Unrug, 1997). In South America and Africa these orogenic events are referred to as the Brasiliano and Pan African orogenies, respectively. The main period of orogenic activity in the Ribeira Belt occurred during the interval 670 - 480 Ma.

The metamorphic grade increases along strike, from low- to medium-grade in the SE to granulite facies in the central and northeastern parts. High-grade gneiss belonging to both reworked basement and deformed cover succession, predominate in the central and northern parts of the belt. Low- to medium-grade siliciclastic and carbonate successions and scarce basement exposures appear in the Ribeira River region. Granitoid rocks are abundant along the entire belt and have been related to several tectonic settings, such as continental magmatic arc, syn-collisional, transpressional and late-collisional settings (Junho, 1993; Wiedemann, 1993; Campos Neto and Figueiredo, 1995; Heilbron *et al.*, 1995, 1998a, 2000; Ebert *et al.*, 1996; Töpfner, 1996; Wernick and Töpfner, 1997; Tupinambá *et al.*, 1998).

A particular characteristic of the belt is the presence of major subvertical deep crustal shear zones with dextral movement, that record an important transpressional component in the tectonic evolution of the belt. The most important of these zones is the Paraíba do Sul Shear Zone (Campanha and Ferrari, 1984; Chrispim and Tupinambá, 1989; Ebert *et al.*, 1991; Correia Neto *et al.*, 1993; Trouw, 1995; Ebert and Hasui, 1998; Heilbron *et al.*, 1998b), which is also called the Atlantic Shear Belt (Machado and Endo, 1993). The presence of these shear zones and the inflexion of the belt around the São Francisco Craton have inspired several authors to propose a model of lateral E-W convergence and escape tectonics related to a corner effect around the southern extremity of the São Francisco Craton (Vauchez *et al.*, 1992, 1994; Trompette *et al.*, 1993).

Although the geological understanding of the belt has increased considerably over the last decade, it is still fragmentary, mainly due to the fact that large areas have not yet been mapped in detail. Radiometric dating is also limited. The belt will be described according to regions that have been worked by different study groups. The central segment of the Ribeira Belt, including the interference zone with the Brasília Belt is described in the present section.

Tectonic framework and main discontinuities

The crustal structure of the Central Ribeira Belt is defined by two major tectono-stratigraphic terranes (in the sense of Howell, 1989): the Occidental and Oriental terranes (Heilbron *et al.*, 1995, 1998b, 2000). The Occidental Terrane comprises an autochthonous domain, the nappe system of the interference zone with the Brasília Belt, and the Juiz de Fora Thrust System. The Oriental Terrane encompasses the Paraíba do Sul Klippe, the Costeiro and the Cabo Frio domains (Fig. 2a). The contact between these two geotectonic units is a NW dipping shear zone that can be traced continuously along at least 200 km. This structure, called the Central Tectonic Boundary (CTB; Almeida *et al.*, 1998), was developed during the later stages of the Brasiliano Orogeny.

The autochthonous domain and the nappe systems in the interference zone (Fig. 2a) were considered to be an independent belt, the Alto Rio Grande Belt, defined by Hasui

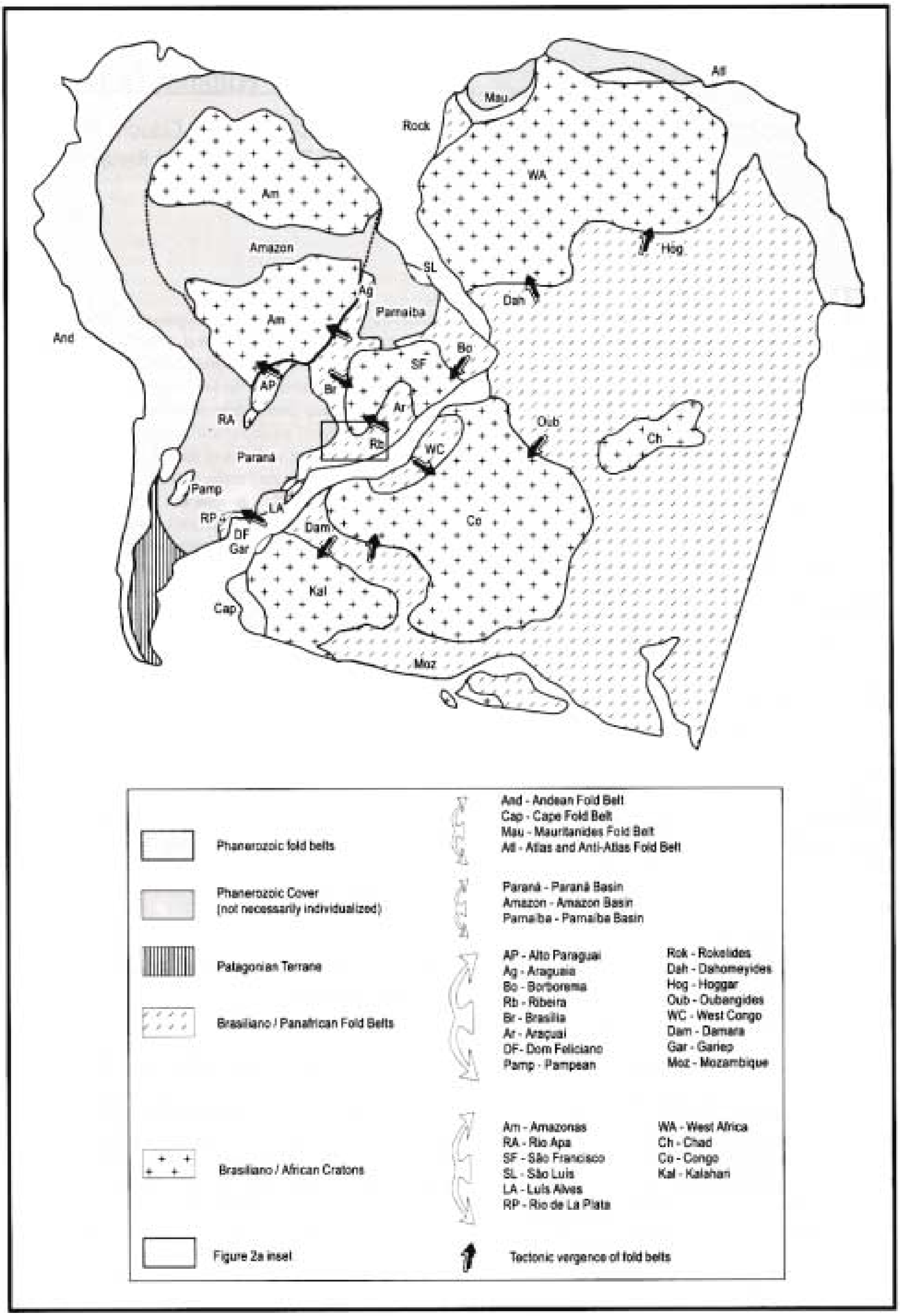


FIGURE 1 - The segment of the Ribeira Belt discussed in this paper in the context of the tectonics of western Gondwana (modified from Almeida and Hasui, 1984; Trompette, 1994).

and Oliveira (1984) as the regional extension of the São João del Rei and Andrelândia groups (Ebert, 1957, 1971). Detailed geological mapping and structural analysis, metamorphic petrology and sequence stratigraphy have improved understanding of the geological evolution of this region (Ribeiro *et al.*, 1995). Reconstruction of the basin geometry, the regional metamorphic pattern and analysis of the tectonic transport directions have led to a redefinition of the Alto Rio Grande Belt as the zone of interference between the Brasília and Ribeira belts (Trouw *et al.*, 1994). The interference zone includes basement units and Proterozoic supracrustal sequences of the São João del Rei, Carandaí and Andrelândia basins. The other tectonic domains also include basement units and Proterozoic cover successions. The Juiz de Fora system is regarded as a crustal scale duplex which resulted from the docking of the Costeiro Domain, during the Proterozoic-Paleozoic transition (Heilbron *et al.*, 1995, 1998b). The Paraíba do Sul Klippe is the uppermost thrust slice of the Central Ribeira Belt, and is composed of the Quirino Orthogneiss and the overlying Paraíba do Sul supracrustal succession. It occupies the hinge zone of the Paraíba do Sul Megasynform, a superposed structure associated with the subvertical dextral Paraíba do Sul Shear Zone (Fig. 2a, b). The Costeiro Domain, also called the Serra do Mar Microplate by Campos Neto and Figueiredo (1995), includes the Italva Basin and the Rio Negro Arc granitoid rocks (Tupinambá *et al.*, 1998; Almeida *et al.*, 1998). The Cabo Frio Domain (Fonseca *et al.*, 1984; Fonseca, 1998) overthrusts the Costeiro Domain. It includes the basement orthogneiss of the Região dos Lagos Complex and the Buzios Proterozoic succession.

Lithological associations

Four major lithological associations were recognised and mapped in the Central Ribeira Belt: reworked Paleoproterozoic/Archean basement, deformed Paleoproterozoic to Neoproterozoic cover successions, late orogenic fault-bounded sedimentary successions and Brasiliano granitoid rocks (Table 1). Tentative correlation between the geotectonic evolution of the Occidental and Oriental terranes are based on the reconstruction of the Proterozoic basins, and on the thermal and tectonic events during the Brasiliano Orogeny. Geochronological and geochemical data support the stratigraphic, metamorphic and structural correlation.

Basement (pre-1.8 Ga)

The basement units within the mobile belts are to a large extent similar to those in the adjacent São Francisco Craton. The limits of the craton are conventionally based on K/Ar metamorphic cooling ages > 1.8 Ga in the basement units. Outside the craton the basement occurs in structural highs and thrust slices.

Five main lithological associations are recognized in the basement of the central segment of the Ribeira Belt. In the Occidental Terrane, the basement units are the greenstone belt, the orthogneiss of the Mantiqueira Complex and the orthogranulite of the Juiz de Fora Complex, and Paleoproterozoic granitoid and gabbroic plutons. In the Oriental Terrane, the basement is essentially composed of the Quirino and Região dos Lagos orthogneiss complexes (Table 1).

Greenstone Belt

The greenstone belt, often referred to as the Barbacena Greenstone Belt, occur as narrow bands usually limited by intrusive Paleoproterozoic granitoid and gabbroic plutons (Fig. 3a). They contain typical ocean basin lithological associations consisting of mafic and ultramafic komatiitic and tholeiitic lavas (Noce *et al.*, 1987; Valença *et al.*, 1998), and metasediments. The latter are mainly grey metapelite with minor metawacke, metachert, manganiferous metachert and metadiamictite intercalations. Meta-andesitic and meta-rhyolitic subvolcanic bodies also occur, some of them intruding the metasediments (Ribeiro *et al.*, 1998; Ávila *et al.*, 1998a). At present it is not clear whether these shallow intrusive rocks are related to the Paleoproterozoic plutons, or to a different magmatic event.

The metamorphism is commonly of upper greenschist or lower amphibolite facies. The main deformation phase generally produced a steep foliation parallel to transposed bedding. However, in some areas beds are gently dipping and primary structures such as spinifex texture and sedimentary layering are preserved. Radiometric ages are not presently available, but the Barbacena Greenstone Belt, both within and outside the craton, has been correlated tentatively with the Rio das Velhas Supergroup, of Archean age, that crops out farther to the NE, in the Quadrilátero Ferrífero region of the São Francisco Craton.

Mantiqueira Complex

This domain has been called the Mantiqueira Series (Barbosa, 1954), Mantiqueira Complex (Brandalise, 1991) or Mantiqueira Metamorphic Complex (Figueiredo and Teixeira, 1996). The orthogneiss often with migmatite appearance is usually banded with granodioritic or tonalitic composition. It contains numerous tabular and lens-shaped bodies of rocks of amphibolitic and locally ultramafic composition. The metamorphic grade is mainly of amphibolite facies, but rocks of granulite facies also occur (Pinto *et al.*, 1992). Most Rb/Sr whole-rock isochron ages fall within the range of 2.3 to 2.0 Ga (Teixeira, 1985, 1993). However, U/Pb zircon ages of *c.* 3.4 and 3.1 Ga and a Rb/Sr whole rock isochron of *c.* 2.9 Ga (Cordani *et al.*, 1973; Söllner *et al.*, 1991) were also reported. According to Figueiredo and Teixeira (1996), chemical composition of the orthogneiss defines two different calc-alkaline suites, both suggesting a magmatic arc environment. According to Duarte (1998), the calc-alkaline rocks comprise four petrogenetic groups, each one generated by partial melting of crustal material. Basic intercalations with tholeiitic characteristics comprise a heterogeneous group generated by varied source materials as depicted by their REE patterns. Other basic intercalations with alkaline affinities display intraplate REE content characteristics, which also suggest a continental setting for the magmatism (Duarte, 1998).

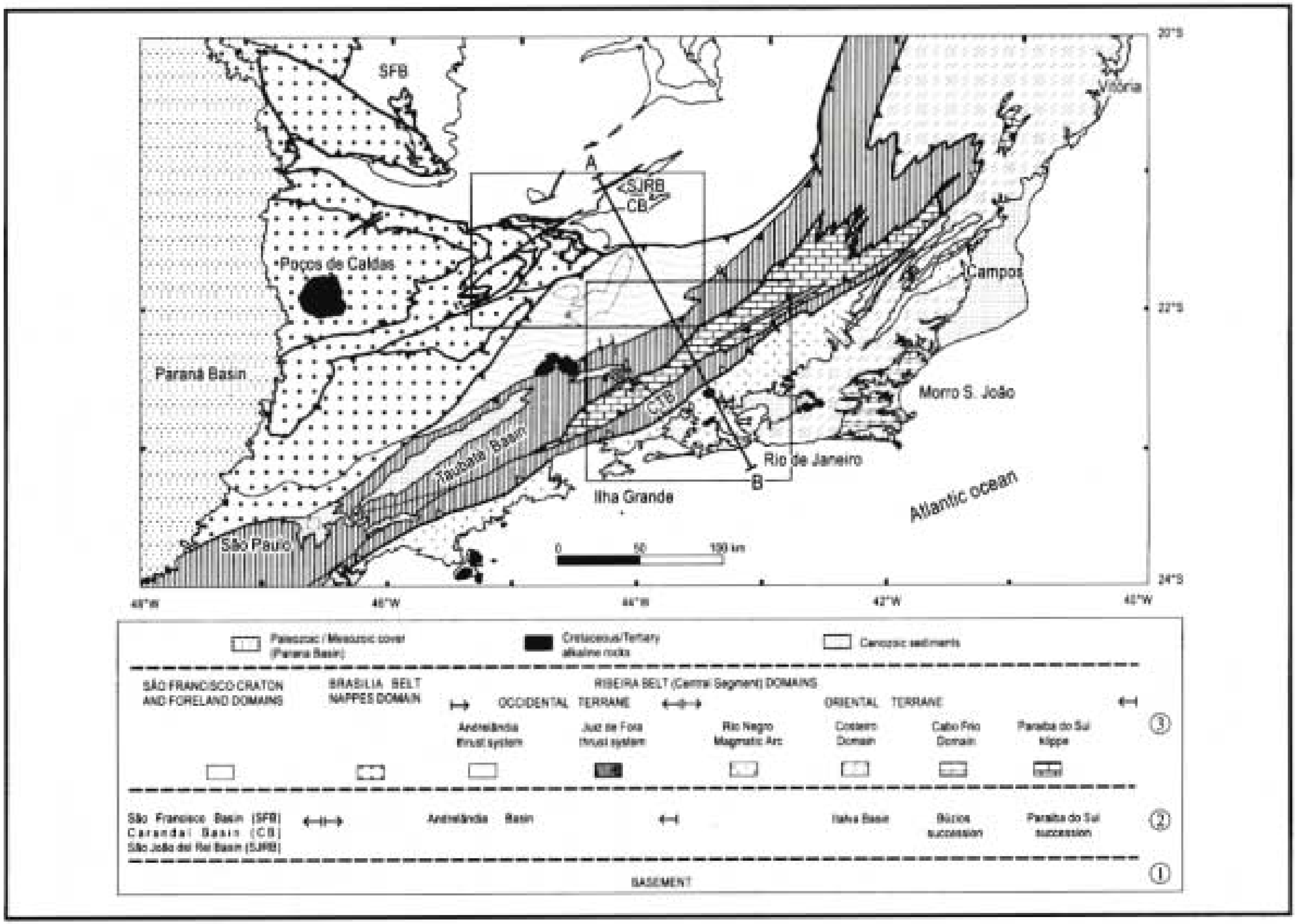


FIGURE 2a - Main tectono-stratigraphic units of the central segment of the Ribeira Belt. 1 - Basement units including paleoproterozoic granitoid rocks; 2 - Proterozoic basins and sucessions; 3 - Neoproterozoic tectono-stratigraphic units. The Andrelândia Thrust System includes the Brasília Belt-related Liberdade Nappe. The foreland of the São Francisco Craton includes the autochthonous domain. Section AB is displayed in Figure 2b. Insets are Figures 3a and 5a.

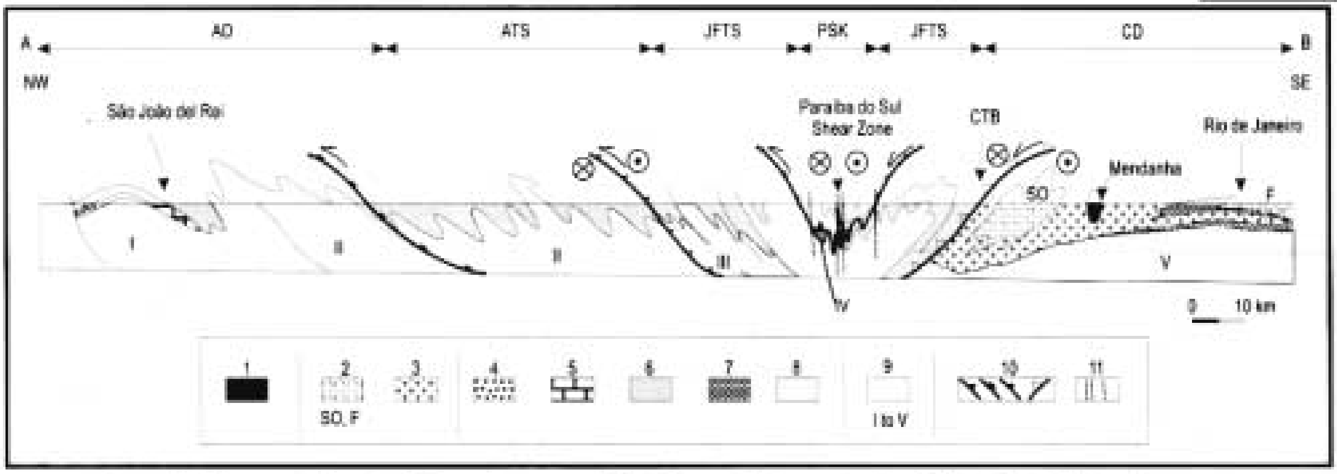


FIGURE 2b - Simplified structural section of the central segment of the Ribeira Belt between São João del Rei and Rio de Janeiro (modified from Heilbron, 1995; Heilbron et al.*, 2000). 1 - Mesozoic-Cenozoic alkaline rocks; 2 - Syn to late collisional granitoid rocks; 3 - Rio Negro Magmatic Arc granitoid rocks. Proterozoic basins and successions: 4 - São João del Rei; 5 - Carandaí; 6 - Andrelândia; 7 - Italva; 8 - Paraíba do Sul. 9 - Basement units: I - Barbacena Greenstone Belt (I) and Mantiqueira (II), Juiz de Fora (III), Quirino (IV) and Região dos Lagos (V) complexes; 10 - Major Thrusts; 11 - Subvertical shear zones. Tectonic domains: AD - Autochthonous domain; ATS - Andrelândia Thrust System; JFTS - Juiz de Fora Thrust System; PSK - Paraíba do Sul Klippe; CD - Costeiro Domain; CTB - Central Tectonic Boundary.*

Juiz de Fora Complex

The Juiz de Fora Complex contains predominantly Paleoproterozoic orthogneiss of granulite facies (Oliveira, 1982; Grossi Sad and Barbosa, 1985; Duarte *et al.*, 1997; Heilbron *et al.*, 1997, 1998a). Migmatitic structures predate peak metamorphic conditions (Sad and Dutra, 1988; Heilbron *et al.*, 1998a). The available isotope data indicate Paleoproterozoic ages (2.2 - 1.9 Ga) for both the protoliths of the orthogneiss and the granulite facies metamorphism (Delhal *et al.*, 1969; Oliveira 1982; Teixeira and Figueiredo, 1991; Machado *et al.*, 1996). Petrological and fluid inclusion studies point to high temperatures with low to intermediate pressure metamorphic conditions. CO_2-rich fluid inclusions (Nogueira *et al.*, 1996) and the lack of substantial LILE depletion may indicate that the CO_2-flushing model for metamorphism fits better than the model based on melting for the metamorphic evolution of orthogranulite (Duarte *et al.*, 1997; Heilbron *et al.*, 1998a). The geochemical data also support the hypothesis that the composition of the granulite protoliths was not substantially modified during metamorphic processes, especially regarding the REE. Geochemical data also suggest three different petrotectonic groups: calc-alkaline granulite of convergent tectonic setting, with pre-collisional arc-related rocks and syn-collisional granite; alkaline basic rocks typical of continental intraplate magmatism; and a heterogeneous tholeiitic group that suggests several tectonic settings such as back-arc basins, continental intraplate and E-MORB basalts (Heilbron *et al.*, 1998a).

Quirino Complex

The Quirino Complex comprises Paleoproterozoic granitic to granodioritic orthogneiss with basic and calc-silicate (tremolite-rich) enclaves that crop out at the Paraíba do Sul Klippe (Machado, 1984; Almeida *et al.*, 1993; Heilbron *et al.*, 1995; Valladares *et al.*, 1997). Ages of *c.* 2.2 Ga were obtained on zircon by the U/Pb method (Machado *et al.*, 1996, Valladares *et al.*, 1997). Geochemical data fall in two different groups: cordilleran magmatic arc and collisional granitoid plutons.

Região dos Lagos Complex

Paleoproterozoic (*c.* 2.0 Ga) granitic to granodioritic orthogneiss and intercalated metabasic bodies comprise the basement in the Cabo Frio Domain (Heilbron *et al.*, 1982; Zimbres *et al.*, 1990; Fonseca, 1994; Schmitt *et al.*, 1999). Sm/Nd model ages (T_{DM}) between 2.6 and 2.3 Ga (Fonseca, 1994) and $\varepsilon_{Nd(2.0\,Ga)}$ of *c.* - 6.0 suggests the existence of older continental crust in the source of the orthogneiss.

Transamazonian Orogeny

During the Paleoproterozoic Transamazonian Orogeny (2.2 - 1.9 Ga), the Mantiqueira orthogneiss and the greenstone belt rocks were reworked and intruded by felsic and mafic igneous bodies, most of them granitoid plutons probably related to the evolution of magmatic arcs (Figueiredo and Teixeira, 1996; Teixeira *et al.*, 1996; Duarte, 1998; Alkmin and Marshak, 1998; Ávila *et al*, 1998b). The protolith of the Juiz de Fora granulite was also related to arc magmatism (Heilbron *et al.*,1998a). The Paleoproterozoic metamorphic gradient increases from greenschist facies in the NW to granulite facies in the Juiz de Fora Complex.

The protolith of the Quirino and Região dos Lagos orthogneiss was probably related to late collisional magmatic events of the Transamazonian Orogeny. Isotope and geological field data suggest that the main metamorphic and tectonic events recorded by these orthogneiss result from the imprint of the Brasiliano orogenic stages (Zimbres *et al.*, 1990; Fonseca, 1994; Schmitt *et al.*, 1999). The development of the much younger and partially parallel Ribeira and Araçuaí belts may have been controlled at least partially by the older regional structure of the Paleoproterozoic belt.

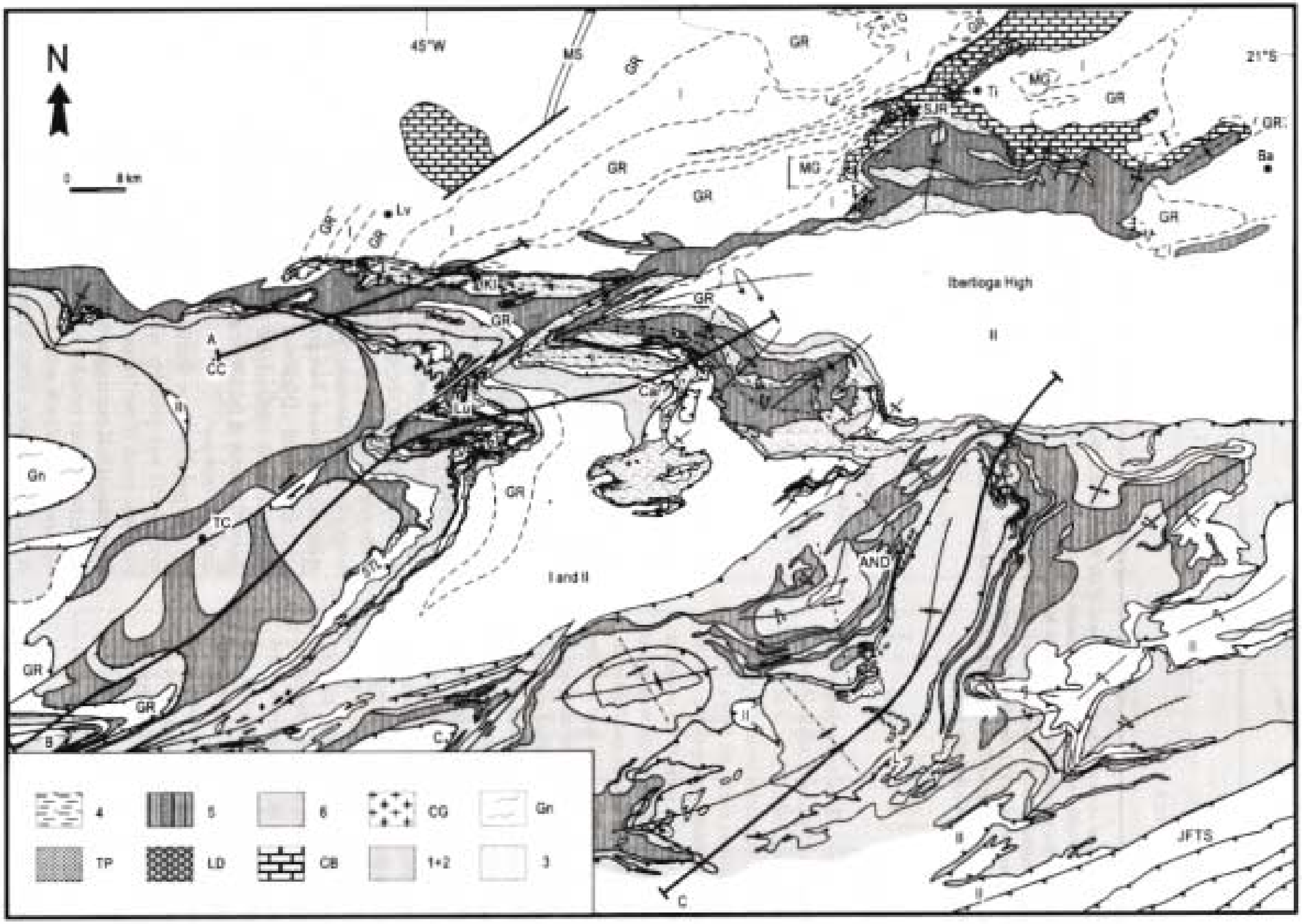


FIGURE 3a - Geological map of the interference zone between the Brasília and Ribeira belts, southern Minas Gerais State (modified after Paciullo, 1997). Basement units: greenstone belts (I), Mantiqueira Complex (II), Minas Supergroup (MS) and Paleoproterozoic intrusives (GR- granitoid and MG- gabroic rocks). Proterozoic units: São João del Rei Basin - quartzite and metapelite of the Tiradentes Platform (TP) and metapelite, quartzite and metaconglomerate of the Lenheiro Delta (LD), Metapelite and metalimestone of the Carandaí Basin (CB). Andrelândia Basin lithofacies associations: 1 and 2 - paragneiss, quartzite, schist and amphibolite; 3 - quartzite, 4 - phyllite/schist and minor quartzite intercalations; 5 - biotite schist/gneiss with local metadiamictite/dropstone, and 6 - biotite schist/gneiss and calcsilicate, metachert and amphibolite intercalations. CG - Syn to late-collisional granite. Gn - gneiss of the Guaxupé Nappe. JFTS - Juiz de Fora Thrust System. For structural symbology, see Figure 6. A, B, and C refer to cross-sections shown on Fig. 3b.

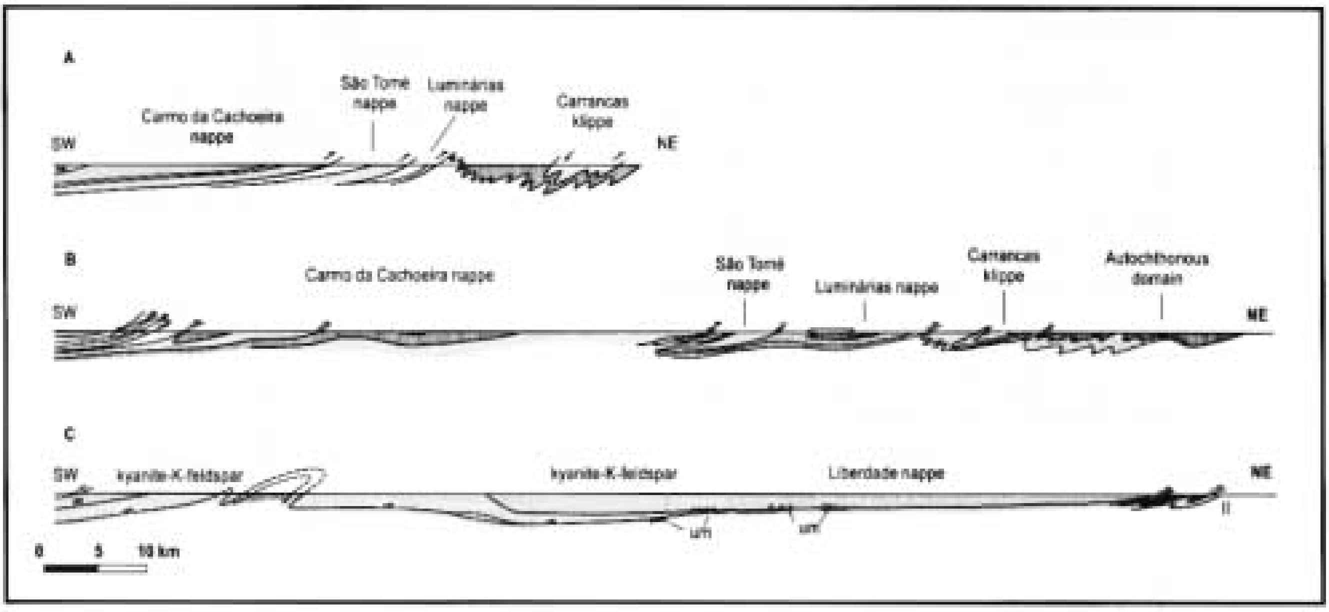


FIGURE 3b - Structural sections through Brasília Belt related nappes, indicated in Figure 3a. A and B: São João del Rei to Três Corações; section C: across the Liberdade Nappe. Um - ultramafic bodies, other simbologies as in Figure 3a.

Proterozoic basins and cover successions

The recognition of regional and intraformational unconformities and associated depositional sequences allowed the identification of Proterozoic basins and their evolution in the autochthonous domain and in the interference zone with the Brasília Belt. The high grade metamorphism, intense deformation and tectonic interleaving with basement units hamper basin analysis in the other domains. However, detailed mapping and geochronology allowed the discrimination of the Proterozoic metasedimentary successions from their basement both in the Juiz de Fora Thrust System and in the Oriental Terrane.

Autochthonous domain and interference zone with the Brasília Belt

These domains include the successions of the superposed São João del Rei, Carandaí and Andrelândia basins (Ribeiro *et al.*, 1995; Paciullo *et al.*, 1998). All three basins developed on a crystalline basement older than 1.8 Ga. The first two are intracratonic and of Paleoproterozoic to Mesoproterozoic age, whereas the Andrelândia Basin is a Neoproterozoic continental margin basin. The approximate geometry of these basins can be estimated from the distribution of their main lithofacies associations (Fig. 3a). The basin evolution, deduced from the deformed metasedimentary successions is outlined below. The time intervals indicated between brackets are only tentatively established, based on field relationships, few Sm/Nd T_{DM} model ages (Trouw *et al.*, 1997) and correlation with other Proterozoic basins (Ribeiro and Silva, 1996; Paciullo, 1997; Paciullo *et al.*, 1993; 1998).

São João del Rei Basin (1.8 - 1.3 Ga)

A quartzite sequence up to 1000 m thick rests unconformably on the basement and is unconformably overlain by the Carandaí Basin succession. Four depositional sequences are recognised based on intraformational unconformities (Fig. 4a). The lowermost quartzitic sequence records the transgression of a shallow sea over the wide Tiradentes Shelf. Foreshore to lower shoreface facies associations deposited under the influence of NE longshore currents and a NE-SW coastline record a pre-rift subsidence stage of the basin. Relative sea level falls, probably correlated with the Espinhaço rifting further N (Ribeiro and Silva, 1996), are responsible for the development of erosional surfaces that limit the two other preserved quartzitic sequences of the Tiradentes Shelf. These sequences are dominated by the stacking of middle shoreface deposits with tidal influence that grade upward, in the third sequence, to a quartzitic-pelitic lagoon-tidal flat facies association, with scarce stromatolitic facies. The fourth and uppermost sequence is an upward coarsening and thickening succession, up to 500 m thick, composed of massive and laminated metapelite beds, fine quartzite, pebbly quartzite and metaconglomerate. It resulted from an abrupt relative sea-level rise that created accommodation space for the progradation of a braided plain delta, the Lenheiro Delta (Ribeiro, 1997). The NE and NW oriented fluvial paleo-currents demonstrate the progressive appearance of an exposed area, the Ibertioga High (Fig. 3a) towards the S of the old shoreline (Paciullo *et al.*, 1993). This represents an important reversal in the southward dip of the local paleo-slope of the São João del Rei Basin towards the N. This episode marks the onset of the main rifting phase and culminated with the intrusion of mafic dikes, tilting of blocks, uplift and erosion.

Carandaí Basin (1.3 - 1.0 Ga)

After an erosional period of unknown duration, renewed rifting promoted the opening of the Carandaí Basin. The first stage in the evolution of this basin is represented by the development of the NE-SW Prados Graben (Ribeiro, 1997). This graben was filled with a thick succession containing hemi-pelagic deposits and pelitic turbidite beds (mainly metasiltite) in its deepest parts and diamictite units along the border. A shallow warm sea covered the graben deposits and the adjacent basement during the post-rift thermal subsidence. This led to the deposition of a thick succession including beds of limestone, marlstone and shale. The main facies associations of the Carandaí Basin are made up of both deep and shallow deposits. The tectonic evolution continued with an increase in area and elevation of the Ibertioga High. The renewed change in paleo-slope from N to S, accompanied by the development of sandy braided river systems that covered wide plains, marked the transition to the Andrelândia Basin.

Andrelândia Basin (1.0 Ga - 600 Ma)

The evolution of this basin was studied in detail by Paciullo *et al.*, (1993); Ribeiro *et al.*, (1995); Paciullo (1997). The Andrelândia Basin metasediments are at least 1500 m thick. The basal lithofacies association is composed of a paragneiss succession with amphibolite intercalations. It is interpreted to represent continental deposits related to the initial rifting stage (unit 1, Fig. 4b). These deposits are overlain by a succession of paragneiss and quartzite intercalations interpreted to representing sets of retrogradational parasequences (unit 2, Fig. 4b). The parasequences were produced by partial reworking of the underlying deposits during sea-level oscillations associated with a slow but persistent shallow marine transgression. This second unit grades upwards to a quartzitic platform succession, the third unit (unit 3, Fig. 4b). A sharp contact on top of these quartzite units separates them from a succession of grey metapelite beds (unit 4, Fig. 4b) that represent a rapid rise of the relative sea level. To the N, these pelite beds partially cover basement areas and also the Carandaí Basin successions. To the SW they grade to deeper lithofacies association (unit 6, Fig. 4b). Although pelite beds continued to be deposited in the deeper parts of the basin, thin-bedded turbidite and pelite were deposited in aprons developed near the platform border. They are now represented by biotite schist with scarce dropstone and debris flow intercalations containing clasts of basement rocks (unit 5, Fig. 4b).

Upwards, this succession is followed by transgressive deposits of pelite (top unit 5) that onlap the basement and older Andrelândia sediments. The turbidite and associated deposits record a lowstand period possibly related to the Riphean glaciation, while the overlying pelite beds represent a highstand period probably produced by glacio-eustatic sea-level rise. The sixth unit is composed of biotite schist and gneiss (including high-pressure granulites) with calc-silicate, manganiferous chert and amphibolite intercalations (unit 6, Fig. 4b). The sixth unit represents the continuous deposition on slope and basin floor areas, during the entire evolution of the Andrelândia Basin. The amphibolite units are interpreted to be derived from transitional to MORB-type basalt (Gonçalves and Figueiredo, 1992; Paciullo, 1997). They record Sm/Nd model ages (T_{DM}) of *c.* 1.2 and 1.05 Ga, and $\varepsilon_{Nd(1.0\,Ga)}$ values of + 4.8 and + 3.1, supporting their derivation from a depleted mantle reservoir and a short crustal residence time (Heilbron *et al.*, 1990).

Proterozoic cover in the Juiz de Fora Thrust System

The Proterozoic cover, of which the Juiz de Fora Complex was the basement, comprises a succession of paragneiss, quartzite and pelitic schist, that grades upward to a pelitic succession (mica schist/gneiss). These successions are correlated to the lithofacies associations 1, 2 and 6 of the Andrelândia Basin. The same cover succession, highly imbricated with the basement units, also appear in the Serra do Mar region (Heilbron *et al.*, 1991) as a consequence of fold-repetition at the southern limb of the Paraíba do Sul Megasynform (Figs. 2b and 5a).

Ages of detrital zircon from quartzite of the Andrelândia Basin, of both the Andrelândia and Juiz de Fora thrust systems, indicate a Paleoproterozoic to Archean source area (Machado and Gauthier, 1996; Söllner and Trouw, 1997; Valladares *et al.*, 1997).

Supracrustal rocks of the Paraíba do Sul Klippe

The metasedimentary succession in this klippe crops out in the core of the Paraíba do Sul Megasynform (Figs. 2b and 5a). It was included in the Paraíba do Sul Group by Rosier (1965) and Ebert (1971). The succession is composed of pelitic schist/gneiss and psammitic gneiss, both with dolomitic marble, calc-silicate and few quartzite intercalations. The quartzite beds are interpreted to be metachert. Almeida *et al.* (1993) subdivided the Paraíba do Sul Group in the southern part of State of Rio de Janeiro into three units: a basal unit containing psammitic gneiss with calc-silicatic, marble and pelitic (schist/gneiss) intercalations; an intermediate pelitic unit (schist/gneiss); and a top unit with psammitic gneiss and abundant pelitic (gneiss/schist), marble, calc-silicatic rock and manganesiferous meta-chert intercalations. Grossi Sad and Dutra (1998) also proposed a regional subdivision for the Paraíba do Sul Group into different formations. They proposed a back-arc tectonic setting in which sedimentation was associated with volcanism.

Inferences about the relative age of the Paraíba do Sul succession are hampered by the absence of geochronological data, by the tectonic character of contacts with the Andrelândia Basin successions and by the lack of defined stratigraphic relationships with the Paleoproterozoic basement (Quirino Complex).

Proterozoic cover of the Costeiro Domain

The supracrustal association of the Costeiro Domain comprises high-grade paragneiss with quartzite, calc-silicate and calcite marble intercalations (Figs. 2b and 5a). This association, together with the Paraíba do Sul Klippe supracrustal rocks, was named the Paraíba-Desengano Series (Rosier, 1965), Paraíba do Sul Series (Ebert, 1957) and Italva Group (Machado Filho *et al.*, 1983). It also has been included in the Paraíba do Sul Group by Ebert (1971), Grossi Sad and Dutra (1998), Tupinambá (1993a) and many other authors. However, recent field work indicates that this Proterozoic cover is an independent unit restricted to the Costeiro Domain (Heilbron *et al.*, 1998a, 2000). The

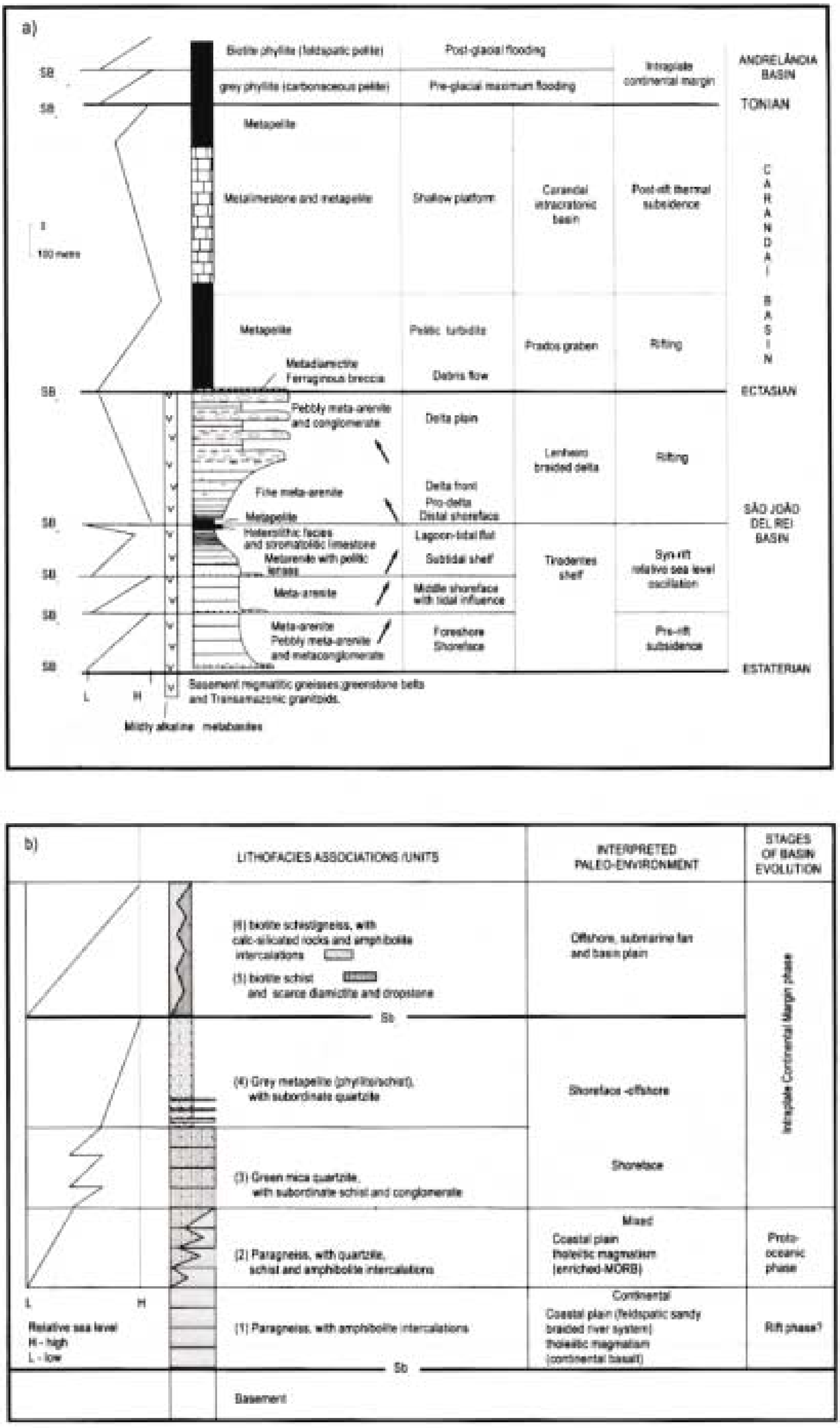
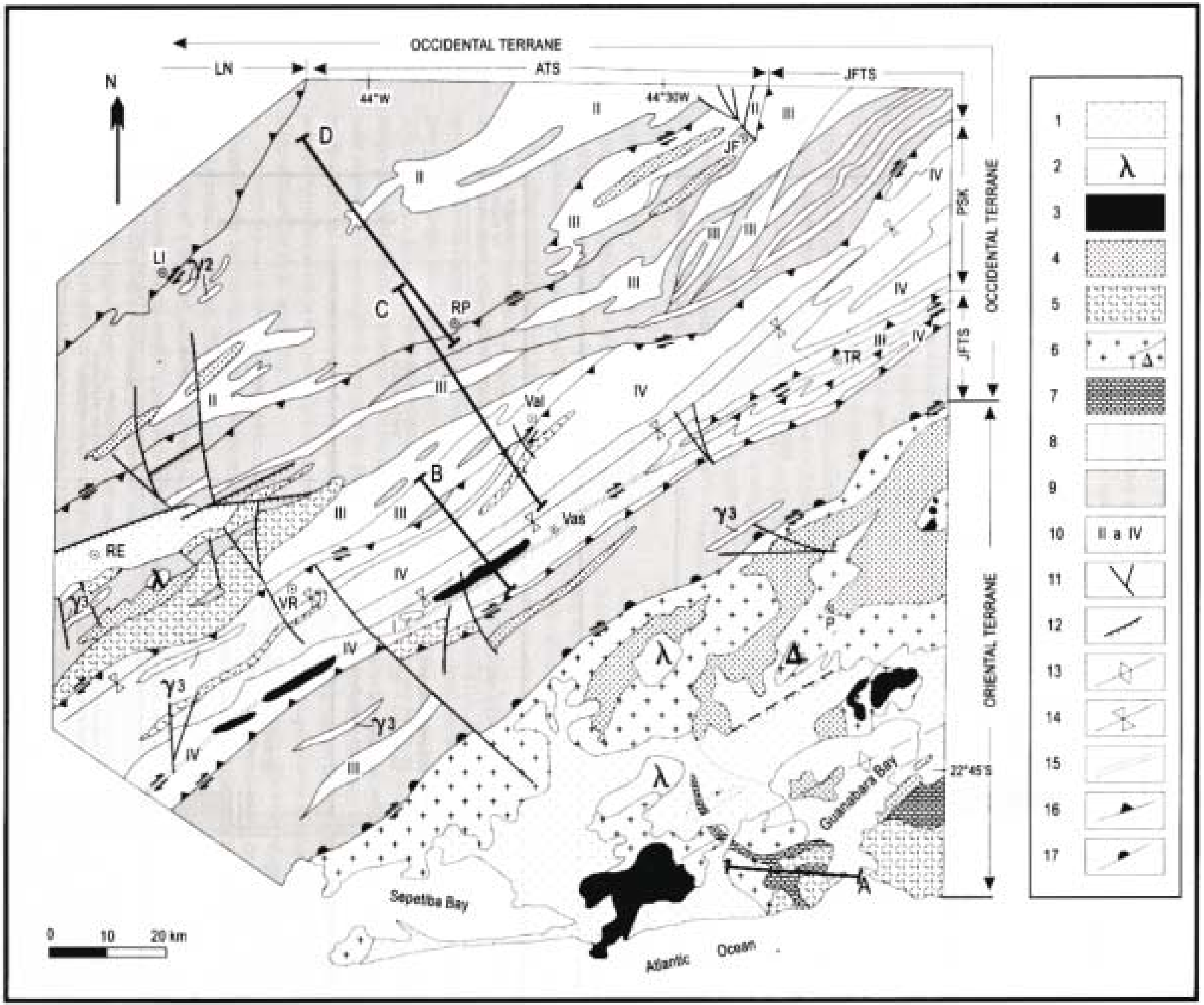


FIGURE 4 - Depositional sequences and lithofacies associations of the São João del Rei, Carandai (a) and Andrelândia basins (b). Sb1 - angular/lithological unconformity; Sb2 and Sb3 - disconformity; Sb4, Sb6 and Sb7 - correlative conformity; Sb5 - angular unconformity. Relative sea level H - high and L - low (modified from Paciullo, 1997; Ribeiro, 1997).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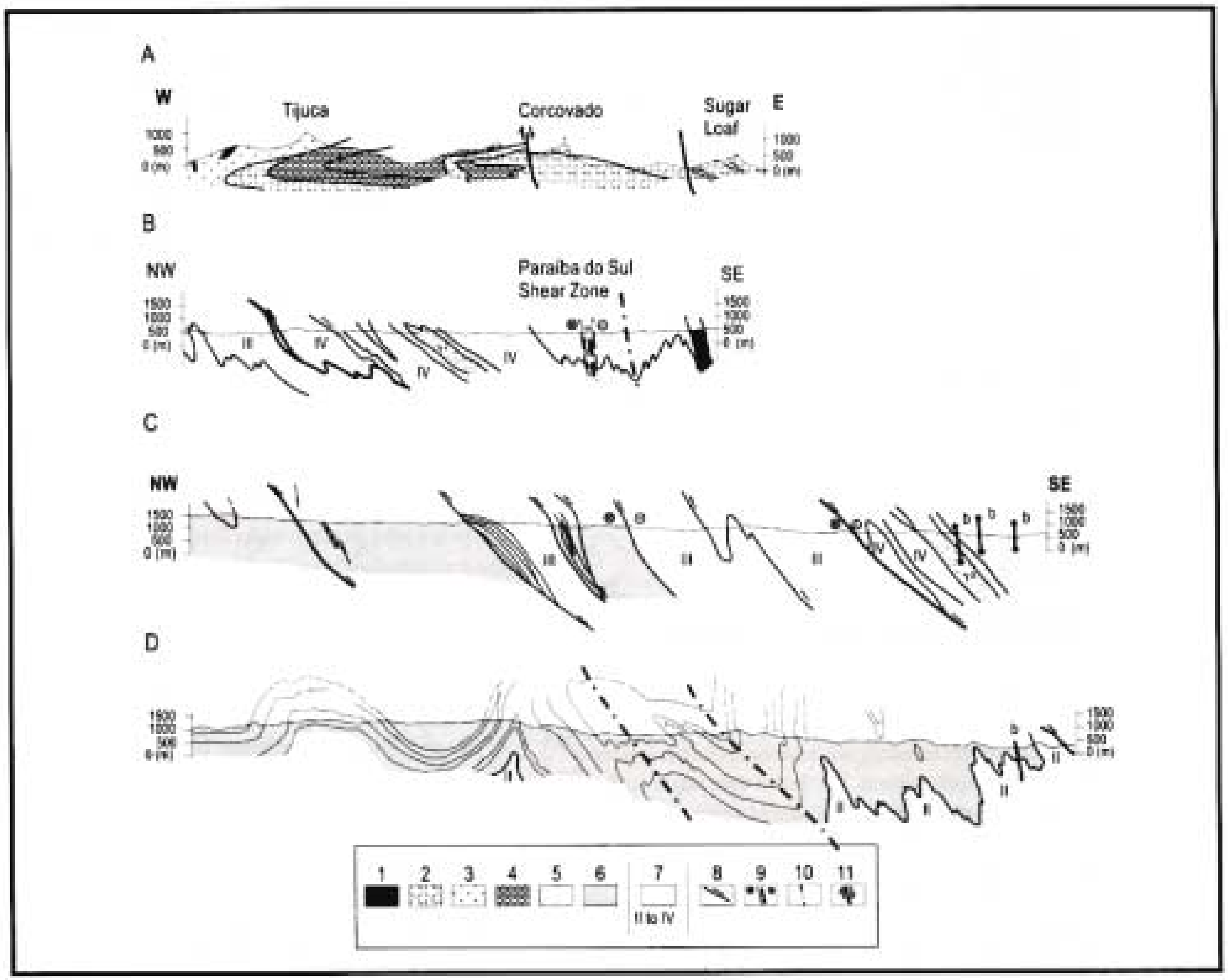


FIGURE 5b - Detailed cross sections showing different structural styles of the tectonic domains. A - the Costeiro Domain in City of Rio de Janeiro with large-scale recumbent folds without clear vergence. Normal faults are related to Meso-Cenozoic tectonism; B - the Paraíba do Sul Domain, with D_1/D_2 tight folds and thrusts refolded by the Paraíba do Sul Megasynform and associated shear zone (D_3), C - the Juiz de Fora Domain with thrust-imbrication toward the SFC and scarce folding; D - the Andrelândia Domain, with NW-vergent D_1/D_2 recumbent folds refolded by the vertical to inclined D_3 folds. Legend: 1-3 - Second Orogeny granitoid rocks (1 - Post-collisional and transitional; 2 - Syn-collisional; 3 - pre-collisional); 4-6 - Basin sucessions (4 - Italva, 5 - Andrelândia, 6 - Paraíba do Sul); 7 - pre- 1.8 Ga Basement associations: II - Mantiqueira, III - Juiz de Fora and IV - Quirino complexes; 8 - Major thrusts; 9 - transpressive shear zones; 10 - D_3 axial planes; 11 - Tertiary normal faults.

FIGURE 5a - Geological map of the central segment of the Ribeira Belt (modified from Heilbron, 1995; Heilbron et al., 1993, 1995, 2000). 1 - Cenozoic cover; 2 - Mesozoic-Cenozoic alkaline rocks; Second orogeny granitoids rocks: 3 - Late-collisional and slab detachment related granitoids; 4 - Late-collisional; 5 - Syn-collisional; 6 - Magmatic arc rocks including alkaline rocks (D). Basin sucessions: 7 - Italva; 8 - Paraíba do Sul and 9 - Andrelândia. 10 - Pre-1.8 Ga basement associations: II - Mantiqueira, III - Juiz de Fora and IV - Quirino complexes. Tertiary faults: 11 - transfer and 12 - normal faults; 13 - Rio de Janeiro Mega-antiform; 14 - Paraíba do Sul Mega-synform; 15 - Paraíba do Sul Shear Zone; 16 - Major thrusts; 17 - Central Tectonic Boundary. A, B,C, and D are the cross-sections of Fig. 5b.

succession was interpreted to be deposited in the Italva Passive Margin Basin (Heilbron *et al.*, 2000). Three main lithofacies associations are recognised in the Italva Basin: a) banded biotite paragneiss with abundant quartzitic and minor mafic and calc-silicate intercalations; b) thick strata of calcite-marble; and c) metapelitic gneiss (kinzigites) with minor calc-silicate, quartzite, and calcite-marble intercalations.

Proterozoic cover of the Cabo Frio Domain

The supracrustal cover of the Cabo Frio Domain comprises kyanite/sillimanite pelitic gneiss with calc-silicate, amphibolite and garnet quartzite intercalations (Fig. 2a). The succession was named the Superior Sequence (Heilbron *et al.*, 1982) and Buzios Group (Machado *et al.*, 1983), and represents the sedimentation in the Buzios Basin.

Brasiliano orogenic stages

Three Brasiliano orogenic stages were recognised in the Central Ribeira Belt. These orogenic stages are partially superimposed in time and space, resulting in complex structural and metamorphic patterns. The first of these stages occurred between 670 and 600 Ma and culminated with the E-W closure of the Brazilides Ocean (Unrug, 1997), resulting in the N-S trending Brasília Belt. The second orogenic stage took place between 630 and 520 Ma during the closure the Ribeira branch of the Adamastor Ocean (Unrug, 1997). The third orogenic stage, between 520 - 480 Ma, was responsible for the final closure of the Buzios Basin. This succession of events has been inferred from analysis of the superimposed structures and their corresponding tectonic transport directions, and the pattern of metamorphic isograds. The time interval of this orogenic stage can be determined from radiometric ages of the metamorphic pulses and of the pre, syn, and late-tectonic granitoid rocks (Figueiredo and Campos Neto, 1993; Heilbron, 1995; Campos Neto and Figueiredo, 1995; Machado *et al.*, 1996; Ebert *et al.*, 1996; Tupinambá *et al.*, 1998; Campos Neto and Caby, in press). Campos Neto and Figueiredo (1995) designated the first orogenic stage as Brasiliano I, and part of the second together with the third stage as the Rio Doce Orogeny.

First orogenic stage

This orogenic stage is related to the southernmost evolution of the Brasília Belt. Radiometric ages of this orogenic stage fall within the range 700 - 600 Ma (Figueiredo and Campos Neto, 1993; Janasi *et al.*, 1993; Ebert *et al.*, 1996; Töpfner, 1996; Wernick and Töpfner, 1997; Campos Neto and Caby, in press). The tectonic evolution began with westward subduction of the São Francisco Plate producing magmatic arc-type granitoid plutonism in the upper plate, locally represented by the Guaxupé Nappe. Subsequent collision between the São Francisco and another probable major continental block presently covered by the Paraná Basin (Guimarães, 1951; Trompette *et al.*, 1993) resulted in a spectacular nappe system with a strongly indented outcrop pattern (Figs. 3a and 6) (Ribeiro *et al.*, 1995; Paciullo, 1997; Campos Neto and Caby, in press). Within the nappe system, there occur three synformal nappes, bounded by lateral ramps with wedge-shaped fronts. From N to S these are the Passos Nappe System with top-to-ESE tectonic transport (Valeriano *et al.*, 1995; Simões, 1995; Valeriano *et al.*, 1998), the Luminarias Nappe System, with tectonic transport to the E and the Socorro-Liberdade Nappe System with ENE-tectonic transport (Trouw *et al.*, 1982; Almeida, 1992; Ribeiro *et al.*, 1995; Ebert and Hasui, 1998; Campos Neto and Caby, in press). In the interference zone of the Brasília and Ribeira belts the Luminarias and the Socorro-Liberdade nappe systems are the units with best-preserved deformational fabrics and metamorphic pattern related to the Brasília stage.

The Luminarias Nappe System includes, in ascending order, the Carrancas Klippe and the Luminarias, São Tomé, Carmo da Cachoeira and Varginha nappes. The Liberdade Nappe is a relict structure of the first orogenic stage within the Andrelândia Thrust System (Fig. 6). The Carrancas Klippe is composed of shelf quartzite beds and pelite of the Andrelândia Basin (units 3 and 4; Fig. 4b) whereas the Luminarias and São Tomé nappes are composed of units 1 to 5 (Fig. 4b). The other nappes are largely composed of distal shelf and deep basin facies associations (units 5 and 6; Fig. 4b). The Varginha Nappe is composed of mainly igneous complexes with high metamorphic grade, partly of granulite facies. Basement slices occur along several of the thrust surfaces.

The eastwards tectonic transport in the Luminarias Nappe System (Fig. 6) approximately coincided with the metamorphic peak temperatures. Ductile deformation has transformed all rock types into strong S-L tectonites but mylonites are rare due to recrystallisation and crystal growth. The nappes can be classified as fold related. Tight to isoclinal folds, ranging in size from centimetric to kilometric scales, are usually oriented with their axes parallel to the E-W stretching lineation and exhibit vergence both to the N and S. With reference to the general kinematic framework they may be interpreted as sheath folds, as oblique folds, or as being due to the lateral confinement of the thrust sheets between oblique lateral ramps (Trouw *et al.*, 1982; Almeida, 1992). Interference patterns, including refolding and crenulation cleavage permit the identification of two coaxial and progressive deformation phases, D_{B1} and D_{B2} (Trouw *et al.*, 1982; Ribeiro *et al.*, 1995).

The metamorphism of the first orogenic stage (M_B) varies from middle greenschist facies in the Autochthonous Domain to upper amphibolite and granulite facies in the upper nappes (Fig. 7).

The metamorphic grade increases towards the upper nappes both as gradational inverted metamorphism within single nappe units (Simões, 1995; Simões *et al.*, 1994; Campos Neto and Caby, in press) and as post-metamorphic thrusts of higher-grade nappes over lower grade ones. Relatively high-pressure metamorphism is recognised in the upper nappes systems and is related to the E-W collision. It is characterised by a kyanite-K-feldspar association in metasediments and by the clinopyroxene-garnet-quartz association in mafic rocks (Ribeiro *et al.* 1995; Trouw and Castro, 1995; Campos Neto and Caby, in press). It is also identified locally in the Andrelândia Thrust System,

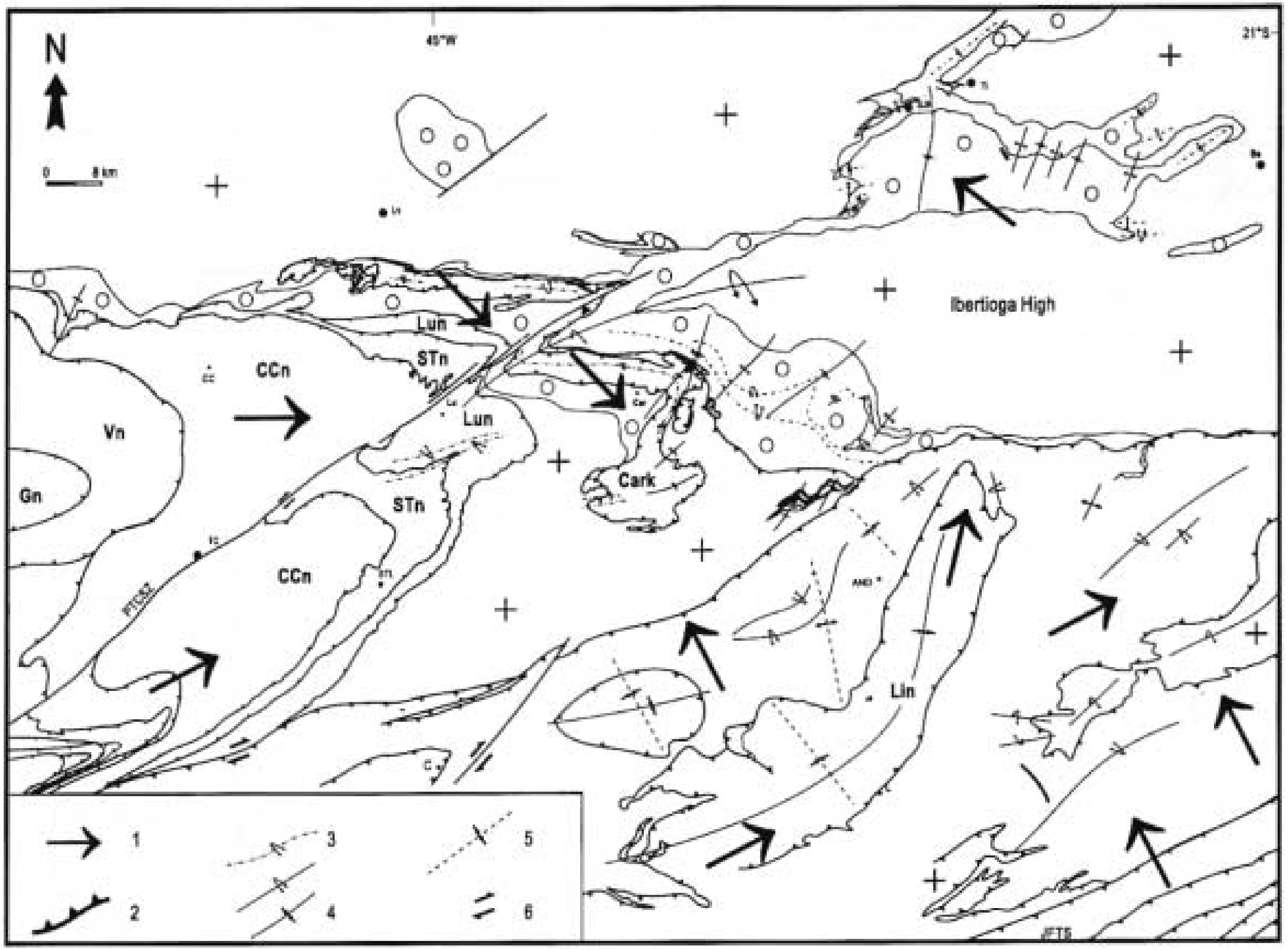


FIGURE 6 - Simplified structural map of the interference zone between the Brasília and Ribeira belts, southern Minas Gerais State (modified after Paciullo, 1997). 1 - Main tectonic transport direction. 2 - Trace of major thrust surfaces. Trace of axial surfaces of: 3 - D_{B1} and D_{B2} folds; 4 - D_{B3} folds and 5 - D_{B4} folds. 6 - Major subvertical shear zones: PTCSZ - Pombeiro-Três Corações shear zone. Cark - Carrancas Klippe. Nappes: Lun - Luminarias, STn - São Tomé, CCn - Carmo da Cachoeira, Vn - Varginha; Gn - Guaxupé; Lin - Liberdade. JFTS - Juiz de Fora Thrust System. Autochthonous domain: basement units (cross) and Proterozoic depositional sequences (circle).

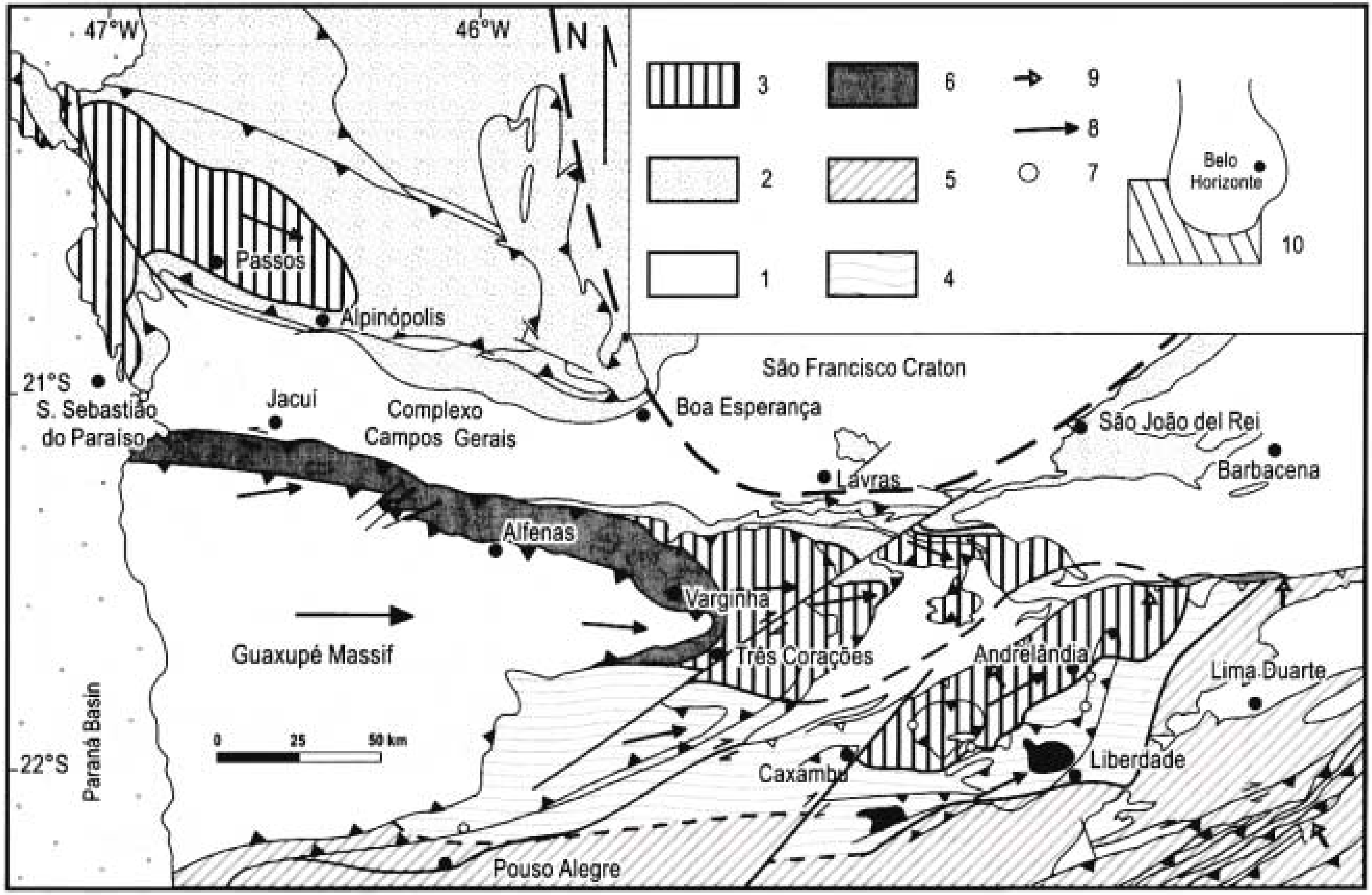


FIGURE 7 - Simplified metamorphic map. 1 - Pre 1.8 Ga basement; 2 - greenschist facies; 3 - kyanite; 4 - kyanite and sillimanite; 5 - sillimanite, in anphibolite facies; 6 - high pressure granulites (kyanite + k-feldspar); 7 - retro-eclogite. Direction of tectonic transport related to 8 - Brasília Belt and 9 - Ribeira Belt. 10 - Location with relation to São Francisco Craton (modified from Ribeiro et al., 1995).

although there it was strongly overprinted by metamorphism related to the Ribeira stage. Peaks metamorphic conditions determined by geothermobarometry, based on data from mafic granulite, are 800° - 900°C and 12-13,5 kbar (Ribeiro *et al.*, 1995). Similar metamorphic conditions are reported within the Varginha Nappe (Del Lama *et al.*, 1993; Campos Neto and Caby, in press).

The presence of proximal shelf facies successions in the lower nappe units (Luminarias, São Tomé) and of distal facies, high pressure granulite and associated retroeclogite (Trouw *et al.*, 1998) in the upper nappes (Liberdade, Carmo da Cachoeira, Varginha) are consistent with continental collision. The distribution of the granulite and retroeclogite along main thrust surfaces outlines the collisional suture (Fig. 7). A similar suture was proposed previously based on gravity anomaly patterns (Lesquer *et al.*, 1981; Malagutti Filho *et al.*, 1996).

Second orogenic stage

This stage is related to the evolution of the Central Ribeira Belt. It initiated with southeastward subduction of the southern margin of the São Francisco Plate beneath the Serra do Mar Microplate, producing magmatic arc-type granitoid plutonism in the upper plate. Based on the tectono-metamorphic evolution, magmatic pulses and U/Pb data, the second orogenic stage is subdivided into five episodes (Fig. 8). A pre-collisional episode took place within the interval 630 - 595 Ma range, based on radiometric ages of the Rio Negro arc-granitoid rocks (Tupinambá *et al.*, 1998). Subsequent oblique collision between these two plates resulted in the closure of the Ribeira branch of the Adamastor Ocean (Fig. 9). This collision produced a complex structural pattern including NW-vergent nappe systems, inverted metamorphic gradients (M_{R1}) and abundant S-type granitoid plutons (Heilbron *et al.*, 1995). Radiometric ages of the early syn-collisional metamorphism fall within the range 595 - 565 Ma (Machado *et al.*, 1996). The late-collisional episode (540 - 520 Ma) produced subvertical dextral ductile shear zones. The superposition of the second orogenic stage on the older southern Brasília Belt disturbed previous structures and metamorphic isograd patterns, resulting in the complex interference zone between the two belts (Fig. 7).

Deformation

Occidental Terrane and the interference zone with the Brasilia Belt

A similar metamorphic and structural evolution is recorded in each tectonic domain of the belt. Different structural styles and metamorphic facies are a consequence of deformation at specific crustal levels before thrust stacking (Fig. 5b). Three main deformational pulses are defined based on field structural data and on their time-relationships with metamorphism and magmatism.

The first pulse resulted in the main deformation phase (D_{R1}+D_{R2}) and was coeval with the M_{R1} metamorphic stage and consistent with an oblique collision model. The tectonic transport to NNW indicates vergence towards the São Francisco Craton. The development of penetrative structures and fabrics (*e.g.*, ductile thrusts, folds, main foliation, stretching and mineral lineations) are a consequence of thrust stacking of the different tectonic domains during this first pulse (Heilbron, 1995; Ribeiro *et al.*, 1995; Heilbron *et al.*, 1998b, 2000).

In the second deformational pulse a superimposed compressive phase (D_{R3}) generated subvertical folds and N trending transpressive shear zones, coeval with the M_{R2} metamorphic stage. Outstanding D_{R3} map-scale structures are the regional asymmetric open refolding of the older Brasília belt nappe systems (Fig. 6) and of the autochthonous domain (Valeriano, 1985, 1986; Ribeiro *et al.*, 1995); the Paraíba do Sul Megasynform, the Rio de Janeiro Mega-antiform (Heilbron, 1995) and the Paraíba do Sul transpressive shear zone (Figs. 2b and 5a).

The third pulse produced open to tight crenulation and folds, with N-S axes and subvertical axial planes, indicative of E-W compression (D_{R4}). Two sets of steep to vertical shear zones are attributed to this phase. The dextral NE-SW set was well developed in the autocthonous domain and in the Brasília belt nappe systems (*e.g.*, Pombeiro - Tres Corações Shear Zone; Fig. 6). Discrete sinistral NW-SE shear zones recorded in all tectonic domains represent the other set. Both sets of D_{R3} and D_{R4} shear zones acted as conducts for the ascent of late-collisional (Valladares *et al.*, 1995; Ebert *et al.*, 1995) to slab-detachment related granite.

Costeiro Domain

Large-scale recumbent isoclinal folding is the most conspicuous internal structure in this domain. The regional pattern of the recumbent folding is symmetric, with highly deformed hinge zones and no evident vergence (Heilbron *et al.*, 1993, 2000). Mylonitic foliation is restricted to the Central Tectonic Boundary. Few shear zones within the Rio Negro granitoid rocks indicate top-to-NW tectonic transport. In addition, the two sets of D_{R4} structures overprint this tectonic fabric (Fig. 5b).

Metamorphism

The second orogenic stage shows two metamorphic events, M_{R1} and M_{R2}. The M_{R1} event is of Neoproterozoic age (595 - 565 Ma), and the M_{R2} is Early Paleozoic (540 - 520 Ma) in age (Fig. 8). The M_{R1} event produced intermediate-pressure mineral parageneses that overlap the M_B high-pressure metamorphism in the Andrelândia Thrust System. Microstructural observations indicate that the main foliation includes the M_{R1} assemblages. Peak metamorphic temperature increases eastwards, from the Occidental to the Oriental terranes, displaying successive metamorphic zones: biotite, garnet, kyanite-sillimanite, sillimanite and K-feldspar (Heilbron, 1985, 1993, 1995; Trouw *et al.*, 1986; Tupinambá, 1993b). Orthopyroxene occurs locally within the Proterozoic successions of the Juiz de Fora Thrust System (Heilbron, 1993; Nogueira *et al.*, 1996; Duarte, 1998). The spatial distribution of the M_{R1} metamorphic zones, with high-grade metamorphic zones thrusted over the lower grade zones, suggests an inverted metamorphic gradient as in the southern portion of the Brasília Belt. Late M_{R1} metamorphic conditions around 700 -750°C and 6 - 7 kbar were reported by Duarte (1998) in the Juiz de Fora Thrust

System. An isothermal decompression P-T-t path was suggested for the metamorphic evolution in this domain.

At the end of the collision the M_{R1} event reached higher temperatures recorded by migmatization and S-type granitoid plutons. Cordierite starts to grow at this time, both in anatectic veins of the Occidental Terrane and as a widespread late-tectonic mineral in the pelitic gneiss of the Costeiro Domain (Pires and Heilbron, 1986).

The M_{R2} metamorphism shows a syn-to-late relationship with D_{R3} deformation. It is represented by retrogressive metamorphic minerals in both the Occidental and Oriental terranes.

Magmatism

The general characteristics of the magmatism and its relationship with the tectonic episodes of the second orogenic stage are summarised in Table 2 and Figure 8, respectively. Arc-related magmatism is represented by the Rio Negro granitoid rocks and occurs only in the Costeiro Domain (Tupinambá *et al.*, 1998; Heilbron *et al.*, 1998b; 200. These granitoid bodies occupy approximately half of the area of the Costeiro Domain, extending for almost 600 km along the Atlantic coast, from northern São Paulo State to southern Espírito Santo State (Fig. 2a). Syn to late-collisional granitoid rocks occur both in the Occidental and Oriental terranes, showing a spatial and temporal polarity within the belt. They are more abundant in the Juiz de Fora Thrust System and towards the coast, indicating crustal thickening as the result of continental collision. The late-collisional period shows metaluminous to slightly peraluminous leucogranite related to the D_{R3} subvertical shear zones. Elongate syn-D_{R3} batholiths and stocks of leucogranite are widespread in the Paraíba do Sul Klippe and in the Oriental Terrane.

Small plutons and stocks of high-K calc-alkaline to alkali-calcic type intrude all lithological units of the Costeiro Domain. They have not been reported within any other terranes of the central Ribeira Belt. Textural patterns and structural relationships with country rocks suggest post-tectonic emplacement.

The third orogenic stage

This stage was recently recognised within the Cabo Frio Domain based on U/Pb ages of *c.* 520 Ma for the metamorphism and the main deformation of the Buzios succession and its basement (Schmitt *et al.*, 1999). This is the last stage of the Brasiliano Orogeny recorded in the Central Ribeira Belt.

The Cabo Frio Domain overthrusts the Costeiro Domain (Fonseca, 1998). The main structural characteristics are a gently dipping coarse schistosity found in both Buzios succession and basement orthogneiss (D_1; Heilbron *et al.*, 1982). The principal sense of tectonic transport within this foliation is top-to-NNW (Schmitt *et al.*, 1997). This foliation is refolded by two phases of recumbent tight to isoclinal folds with NNW trending axes (D_2 and D_3; Heilbron *et al.*, 1982), that also deform the basement-cover contact on a regional scale. Two sets of orthogonal subvertical folding related to shear zones are the latest deformational event recorded in this domain. This phase could be possibly related to the E-W regional compression that produced the D_{R4} deformation throughout the Ribeira and Brasília belts.

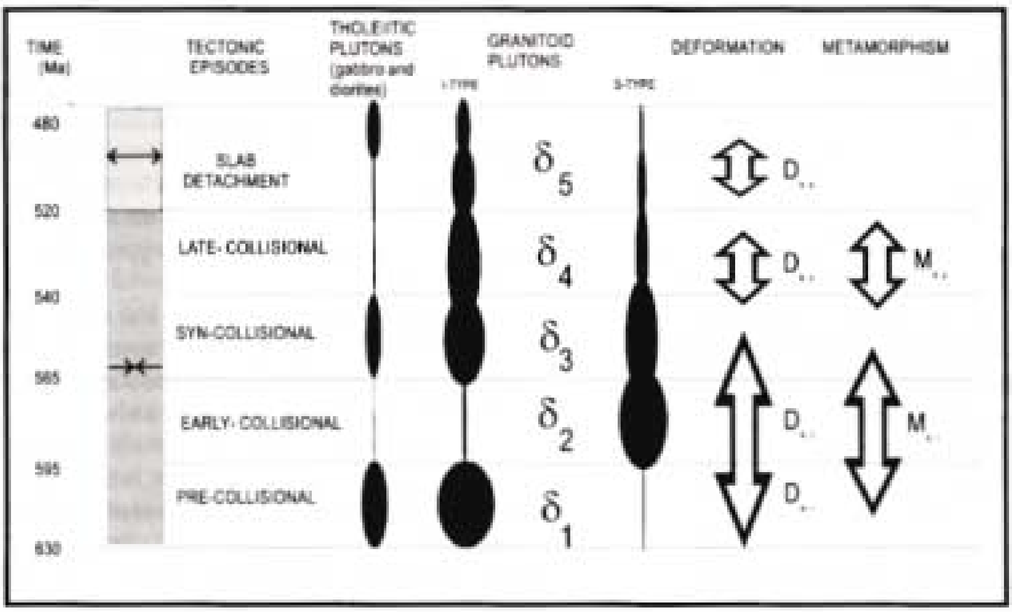
The metamorphism of the third orogenic stage is well recorded within the Búzios succession. It reached temperatures from the upper amphibolite facies to the transition with the granulite facies. It also displays higher metamorphic pressures when compared with the Costeiro Domain. The relatively high-pressure parageneses were developed within the main foliation (D_1), and includes kyanite, garnet, sillimanite and K-feldspar (Heilbron *et al.*, 1982).

Final Remarks

An evolutionary tectonic model for the central segment of the belt presented below is based on the observed relationships between structural evolution and metamorphic/magmatic features, also supported by geochronological and geochemical data.

At the end of the Transamazonian Orogeny (*c.* 1.9 Ga) the docking of terranes such as microplates and magmatic arcs, generated a major continent comprising the present southeastern Brazil. A variety of geological and geophysical evidence indicates the existence of large continental masses during the Orosirian period of the Paleoproterozoic (Van Schmus *et al.*, 1993; Brito Neves *et al.*, 1995; Rogers, 1996). Pieces of the Transamazonian amalgamated paleocontinent can partly be restored from the pre-1.8 Ga basement units in the Ribeira Belt. The basement of the Occidental Terrane is related to the evolution of the São Francisco Craton. Archean gneiss of the Mantiqueira Complex and supracrustal successions of the Barbacena Greenstone Belt are the oldest units in this terrane. The oceanic arc-derived orthogneiss of the Juiz de Fora Complex was accreted to these older units during the Transamazonian Orogeny (2.2 - 1.9 Ga). The pre- to syn-collisional orthogneiss of the Quirino and Região dos Lagos complexes that constitute the pre-1.8 Ga basement of the Oriental Terrane may represent the southeastern part of the continental crust (Serra do Mar Microplate) between the São Francisco and Congo cratons during the Paleoproterozoic (Fig. 9). Alternatively, the orthogneiss could belong to the western side of the Congo Plate.

The Staterian period is manifest by widespread taphrogenesis in the São Francisco-Congo paleocontinent. Important rift systems developed at this time, *e.g.*, the Espinhaço and São João del Rei rifts (Brito Neves *et al.*, 1995). Felsic volcanism around 1.8 - 1.75 Ga is related to the initial rift stage. The resulting NE-SW regional structure controlled the orientation of the Mesoproterozoic intracontinental rifts that initiated the evolution of both São João del Rei and Carandaí basins. Two major evolutionary stages were recognised in these basins. In the São João del Rei Basin, the Tiradentes Shelf records the pre to early rifting stage, whereas the Lenheiro Delta marks the onset of rifting. Paleoproterozoic U/Pb zircon ages of basement granitoid rocks and Mesoproterozoic Sm/Nd model ages of mafic dykes constrain the deposition in this basin to the interval 1.8 - 1.3 Ga.



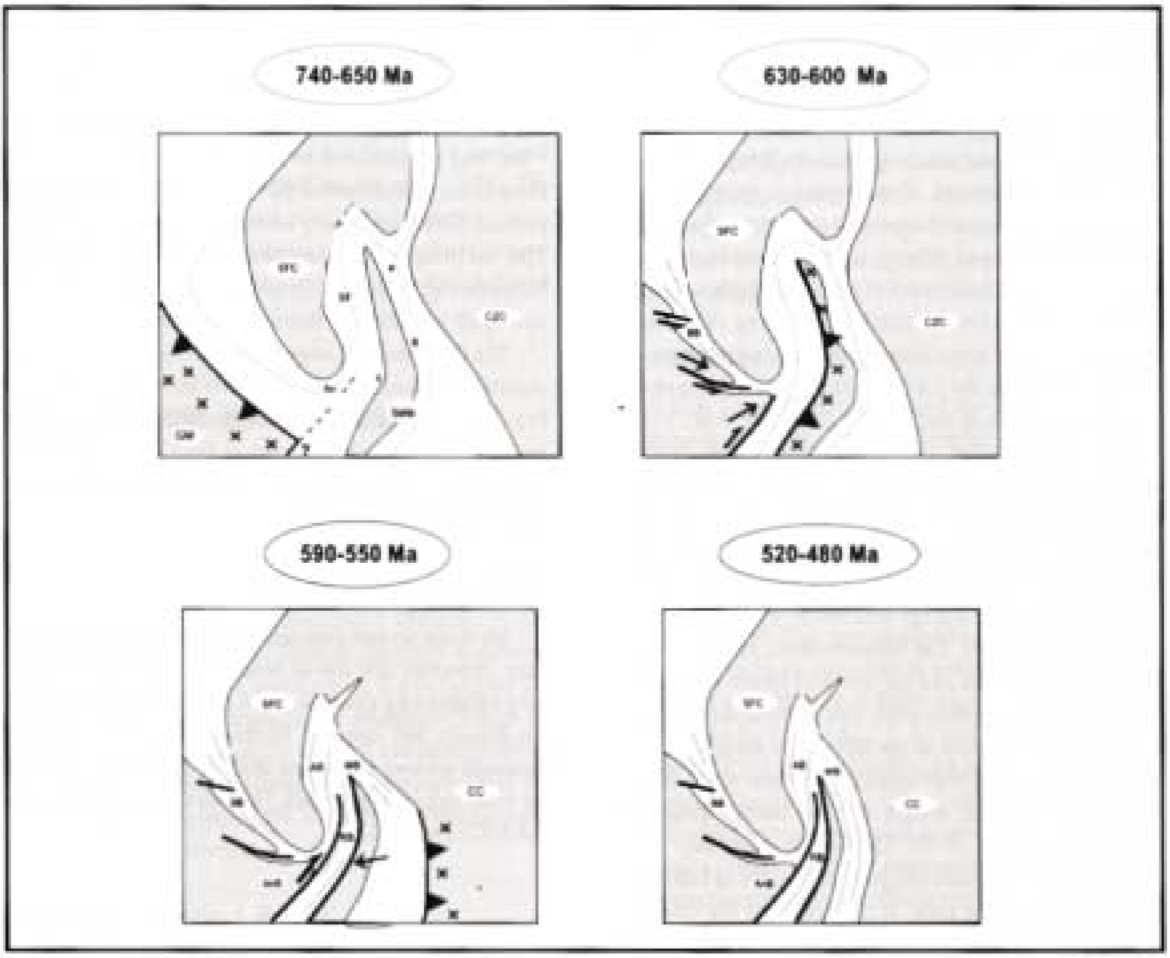


FIGURE 8 - The tectonic episodes of the second orogenic stage and related deformational (D), metamorphic (M) and magmatic (δ) (modified from Heilbron et al., 2000); Time intervals are based on U/Pb data from Machado et al., 1996; Tupinambá et al., 1998.

FIGURE 9 - Interpretative cartoons for the evolution of the Brasiliano collage. SFC - São Francisco and C - Congo continents; SMM - Serra do Mar Microplate. Basins: A - Andrelândia, It - Itabra, SF - São Francisco, W - West Congo, B - Búzios. Brasiliano belts: BB - Brasília, RB - Ribeira, AB - Araçuaí and WB-West Congo. SA - Sangha aulacogen. X - magmatic arc. Arrows indicate the tectonic vergence.

According to Ribeiro *et al.* (1995) and Trouw *et al.* (1997), another extensional event, poorly constrained, but tentatively related to the Ectasian period (*c.* 1.3 Ga) of the Mesoproterozoic, led to the deformation and tilting of the sequences in the São João del Rei rift basin, predating the Carandaí Basin (Ribeiro *et al.*, 1995; Ribeiro, 1997). Basic dykes, related to this extensional episode, intruded the rift successions of the São João del Rei Basin. The main rifting stage is represented by the NE-SW Prados Graben, whereas the shallow carbonate Carandaí Platform records the post-rift thermal subsidence phase. The repetitive sedimentary successions and the absence of structures and magmatism suggest conditions of tectonic stability during the development of the Carandaí Basin.

At the end of the Mesoproterozoic, the Grenville amalgamation led to the formation of the Rodinia Supercontinent (Hoffman, 1991). Clear evidence of the imprint of the Greenville Orogeny has only been reported in the central, northern and northeastern regions of Brazil (Brito Neves *et al.*, 1995). Probably, the southeastern region of Brazil was located relatively far from Grenville front.

The Mesoproterozoic to Neoproterozoic transition displays the break up and dispersion of the Rodinia and other major continental masses. Neoproterozoic rifting along the older NE-SW Transamazonian Suture was responsible for the opening of the Ribeira Trough, a branch of the Adamastor Ocean. These processes were probably diachronic, and generated important passive margin basins. Both the Brasília and Ribeira passive margins, located westward and southeastward of the São Francisco-Congo continent, respectively, developed during this time. Although poorly constrained, this tectonic phase is probably related to the 1.0 Ga - 900 Ma interval, based on Sm/Nd model ages of mafic dykes (Heilbron *et al.*, 1990; Trouw and Pankhurst, 1993). The Ribeira Belt passive margins may be tentatively restored within the different tectonic domains. The Andrelândia Basin developed at the border of the São Francisco Paleocontinent in response to the break-up of Rodinia (Fig. 9). The Andrelândia Basin comprised an early rift stage followed by a passive margin stage (Paciullo, 1997). The Neoproterozoic successions of the Andrelândia Basin (1.0 Ga - 600 Ma) record a dramatic change in regional paleo-slope and tectonic setting. The nature and distribution of the lithofacies associations in this basin indicate that deposition took place in an intraplate continental margin with the basin depocentre located to the SW of an NW-SE oriented coast line.

Another Neoproterozoic passive margin is inferred, probably at the same time, at the border of the Oriental Terrane (Serra do Mar Plate or Microplate). The Italva Basin comprises a (meta) pelite-carbonate lithological association that, in spite of the high-grade metamorphism (granulite facies or the transition between amphibolite and granulite facies) and intense tectonic transposition, may be regarded as indicative of a former shelf environment (Heilbron *et al.*, 2000). The Búzios Basin, located eastward of the Serra do Mar Microplate, is represented by a pelitic succession with amphibolitic layers probably related to a back-arc setting.

The Brasiliano orogenic stages were responsible for the closure of the Neoproterozoic passive margins. The first orogenic stage occurred between 740 - 640 Ma and was related to the evolution of the southern segment of the Brasília Belt, that began with westward subduction of the western margin of the São Francisco Plate, generating a magmatic arc located in at the upper Paraná Plate, now represented by the Guaxupé Terrane (Fig. 9). Subsequent collision produced synformal nappes with top-to-E movement, associated with high pressure metamorphism. The distribution of granulite and retro-eclogite outlines the collisional suture.

The second orogenic stage (630 - 520 Ma) is related to the interaction of the Serra do Mar Microplate, that resulted in eastward subduction of the southeastern and eastern margins of the São Francisco Pate (Fig. 9). A new magmatic arc, the Rio Negro Arc, was emplaced in the Serra do Mar microplate region (Tupinambá *et al.*, 1998; Heilbron *et al.*, 1998b, 2000). The magmatic rocks intruded the Italva passive margin. Subsequent oblique collision (590 - 550 Ma) between these two plates caused the development of a W-vergent thrust system with an important dextral component (Fig. 9), intermediate pressure metamorphism, and the emplacement of abundant granitoid plutons, derived from the remelting of both basement and cover successions (Machado *et al.*, 1996; Heilbron *et al.*, 1995). The Central Tectonic Boundary outlines the suture zone of the collision. The late tectonic episode of the second orogenic stage (520 - 480 Ma) is associated to subvertical folding and steep shear zones that controlled the emplacement of widespread granitic plutonism, mostly associated with tholeiitic rocks. The melting of crustal and mantle rocks at high temperatures was probably related to the slab detachment and uplift just after the Brasiliano collision.

The third orogenic stage (520 - 480 Ma) is related to the closure of the Buzios Basin, with NNW thrusting of the Cabo Frio Terrane and associated intermediate to high-pressure metamorphism, only detected at the Cabo Frio-Búzios Cape region (Heilbron *et al.*, 1982; Schmitt and Trouw, 1997; Schmitt *et al.*, 1999).

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Time relationship with deformation	Age (Ma)	Geochemistry	Structures	Tectonic Domains	Tectonic setting	References
Slab detachment related granitoid	< 520	Calc-alkaline to alkali-calcic plutons Tholeitic gabbro and diorite	Non-foliated and mostly zoned stocks and sheets, primary flux foliation and layering, dikes, proposition of mixing and minging process	Only detected in the Oriental Terrane	Extensional Collapse of the orogenic belt	Wiedemann,1993; Junho, 1993; Figueiredo and Campos Neto, 1993
Late-collisional syn-D3	540-520	Alkali-calcic stocks of slightly peraluminous leucogranites	Subvertical dykes and plutons related to D3 shear zones, weakly foliated at the borders	Recorded in both tectonic terranes	Crustal reworking of basement and cover rocks	Heilbron, 1993, 1995; Heilbron et al., 1995; Machado et al., 1996
Syn-collisional late-D2	565-540	Abundant metaluminous I-type granites with basic enclaves Subordinated two-mica peraluminous S-type leucogranite	Weakly foliated plutons and sheets	Recorded in both tectonic terranes	Crustal reworking of basement and cover rocks	Grossi Sad and Barbosa, 1985; Tupinambá, 1993a; Heilbron, 1995; Heilbron et al., 1995
Early Syn-collisional syn-D1+D2	595-565	Abundant peraluminous S-type granite Metaluminous I-type granite	Foliated and mylonitic plutons	Recorded in both tectonic terranes	Crustal reworking of basement and cover rocks	Tupinambá, 1993a; Heilbron, 1995; Heilbron et al., 1995; Machado et al., 1996
Pre-collisional pre-D1	630-595	Tonalites to granodiorite and tholeitic gabbro	Foliated plutons, mylonitic structures associated with tectonic boundaries	Only detected in the Oriental Terrane	Cordilleran Magmatic Arc	Figueiredo and Campos Neto, 1993; Tupinambá et al., 1998

Table 1: Basement, Proterozoic basins and granitoid rocks within the tectonostratigraphic units of the Central Ribeira Belt (modified from Heilbron et al., 2000).

Tectono-stratigraphic units	**Pre-1.8 Ga basement**	**Post-1.8 Ga basins**	**Brasiliano orogeny granitoid pluton**
Occidental Terrane			
Autochthonous Domain	Greenstone belt Barbacena Mantiqueira Complex	Andrelândia Basin Carandaí Platform Prados Graben São João del Rei Basin	
Nappe systems within the interference zone	Greenstone belt Barbacena Mantiqueira Complex	Andrelândia Basin	
Andrelândia Thrust System	Mantiqueira Complex	Andrelândia Basin	syn to late collisional granitoid plutons
Juiz de Fora Thrust System	Juiz de Fora Complex	Andrelândia Basin	syn to post collisional granitoid plutons
Oriental Terrane Paraíba do Sul Klippe	Quirino Complex	Paraíba do Sul Basin (?)	syn to post collisional granitoid plutons
Costeiro Domain	?	Italva Basin	pre, syn to post collisional and post tectonic granitoid plutons
Cabo Frio Domain	Região dos Lagos Complex	Búzios Succession	Leucogranite

Table 2 - General characteristics of the Brasiliano granitoids of the Ocidental and Oriental terranes, classified according to time relationships with the deformation phases (modified after Heilbron et al; 1995)

REFERENCES

Alkmim, F.F. and Marshak, S. (1998). Transamazonian Orogeny in the Southern São Francisco Craton Region, Minas Gerais, Brazil: evidence for Paleoproterozoic collision and collapse in the Quadrilátero Ferrífero. *Precambrian Research*, **90**, 29-58.

Almeida, F.F.M. and Hasui, Y. (1984). *O Pré-Cambriano do Brasil.* Edgard Blucher, 378 p.

Almeida, F.F.M. de, Amaral, G., Cordani, U. and Kawashita, K. (1973). The Precambrian evolution of the South American Cratonic Margin, South of Amazon River. In: *The ocean basin and margins*, 1, eds. Nairn and Stelli, pp. 411-416. Plenum, Nova York.

Almeida, F.F.M. de, Hasui, Y. and Brito Neves, B.B. (1976). The Upper Precambrian of South America. *Bol. do Instituto de Geociências USP*, 7, 45-80, São Paulo, SP.

Almeida, J.C.H. (1992). *Mapeamento geológico da Folha Luminárias-MG (1:50.000), com ênfase na análise estrutural dos metassedimentos do CicloDeposicional Andrelândia.* Tese de Mestrado, IGEO-UFRJ, Rio de Janeiro, 102 p.

Almeida, J.C.H., Silva, L.G.E. and Valladares, C.S. (1993). O Grupo Paraíba do Sul e rochas granitóides na região de Bananal (SP) e Rio Claro (RJ): uma proposta de formalizacão litoestratigráfica. In: *Atas do III Simpósio de. Geologia do Sudeste, Rio de Janeiro, SBG*, 161-167.

Almeida, J.C.H., Tupinambá, M., Heilbron, M. and Trouw, R. (1998). Geometric and kinematic analysis at the Central Tectonic Boundary of the Ribeira belt, Southeastern Brazil. In: *Anais 39 Congresso Brasileiro Geologia*, Belo Horizonte, 32 p.

Ávila, C.A., Valença, J.G. and Dutra, D.C. (1998a). Rochas meta-subvulcânicas félsicas presentes no embasamento de São João del Rei, Minas Gerais. Anais do XL Congresso Brasileiro de Geologia, Belo Horizonte, 494 p.

Avila, C.A., Moura, C.A.V., Valença, J.G., Ribeiro, A. and Paciullo, F.V.P. (1998b). Idades Pb-Pb em zircões de corpos plutônicos da região de São João del Rei, borda sul do Cráton do São Francisco, sul de Minas Gerais. *Cong. Brasil. Geol. 40 Belo Horizonte, MG. Anais*, 34 p.

Barbosa, O. (1954). Évolution du géosynclinal Espinhaço. In: INTER. GEOL. CONGR. 19, Alger, 1952, *Comptes Rendus*, Alger, sec. 13, fasc. 14, p. 17-36.

Brandalise, L. A. et al. (1991). *Folha Barbacena, SF.23-X-C-III, escala 1:100.000, Estado de Minas Gerais. Programa Levantamentos Geológicos Básicos do Brasil.* Brasília, DNPM/CPRM, 162 p. (Texto explicativo).

Brito Neves, B.B. and Cordani, U.G. (1991). Tectonic Evolution of South America during the Late Proterozoic. *Precambrian Research*, 53, 23-40.

Brito Neves, B.B., Sá, J.M., Nilson, A.A. and Botelho, N.F. (1995). A tafrogênese Estateriana nos blocos paleoproterozóicos da América do Sul. *Bol. do IG-USP, Série Científica*, **27**, 1-40.

Campanha, G.A.C. and Ferrari, A.L. (1984). Lineamento Além Paraíba: um exemplo de zona de cisalhamento. In: *Anais 38 Congresso Brasileiro de Geologia, Rio de Janeiro*, 12, 5425-5432.

Campos Neto, M.C. and Figueiredo, M.C.H. (1995). The Rio Doce orogeny, Southeastern Brazil, *Journal of South American Earth Sciences*, 8(2), 143-162.

Campos Neto, M.C. and Caby, R. (1999). Neoproterozoic high-pressure metamorphism and tectonic constrain from nappe system south of the São Franciso craton, southeast Brazil. *Precambrian Research* (in press).

Chrispim, S.J. and Tupinambá, M. (1989). Observações sobre o caráter transpressivo da deformação entre as zonas de cisalhamento de Guaçuí e Paraíba do Sul. In: *Anais do I Simp. Geol. Sudeste*, Rio Janeiro, 120-121.

Cordani, U.G., Delhal, J. and Ledent, D. (1973). Orogenéses Superposées dans Le Precambrien du Brésil Sud-Oriental (États de Rio de Janeiro et de Minas Gerais). *Revista Brasileira de Geociências*, 3, 1-22.

Cordani, U.G., Brito Neves, B.B., Fuck, R.A., Porto., R., Thomaz Filho, A. and. Cunha, F.M.B. (1984). Estudo preliminar de integração do Pré-Cambriano com os eventos tectônicos das bacias sedimentares brasileiras, *Petrobras, Série Ciência-Técnica-Petróleo*, Rio de Janeiro, 15, 70 p.

Correia Neto, A. V., Dayan, H., Valença, J.G. and Cabral, A. (1993). Geologia e Estrutura da Zona de Cisalhamento do Rio Paraíba do Sul e Adjacências, no trecho entre Três Rios e Sapucaia (RJ). In: *Atas III Simpósio de Geologia do. Sudeste, Rio de Janeiro, SBG*, 194-200.

Cunningham, D., Alkmim, F.F. and Marshak, S. (1998). A Structural transect across the Coastal Mobile Belt in the Brazilian Highlands (latitude 20 S): The roots of a Precambrian transpressional orogen. *Precambrian Research*, **92**, 251-275

Del Lama, E.A., Oliveira, M.A.F. and Zanardo, A. (1993). Geotermobarometria aplicada a rochas do sul do complexo Campos Gerais. In: *Simpósio de Geologia de Minas Gerais, 7. Belo Horizonte, 1993, Bol.* **12**, 93-94.

Delhal, J., Leden, D. and Cordani, U.G. (1969). Ages Pb/U, Sr/Rb et Ar/K de Formations Métamorphiques et Granitiques du Sud-Est du Brésil (États de Rio de Janeiro e Minas Gerais). *Annales de la Societé Géologique de Belgique*, **92**, 271-283.

Duarte, P.D. (1998). *Evolução tectônica dos ortognaisses dos Complexos Juiz de Fora e Mantiqueira na região de Juiz de Fora, MG: Geologia, petrologia e geoquímica.* Thesis, Instituto de Geociências, Universidade de São Paulo, 280 p. (unpublished).

Duarte, B.P., Figueiredo, M.C.H., Campos Neto, M. and Heilbron, M. (1997). Geochemistry of the Granulite Facies Orthogneisses of Juiz de Fora Complex, Central Segment of Ribeira belt, Southeast Brazil, *Revista Brasileira de Geociências*, **27**, 1, 67-82.

Ebert, H. (1955). Relatório de Atividades. In: *Relatório Anual do Diretor, ano 1954, DNPM Div. Geol. Min, Rio de Janeiro*, 79-89.

Ebert, H. (1957). Aspectos principais da geologia de São João del Rei, Estado de Minas Gerais. Relatório inédito da PROSPEC, 120 p. *SBG Núcleo São Paulo Publ.* **12**/1984, 1-71.

Ebert, H. (1971). Os Paraíbiades entre São João del Rei, MG e Itapira, SP, e a bifurcação entre Paraíbiades e Araxaíades. Boletim da Faculdade de Filosofia Ciências e Letras de Rio Claro, SP. SBG Núcleo São Paulo. Publ., Bol. Especial 1, 12/1984, 72-103.

Ebert, H. D., Hasui, Y. and Costa, J.B.S. (1991). O caráter transpressivo do Cinturão de Cisalhamento Rio Paraíba do Sul. In: *Boletim III Simpósio Nacional de Estudos Tectônicos, Rio Claro, SBG*, 139-141.

Ebert, H.D. and Hasui, Y. (1998). Transpressional tectonics and strain partitioning during oblique collision beteween three plates in the Precambrian of southeast Brazil. In: Continental Transpressional and Transtensional Tectonics, eds. Holdsworth, R.E., Strachan, R.A. and Dewey, J.D. *Geological Society, London, Special Publications*, **135**, 231-252.

Ebert, H.D., Neves, M.A., Hasui, Y., Lopes, J.A. and Guerra, M. (1995). Compartimentação crustal e evolução cinemática da Província Mantiqueira através de modelagem física. In: *V Simpósio Nacional de Estudos Tectônicos*, Gramado, 26-28.

Ebert, H.D., Chemale Jr., F., Babinski, M., Artur, A.C. and Van Schmus, W.R. (1996). Tectonic setting and U/Pb zircon dating of the plutonic Socorro Complex in the Transpressive Rio Paraíba do Sul Shear Belt, SE Brazil. *Tectonics*, 15(2), 688-699.

Figueiredo, M.C.H. and Campos Neto, M.C. (1993). Geochemistry of the Rio Doce Magmatic Arc, Southeastern Brazil. *Anais da Academia Brasileira de Ciências*, **65**(1), 63-82.

Figueiredo, M.C.H. and Teixeira, W. (1996). The Mantiqueira

Metamorphic Complex, Eastern Minas Gerais State: preliminary geochronological and geochemical results. *Anais da Academia Brasileira de Ciências*, 68(2), 223-246.

Fonseca, A.C. (1994). *Esboço Geocronológico da Região de Cabo Frio, Estado do Rio de Janeiro*. Thesis, IG/USP, São Paulo, 186 p. (unpublished).

Fonseca, A.C., Cordani., U.G. and Kawashita, K. (1984). Dados preliminares sobre a geocronologia e suas encaixantes na cidade do Rio de Janeiro. Método Rb/Sr. In: *Anais do XXXIII Congresso Brasileiro de Geologia, Rio de Janeiro*, 6, 2333-2345.

Fonseca, M.J.G. (1998). *Mapa geológico do Estado do Rio de Janeiro Escala 1:400.000, Texto Explicativo*. DNPM/MME Rio de Janeiro, 141 p.

Fonseca, M.J.G., Heilbron, M. and Chrispim, S.J. (1984). Geologia da área de Cabo Frio e Armação dos Búzios, in *Anais XXXIII Congresso Brasileiro de. Geologia, Rio de Janeiro, SBG*, 12, 5393-5424.

Gonçalves, M.L. and Figueiredo, M.C.H. (1992). Geoquímica dos Anfibolitos de Santana do Garambéu (MG): Implicações sobre a Evolução do Grupo Andrelândia. *Geochimica Brasiliensis*, 6(2), 127-140.

Grossi Sad, J.H. and Barbosa, A.L. (1985). A origem dos charnockitos e rochas afins da região do Médio Paraíba do Sul, estado do Rio de Janeiro, Contribuições à geologia e petrologia. *Boletim Especial da SBG, Núcleo de Minas Gerais*, 15, 281 p.

Grossi Sad, J.H. and.Dutra, C.V. (1998). Chemical composition of supracrustal rocks from Paraíba do Sul group, Rio de Janeiro State, Brazil. *Geochimica Brasiliensis*, 2(2), 143-166.

Guimarães, D. (1951). *Arqui-Brasil e sua Evolução Geológica*. Div. Fom. Prod. Min. DNPM, Min. da Agric. Boletim 88, ano de 1951, Rio de Janeiro, DF, 317 p.

Hasui, Y. and Oliveira, M. A. F. 1984. Província Mantiqueira, setor central. In: *O Pré-Cambriano do Brasil*, eds. Almeida, F.F.M. and Hasui, Y. Edgard Blücher Ltda, São Paulo, 378 p.

Heilbron, M. (1985). O metamorfismo da área de Itutinga - Madre de Deus de Minas, MG. *Simp. Geol. Minas Gerais, 3,-SBG-MG, Belo Horizonte, Bol.* 5, 219-234.

Heilbron, M. (1993). *Evolução tectono-metamórfica da seção Bom Jardim de Minas (MG) - Barra do Piraí (RJ). Setor Central da Faixa Ribeira*. Thesis, IG/USP, 268 p. (unpublished).

Heilbron, M. (1995). *O segmento central da Faixa Ribeira: síntese geológica e ensaio de evolução geotectônica*. Thesis, Faculdade de Geologia, UERJ, Rio de Janeiro, 110 p. (unpublished).

Heilbron, M., Chrispim, S.J., Simões L.S.A. and Alves, R.P. (1982). Geologia do cabo dos Búzios, *Anais da Academia Brasileira de Ciências*, 54(3), 553-562.

Heilbron, M., Gonçalves, M.L., Teixeira, W., Trouw, R., Kawashita, K. and Padilha, A. (1990). Geocronologia da área entre Lavras, São João del Rei, Lima Duarte e Caxambú, *Anais da Academia Brasileira de Ciências*, 61(2), 177-199.

Heilbron, M., Valeriano, C.M., Almeida, J.C.H. and Tupinambá, M. (1991). A Megassinforma do Paraíba do Sul e sua implicação na compartimentação tectônica do setor central da Faixa Ribeira. In: *Atas 2 Simpósio de Geologia do Sudeste, São Paulo, SBG*, 519-527.

Heilbron, M., Pires, F.R.M., Valeriano, C.M. and Bessa, M. (in memoriam). (1993). Litoestratigrafia, evolução tectono-metamórfica e magmatismo no Precambriano do setor sudeste do Município do Rio de Janeiro. *Anais do III Simpósio de Geologia do Sudeste, Rio de Janeiro, Outubro/1993. SBG, Núcleos Rio de Janeiro/São Paulo*, 174-179.

Heilbron, M., Valeriano, C.M., Valladares, C.S., and Machado, N. (1995). A orogênese Brasiliana no segmento central da Faixa Ribeira, Brasil, *Revista Brasileira de Geociências*, 25(4), 249-266.

Heilbron, M., Machado, R. and Figueiredo, M.C.H. (1997). Lithogeochemistry of Paleoproterozoic Orthogranulites from the Bom Jardim de Minas (MG)-Vassouras (RJ) Transect, Central Ribeira belt, SE Brazil. *Revista Brasileira de Geociências*, 27(1), 83-98.

Heilbron, M., Duarte, B.P. and Nogueira, J.R. (1998a). The Juiz de Fora Complex of the Central Ribeira belt, SE Brazil: a segment of Paleoproterozoic granulitic crust thusted during the Pan African Oregon, *International Geoscience Journal/Gondwana Research*, 1(3-4), 373-382.

Heilbron, M., Tupinambá, M., Almeida, J.C.H.,Valeriano, C.M., Valladares C.S. and Duarte, B.P. (1998b). New constraints on the tectonic organization and structural styles related to the Brasiliano collage of the central segment of the Ribeira belt, SE Brazil, in *Abstracts of the International Conference on Precambrian and Craton Tectonics/14th International Conference on Basement Tectonics*, Ouro Preto, Minas Gerais, Brasil, Extended Abstracts, pp. 15-17.

Heilbron, M., Mohriak, W., Valeriano, C.M., Milani, E., Almeida, J.C.H and Tupinambá, M. (2000) From Collision to Extension: The Roots of the Southeastern Continental Margin of Brazil. In: Mohriak, W.U. and Talwani, M. (Eds.), *Atlantic rifts and continental margins*. Washington, American Geophysical Union, Geoph. Monogr. Ser. 115, pp. 1-32.

Hoffman, P.F. 1991. Did the break-up of laurentia turn Gondwana inside-out? *Science*, 254(5011):1409-1412.

Howell, D.G. 1989, *Tectonics of Suspect Terranes*. Chapman and Hall (eds.), London, 232 p.

Janasi, V.A., Vlach, S.R.F. and Ulbrich, H.h.G.J. (1993). Enriched-mantle contributions to the Itu-Granitoid belt, Se Brazil: Evidence from K-rich diorites and syenites. *Anais da Academia Brasileira de Ciências*, 65, 177-118.

Junho, M.C.B. (1993). Granitóides Brasiliano s da região central do Rio de Janeiro- geoquímica preliminar, *Anais da Academia Brasileira de Ciências*, 65, 2: 162-179.

Leonardos Jr., O.H. and Fyfe, W.S. (1974). Ultrametamorphism and melting of a continental margin: The Rio de Janeiro region, Brazil. *Contributions to Mineralogy and Petrology*, 46, 201-214.

Lesquer, A. (1981). Signification structurale des anomalies gravimétriques de la partie sud du craton de São Francisco (Brésil). *Tectonophysics*, 76, 273-293.

Machado Filho, L., Ribeiro, M., Gonzales, S.R., Schenini, C.A., Santos Neto, A., Palmeira, R.C., Pires, J.L., Teixeira, W. and Castro, H.E.F. (1983). *Geologia das folhas Rio de Janeiro (SF 23/24) escala 1:1.000.000, mapa e texto explicativo*. RADAM Brasil-Ministério das Minas e Energia, Rio de Janeiro, 780 p.

Machado, N. and Gauthier, G. (1996). Detrmination of 207Pb/206Pb ages on zircon and monazite by laser-ablation ICPMS and aplication to sedimentary provenance and metamorphism in sotheastern Brazil. *Geochimica et Cosmochimica Acta*, 60 (24), 5063-5073.

Machado, N., Valladares, C.S., Heilbron, M. and Valeriano, C.M. (1996). U/Pb Geochronology of Central Ribeira belt. *Precambrian Research*, 79, 347-361.

Machado, R. (1984). *Evolução geológica, análise estrutural e metamórfica da região de Vassouras e Paracambi, Rio de Janeiro*. Thesis IG-USP, 196 p. (unpublished).

Machado, R and Endo, I. (1993). Estruturas transcorrentes na borda sul do Cráton do São Francisco. In: *Simpósio sobre o Cráton do São Francisco, II, Salvador, Anais, Salvador, Bahia, SBG*, 269-271.

Malagutti Filho, W., Ebert, H.D., Hasui, Y., Haralyi, N.L.E., Hackspacher, P.C., Sturaro, J.R., Souza, C.A. and Almeida, S.H.S. (1996). Gravimetria e compartimentação crustal do sul de Minas Gerais. *Geociências* (Ed. Unesp, São Paulo), 15 (n. esp.), 199-217.

Noce, C.M., Hoppe, A., Trouw, R.A.J. and Karfunkel, J. (1987). O Pré-

Cambriano inferior da região de Barbacena-Tiradentes, Minas Gerais. *Simpósio de Geologia de Minas Gerais 4, Belo Horizonte, MG, Anais*, 149-163.

Nogueira, J.R., Choudhuri, A. and Trouw, R. (1996). Fluid inclusions studies, P-T-t paths and geotectonic evolution of the granulites facies terrain of the Juiz de Fora, southeastern Brazil. In: *Proc. Intern. Symposium in Charnockite and Granulite facies rocks*. Geolgist'Association of Tamil Nadu, Madras, India.

Oliveira, M.A.F. (1982). Bulk geochemistry of the Paraíba do Sul Granulitic belt, *Revista Brasileira de Geociências*, **12**(1-3), 369-374.

Paciullo, F.V.P. (1997). *A sequência deposicional Andrelândia*. Thesis, Instituto de Geociências da Universidade Federal do Rio de Janeiro, Rio de Janeiro, 245 p. (unpublished).

Paciullo, F.V.P., Ribeiro, A. and Andreis, R.R. (1993). Reconstrucão de uma bacia fragmentada: o caso do Ciclo Deposicional Andrelândia. In: *Anais II Simpósio sobre o Cráton do São Francisco, Salvador, Bahia, SBG*, 224-226.

Paciullo, F.V.P. Ribeiro, A., Andreis, R.R. and Trouw, R.A.J. (1998). Sedimentary, igneous and thermo-tectonic events in the folded belts at the southern border of the São Francisco craton. In: *Abstracts of the International Conference on Precambrian and Craton Tectonics/14th International Conference on Basement Tectonics, Ouro Preto, Brazil*, 68-69.

Pinto, C.P., Brandalise, L.A., Viana, H.S. and Bruno, E.M. (1992). Suíte metamórfica São Bento dos Torres, Serra da Mantiqueira - MG. *Revista Escola de Minas, Ouro Preto*, **45**(1-2), 187-189.

Pires, F. R. M. and Heilbron, M. (1986). Polimetamorfismo em gnaisses do Rio de Janeiro. In: *XXXIV Congresso Brasileiro de Geologia, Anais*, 1451-1463.

Ribeiro, A. (1997). *Estratigrafia e paleoambientes nas sucessões metassedimentares proterozóicas das serras do Lenheiro e São José, São João del Rei, Sul de Minas Gerais*. Thesis, Instituto de Geociências da Universidade Federal do Rio de Janeiro, Rio de Janeiro, 167 p. (unpublished).

Ribeiro, A. and Silva, R.R. (1996). Possíveis correlações entre bacias proterozóicas a sul e sudeste do Cráton do São Francisco, Minas Gerais. In: *Congr. Bras. Geol, 39, Salvador, BA, 1996. Anais, Salvador, SBG*, **6**, 11-13.

Ribeiro, A., Trouw, R.A.J., Paciullo, F.V.P., Andreis, R.R. and Valença, J.G. (1995). Evolução das bacias proterozóicas e termo-tectonismo Brasiliano na margem sul do Cráton do São Francisco, *Revista Brasileira de Geociências*, **25**(4), 235-248.

Ribeiro, A.,. Paciullo, F.V.P., Valença, J.G., Ávila, C.A., Andreis, R.R. and Trouw, R.A.J. (1998). Domínios do embasamento na região de São João del Rei, borda sul do Cráton do São Francisco, sul de Minas Gerais. *Cong. Brasil. Geol. 40 Belo Horizonte, MG. Anais*, 21.

Rogers, J.J.W. (1996). A history of the continentes in the past three billion years. *Journal of Geology*, **4**, 1405-1412.

Rosier, G.F. (1965). *Pesquisas geológicas na parte oriental do Estado do Rio de Janeiro e na parte vizinha de Minas Gerais*. DNPM, Div. Geol. Min., Boletim 222, Rio de Janeiro.

Sad, J.H.G. and Dutra, C. (1988). Chemical composition of supracrustal rocks from the Paraíba do Sul Group, Rio de Janeiro State, Brazil. *Geochimica Brasiliensis*, **7**(2), 143-174.

Schmitt, R.S. and Trouw, R.A.J. (1997). Estudo preliminar dos indicadores de transporte tectônico pré-cambriano na Região dos Lagos, Rio de Janeiro. In: *Atas do V Simpósio de Geologia do Sudeste, Penedo, Rio de Janeiro*, 142-144.

Schmitt, R.S., Trouw, R.A.J. and Van Schmus, W.R. (1999). The characterization of a Cambrian (~ 520 Ma) tectonometamorphic event in the Costeiro Domain of the Ribeira belt, using U/Pb in syntectonic veins. In: *Extended Abstracts South American Symposium on Isotope Geology* (in press).

Simões, L.S.A. (1995). *Evolução tectonometamórfica da nappe de Passos, sudoeste de Minas Gerais*. Tese de Doutorado, IG-Universidade de São Paulo, 149 p. (inéd.).

Simões, L.S.A. and Valeriano, C.M. (1990). Porção meridional da Faixa de Dobramentos Brasília: estágio atual do conhecimento e problemas de correlação tectono-estratigráfica. *Anais do XXXVI Congresso Brasileiro de Geologia, Natal, SBG*, **6**, 2564-2575.

Simões, L.S.A., Girardi, V.A.V. and Valeriano, C.M. (1994). O metamorfism-o inverso da porção meridional da Faixa Brasília. *Anais do XXXVIII Congresso Brasileiro de Geologia, Balneário Camboriú, SBG*, **1**, 253-254.

Söllner, F., Lammerer, B., Weber-Diefenbach, K. (1991). Die Krustenentwicklung in der Kusternregion nördlich von Rio de Janeiro/Brasilien. Munchner Geologische Hefte 4, Munchen, 100 p.

Söllner, F. and Trouw, R.A.J. (1997). The Andrelândia depositional cycle (Minas Gerais, Brazil), a post-Transamazonic sequence south of the São Francisco craton: evidence from U-Pb dating on zircons of a metasediment. *Journal of South American Earth Sciences*, **10**(1), 21-28.

Teixeira, W. (1985). *A evolução geotectônica da porção meridional do Cráton do São Francisco, com base em interpretações geocronológicas*. Tese de Doutoramento, IG-Universidade de São Paulo, 207 p. (inéd.).

Teixeira, W. (1993). Avaliação do acervo dedados geocronológicos e isotópicos do Cráton do São Francisco - implicações tectônicas. In: Dominguez, . M. L. and Misi, A. (eds.), *O Cráton do São Francisco*: trabalhos apresentados na reunião preparatória do II Simpósio sobre o Cráton do São Francisco, Salvador, 1992. SBG-SGM-CNPq, pp.11-34.

Teixeira, W. and Figueiredo, M.C.H. (1991). An outline of Early Proterozoic crustal evolution in the São Francisco Craton, Brazil: a review. *Precambrian Research*, 53, 1-22.

Teixeira, W., Carneiro, M.A., Noce, C.M., Machado, N., Sato, K. and Taylor, P.N. (1996). Pb, Sr and Nd isotope constraints on the Archean evolution of gnaissi-granitoid complexes in the Southern São Francisco Craton, Brazil. *Precambrian Research*, **78**, 151-164

Töpfner, C. (1996). *Brasiliano -granitoide in den Bundestaten São Paulo und Minas Gerais, Brasilien- eine vergleischende studie*. Tese de Doutorado, Universidade de Munique, *Münchner Geologische Hefte* A17, 258 p.

Trompette, R., Egydio Silva, M., Tommasi, A., Vauchez, A., and Uhlein, A. (1993). Amalgamação do Continente Gondwana no Panafricano-Brasiliano e o papel da geometria do CSF na arquitetura da Faixa Ribeira. *Revista Brasileira de Geociências*, **23**(3), 187-193.

Trompette, R. (1994). *Geology of Western Gondwana (2000-500 Ma)*. A.A. Balkema, Rotterdam, 350 p.

Trouw, R.A.J. (1995). Uma avaliação de modelos cinemáticos propostos para o megassinformal do Vale do Rio Paraíba do Sul, baseada em lineações de estiramento. *V Simpósio Nacional de Estudos Tectônicos, Gramado*, 103-104.

Trouw, R.A.J. and Pankhurst, R.J. (1993). Idades radiométricas ao sul do cráton do São Francisco: região da folha Barbacena, Minas Gerais. In: *Anais do II Simpósio sobre o Cráton do São Francisco, Salvador, Bahia, SBG*, 260-263.

Trouw, R.A.J. and Oliveira Castro, E.M. (1995). O significado tectônico de granulitos Brasiliano s de alta de pressão no sul de Minas Gerais. *Congresso Brasileiro de Geologia*, Salvador, BA, **39**(6), 145-151.

Trouw, R. A. J., Paciullo, F. V. P., Chrispim, S. J. and Dayan, H. (1982). Análise de deformação numa área a SE de Lavras, Minas Gerais. *Congresso Brasileiro de Geologia*, Salvador, BA, **32**(1), 187-198.

Trouw, R.A.J., Ribeiro, A. and Paciullo, F.V.P. (1986). Contribuição à geologia da folha Barbacena 1:250.000. In: *Anais do XXXIV Congresso Brasileiro de Geologia, Goiânia, SBG*, 2, 972-986.

Trouw, R.A.J., Paciullo, F.V.P. and Ribeiro, A. (1994). A Faixa Alto Rio Grande reinterpretada como a Zona de Intereferência entre a Faixa Brasília e a Faixa Ribeira in *Boletim de Resumos Expandidos, XXXVIII Congresso Brasileiro de Geologia, Camboriu*, 3, 234-235.

Trouw, R.A.J., Ribeiro, A., Paciullo, F.V.P. and Andreis, R.R. (1997). A passagem do mesoproterozóico para o neoproterozóico na região sul de Minas Gerais. In: *IX Simpósio de Geologia de Minas Gerais, Ouro Preto, SBG/MG*, 14, 10-11.

Trouw, R. A. J., Paciullo, F. V. P and Ribeiro, A. (1998). Tectonic significance of Neoproterozoic high-pressure granulites in southern Minas Gerais. *14th International Conference on Basement Tectonics. Ouro Preto, MG. International Basement Tectonics Association-UFOP, Abstracts*, 69-71.

Tupinambá, M. (1993a). Rochas intrusivas e metassedimentos granulíticos do Complexo Paraíba do Sul na parte setentrional da Faixa Ribeira. In: *Atas do III Simpósio de Geologia do Sudeste, SBG/RJ, Rio de Janeiro*, 187-193.

Tupinambá, M. (1993b). Metamorfismo granulítico do Complexo Paraíba do Sul na porção setentrional da Faixa Ribeira, a leste do Cráton do São Francisco in *Anais do II Simpósio de Geologia do Cráton do São Francisco e Faixas Marginais*, SBG/BA, Salvador, 266-268.

Tupinambá, M., Teixeira, W. and Heilbron, M. (1998). The Pan-African/Brasiliano Arc-related Magmatism at the Costeiro Domain of the Ribeira belt, Southeastern Brazil, in *Abstracts of the International Conference on Precambrian and Craton Tectonics/14th International Conference on Basement Tectonics*, Ouro Preto, MG, 12-14.

Unrug, R. (1997). Rodinia to Gondwana: the geodynamic map of Gondwana supercontinent assembly. *GSA Today*, 7(1), 1-6.

Valença, J.G., Paciullo, F.V.P., Ribeiro, A. and Andreis, R.R. (1998). Komatiítos da Fazenda Chapadinha, Faixa Greenstone Itumirim-Tiradentes, borda sul do Cráton do São Francisco, Minas Gerais. *Cong. Brasil. Geol. 40, Belo Horizonte, MG. Anais*, 62.

Valeriano, C.M. (1985). *Geologia estrutural e estratigrafia do Grupo São João del Rei, na região de São João del Rei, MG.* Thesis. Universidade Federal do Rio de Janeiro, 98 p. (unpublished).

Valeriano, C. M. (1986). Geologia estrutural e estratigrafia do Grupo São João del Rei, na região de São João del Rei, MG. *Anais do 34° Congr. Bras. Geol.*, SBG, núcleo Centro-Oeste, Goiânia, Out/1986, 2, 999-1018.

Valeriano, C.M. (1992). *Evolução tectônica da extremidade meridional da Faixa Brasília, região da Represa de Furnas, Sudoeste de Minas Gerais.* Thesis, IG - Universidade de São Paulo, 192 p. (unpublished).

Valeriano, C.M., Almeida, J.C.H., Simões, L.S.A., Duarte, B.P. and Heilbron, M. (1995). Evolução estrutural do domínio externo da Faixa Brasília no sudoeste de Minas Gerais: registros de uma tectônica pré-Brasiliana. *Revista Brasileira de Geociências*, 25(4):221-234.

Valeriano, C.M., Simões, L.S.A., Teixeira, W. and Heilbron, M. (1998). Southern Brasilia belt (SE Brazil): thrust-discontinuities and evolution during the Neoproterozoic Brasiliano orogeny. *14th International Conference on Basement Tectonics, Ouro Preto (MG)*, 1998, 62-65.

Valladares, C.S., Heilbron, M. and Figueiredo, M.C.H. (1995). O Granito Getulândia e sua relação com a Zona de Cisalhamento Taxaquara. In: *Anais do V Simpósio Nacioanl de Estudos Tectônicos* Gramado, RS, 217-218.

Valladares, C.S., Heilbron, M., Machado, N. and Valeriano, C.M. (1997). Provenance of Central Ribeira belt metasediments using 207Pb/206Pb ages on zircon by laser-ablation ICPMS. In: *Extended Abstracts South American Symposium on Isotope Geology*, Campos do Jordão, SP, 323-325.

Van Schmus, W.R., Bickford, M.E. and Condie, K.C. (1993). Early Proterozoic evolution. In: *Precambrian: conterminous*, eds. Redde, J.C., M.E. Bickford, R.S. Houston, P.K. Link, Rankin, D.W., Sims, P.K., W.R. Van Schmus. U.S. Boulder, Geological Society of America. *The Geology of North America*, C-2, 270-281.

Vauchez, A., Tommasi, A., Egydio da Silva, M. and Tromppette, R. (1992). Tectônica de escape na terminação de um craton: A Faixa Ribeira. In: *XXXVII Congresso Brasileiro de Geologia, São Paulo*, 2, 373-375

Vauchez, A. Tommasi, A. and Egydio Silva, M. (1994). Self-indentation of a heterogeneous continental lithosphere. *Geology*, 22, 967-970.

Wernick, E. and Töpfner, C. (1997). Idades do magmatismo granitóide Brasiliano nos estados de São Paulo e Minas Gerais: implicações na caracterização temporal da evolução do cinturão Ribeira. *V Simpósio de Geologia do Sudeste, Águas de São Pedro*, 8-10.

Wiedemann, C.M. (1993). Early Paleozoic, late- to post-collisional magmatic arc of the coastal mobile belt in the state of Espírito Santo, eastern Brazil. *Anais da Academia Brasileira de Ciências*, 65, supl 1, 162-181.

Zimbres, E., Kawashita, K. and Van Schmus, V.R. (1990). Evidências de um núcleo transamazônico na região de Cabo Frio, RJ e sua conexão com o Cráton de Angola. In: *Anais XXXVI Congresso Brasileiro de Geologia, Natal, RN, SBG*, 6, 2735-2743.

The Dom Feliciano Belt of Brazil and Uruguay and Its Foreland Domain, the Rio de La Plata Craton

Framework, Tectonic Evolution and Correlation with Similar Provinces of Southwestern Africa

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In southeastern South America magmatic, metamorphic, structural and geotectonic features record the superposition of Neoproterozoic-Eopaleozoic orogenies. The present geometry of these geotectonic units reflects collages of distinct terranes, culminating with the amalgamation of the Gondwana Supercontinent. The Brasiliano (900 - 620 Ma) and Rio Doce (620 - 530 Ma) orogenies that took place in the region are manifest by remnants of magmatic arcs and metavolcano-sedimentary cover, well represented by the Ribeira and Dom Feliciano belts.

Metamorphic episodes and collisional events of the Brasiliano Cycle reached their climax around 700 ± 50 Ma and 640 ± 20 Ma, respectively. These orogenic episodes finished with an ample allochthony with preferential NNW transport. Late magmatism (600 ± 10 Ma) is conspicuous in southern Brazil, marking the end of the extensional regime of events related to the Brasiliano Cycle (volcano-sedimentary basins and alkaline-peralkaline granitoid of the Serra do Mar Suite).

With the stabilization of the Brasiliano Orogeny, a series of magmatic arcs that lie today along the Brazilian coast, began to form around 620 ± 20 Ma, E of the landmass formed by the São Francisco and Rio de la Prata cratons, constituting the main domains of the terranes that evolved during the Rio Doce Orogeny (Campos Neto and Figueiredo, 1995). This scenario was followed by important magmatism (600 - 580 Ma). The foreland-type basins, with sedimentation around *560 ± 20 Ma and deformation around 530 Ma, represent the* main volcanic-sedimentary record of the Rio Doce Orogeny in southern Brazil.

Figure 1 shows the distribution of the two main tectonic domains of southern Brazil and Uruguay where the Luís Alves Microplate and Rio de la Plata Craton stand out in the western part, whereas the Dom Feliciano Belt (DFB) predominates in the eastern part. Great emphasis is given to the internal DFB segmentation, pointing out the tectonic contacts between the several DFB segments. Thus, the Major Gercino (SC) separates the Granite Belt from the supracrustal rocks, Cordilheira (RS) and partly the Sierra Ballena (UY) shear zones. Similarly, the contact between the Schist Belt and the foreland basins is also tectonic, prevailing low-angle mylonitic belts very well developed in the State of Santa Catharina (SC), where the Brusque Group metamorphites overlie the Itajaí Group sedimentary rocks.

Dom Feliciano Belt

Hasui *et al.* (1975) arranged the supracrustal rocks of southern Brazil and Uruguay in two major groups represented by the Tijucas (SC, RS and UY) and Rocha (UY) fold belts; these segments are separated by median massifs, formed by older granite-gneiss terranes reworked in the Neoproterozoic. Fragoso Cesar (1980) proposed for southern Brazil and Uruguay the Dom Feliciano Belt, separating it from the Ribeira Belt. An important geological synthesis for the study area was presented by Almeida *et al.* (1981), who placed in the Mantiqueira Province the eastern Brazil crystalline terranes to the S of the São Francisco Craton.

In Fragoso Cesar's (1980) original proposal, the Dom Feliciano Belt resulted from subduction involving a W-dipping oceanic plate (in agreement with Porada's 1979 proposal), producing a magmatic arc during the Neoproterozoic, the granitoid plutons of the eastern domain; a series of supracrustal rocks in a back-arc basin setting; and a molasse basin in its external portion, filled with sediments and anchimetamorphic volcanic rocks to a great extent installed on the eastern border of the Rio de La Plata Craton.

Fragoso Cesar's (1980) model has been improved by changes resulting from detailed studies carried out in different DFB segments, such as Basei (1985); Silva and Dias (1981); Silva *et al.* (1985); Soliani (1986); Fernandes *et al.* (1995a, b); Hartmann *et al.* (1998); Bossi and Campal (1992); *Preciozzi et al. (1985); Sanches Bettuci (1998).* The main synthesis of the gravimetric data can be found in Schukowsky *et al.* (1991); Halinann and Mantovani (1993); Quintas (1994).

Figures 2, 3 and 4 show three NW-SE geological sections at different latitudes along the DFB. The spatial relationships between the DFB segments are pointed out, stressing the

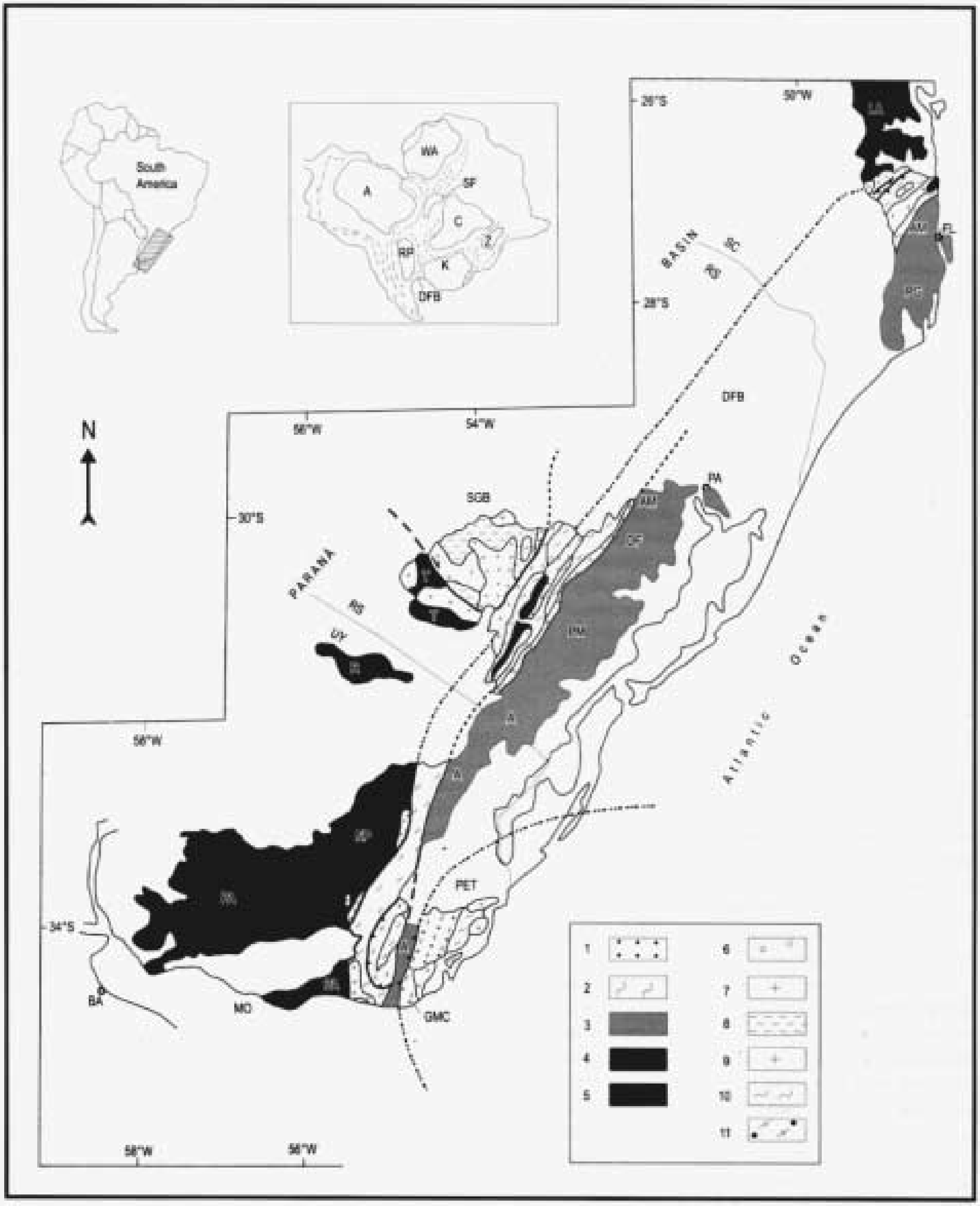


FIGURE 1 - Geological map of southeastern Brazil and Uruguay.

Dom Feliciano Belt (DFB), 1 - Foreland basins: Itajaí, Camaquã, El-Soldado-Piriapolis; 2 - Schist belts and intrusive granitoid: Brusque Metamorphic Complex, Porongos Metamorphic Complex, Lavalleja Metamorphic Complex; 3 - Granite belt: Florianópolis Batholith (AM, PG), Pelotas Batholith (AM, PG, PM), Aiguá Batholith (A); 4 - Basement inliers: Morro do Boi, Encantadas, Punta Rasa; 5 - Foreland, internally preserved of Neoproterozoic overprint: Luis Alves Microplate (LA) and Piedra Alta Terrane (PA); affected by Neoproterozoic heating and granitogenesis: Taquarembó (T), Rivera (R) and Nico Perez (NP); São Gabriel Block (SGB), 6 - Foreland basins (Maricá and Santa Bárbara); 7 - Intrusive granitoid (São Sepé, Caçapava and São Gabriel); 8 - Metamorphic rocks (Cambaí and Vacacaí gneiss); Punta del Este Terrane (PET), 9 - Intrusive granitoid (Santa Tereza and San Ignacio); 10 - Metasedimentary cover (Rocha Group); 11 - Basement (orthogneiss with metasedimentary enclaves). Geographic references: SC - Santa Catarina State; RS - Rio Grande do Sul State; UY - Uruguay; FL-Florianópolis; PA - Porto Alegre; MO - Montevideo; BA - Buenos Aires types of contacts observed between these segments and the internal framework of each domain. Common relationships are the tectonic contacts between the different DFB segments, as well as their constancy along the *c.* 1200 km-long belt. Despite regional differences, it is noted that the DFB framework is characterized by the marked tectonic vergence from E to W with transport of all its units against the foreland. This deformation, despite the constancy of the transport direction, reflects different tectonic pulses that took place between 760 and 530 Ma, with the last phases reflecting compression associated with the approach of the Granite Belt.

Granite Belt

It represents the innermost part of the DFB petrotectonic domain, formed by an essentially igneous complex, with paraderived rocks occurring as roof pendants among the granitoid plutons. It is represented by three great segments occupying the eastern parts of the Brazilian states of Santa Catarina (SC) and Rio Grande do Sul (RS) and Uruguay (UY). From N to S these are the Florianópolis (SC), Pelotas (RS) and Aiguá (UY) batholiths, the relationships of which are obliterated by a Phanerozoic cover of the Paraná Basin.

The granitoid rocks of these batholiths were considered by several authors (Porada, 1989; Fragoso Cesar, 1980) as the roots of magmatic arcs produced by westward subduction of a Neoproterozoic oceanic crust. These are the supracrustal rocks that form the Schist Belt, the remaining materials of the corresponding back-arc basins. Alternatively, Basei and Hawkesworth (1993, 1994), based on geochemical and isotope data, suggested that these granitoid rocks formed in a tectonic context not related to the evolution of the Schist Belt. They proposed that the generation and following juxtaposition of these segments would have happened in a collisional context, only after the metamorphic peak of the supracrustal rocks that constitute the Schist Belt.

Based on the available ages for the different batholiths recognized along the Granite Belt, it is possible to suggest a trend of decreasing radiometric ages for the calc-alkaline rocks from N to S. For example, the tonalitic-granodioritic rocks of the Florianópolis Batholith are slightly older (*c.* 620 Ma) than the equivalent rocks of the Pelotas Batholith (*c.* 610 Ma) and both are older than the Aiguá Batholith (*c.* 580 Ma). On the other hand, older Rb/Sr ages (between 800 - 700 Ma) that may represent a magmatic event prior to the development of the Granite Belt, are more consistently found in its southern domain (Soliani Jr. 1986; Silva *et al.*, 1997).

Florianópolis Batholith

The north-northwestern limit of the Florianópolis Batholith is defined by the Major Gercino Shear Zone (MGSZ), which separates this batholith from the Schist Belt supracrustal rocks (Brusque Group). This transpressional shear zone that evolved polycyclically was developed in low temperature conditions, having a general NE framework, ductile-brittle characteristics and an important oblique component associated with the predominant dextral movement. Based on field relationships and on petrographic and geochemical characteristics, three major suites can be defined in the Florianópolis Batholith: Águas Mornas, São Pedro de Alcântara and Pedras Grandes.

Águas Mornas Suite

This suite consists of deformed granitoid plutons, there predominating migmatitic with granodioritic to monzogranitic leucosomes and more mafic mesosomes/ paleosomes. Basic rocks (amphibolite, gabbro, diorite) of varied dimensions also occur. Characteristic rocks are the granitoid plutons of the Santo Amaro da Imperatriz region (Águas Mornas Complex, Zanini *et al.*, 1997). Mafic enclaves are normally composed of tonalitic rocks with variable proportions of amphibole and biotite. The mineralogical composition is plagioclase (andesine), quartz, biotite, hornblende and potassic feldspar. In the leucosomatic, quartz-monzonite contains plagioclase, potassic feldspar, quartz, biotite and amphibole. U/Pb values for zircon and titanite indicate an age of 606 ± 12 Ma (Basei, 1985), showing that these granitoid rocks originated in the Neoproterozoic, and underwent mylonitic deformation soon after their formation.

Belonging to an ample series of deformed granitoid bodies, the rock-types of the Paulo Lopes Suite consist of very deformed, coarse to very coarse-grained, protomylonitic biotite granitoid, with augen-type microcline megacrysts in a foliated, coarse-grained, biotite-rich matrix. Foliated and non-foliated aplitic veins and N-S trending, metric quartz-porphyry dykes affect these rocks. The predominant foliation has attitude N10° - 30°E / 45° - 50°SE. Compositionally, they are protomylonitic biotite monzogranite with augen-type, centimetric, white-rosy feldspar megacrysts in a foliated matrix with marked recrystallization due to mylonitization that took place under low temperatures. Mineralogically, they consist mainly of oligoclase/andesine (30%), quartz (35%), biotite (20%), hornblende (10%) and microcline (5%). Titanite, apatite, zircon, epidote and chlorite occur in trace amounts. Similar granitoid rocks that underwent different degrees of alteration occur throughout the granitic domain.

Pb/Pb ages for zircon from the Paulo Lopes Granitoid plutons indicated an age of 642 ± 46 Ma (Silva *et al.*, 1997), whereas more precise U/Pb analyses by SHRIMP yielded a value of 629 ± 8 Ma (Silva, PhD thesis, in preparation), a better indication of the age of this suite.

São Pedro de Alcântara Suite

This suite is composed mainly of equi to inequigranular, medium-grained, medium-grey biotite granitoid that is slightly deformed and frequently presents schlieren structures. Mafic (amphibolitic or biotitic) enclaves are common, showing locally white leucogranitic bands with melanosomatic rims. They show lateral variations in composition from granodiorite to quartz-diorite granitoid. Different degrees of deformation are imprinted, predominating slightly foliated types with the following average mineralogical composition: oligoclase/andesine (40 - 50%), microcline (20 - 30%), quartz (25 - 30%) and biotite

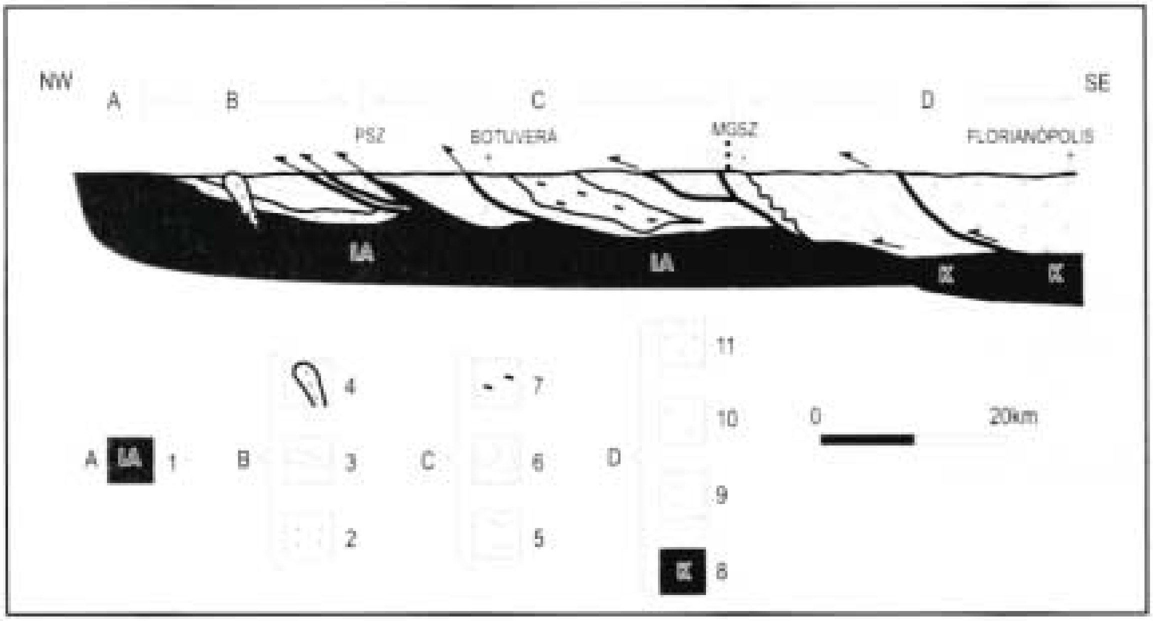


FIGURE 2 - Geological cross section throughout the northern part of DFB. A - Foreland, 1 - Luis Alves Microplate; B - Itajaí Foreland Basin, 2 - arkosic sandstone and conglomerate; 3 - marine turbiditic association; 4 - Subida Sienogranite; C - Brusque Schist Belt, 5 - mica schist; 6 - phyllite; 7 - Valsungana biotite-coarse-grained granitoid; D - Granite Belt, 8 - "African affinity" basement; 9 - Aguas Mornas Complex (calc-alkaline deformed granitoid); 10 - sienogranite related to the Major Gercino Shear Zone; 11 - Pedras Grandes sienogranite. MGSZ - Major Gercino Shear Zone; PSZ - Perimbó Shear Zone.

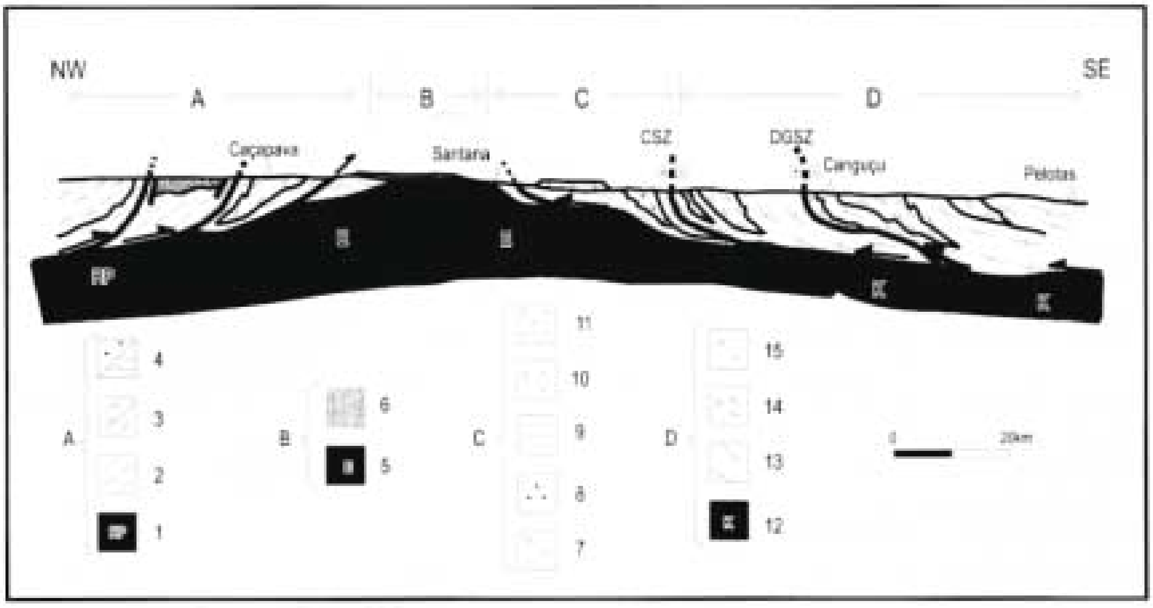
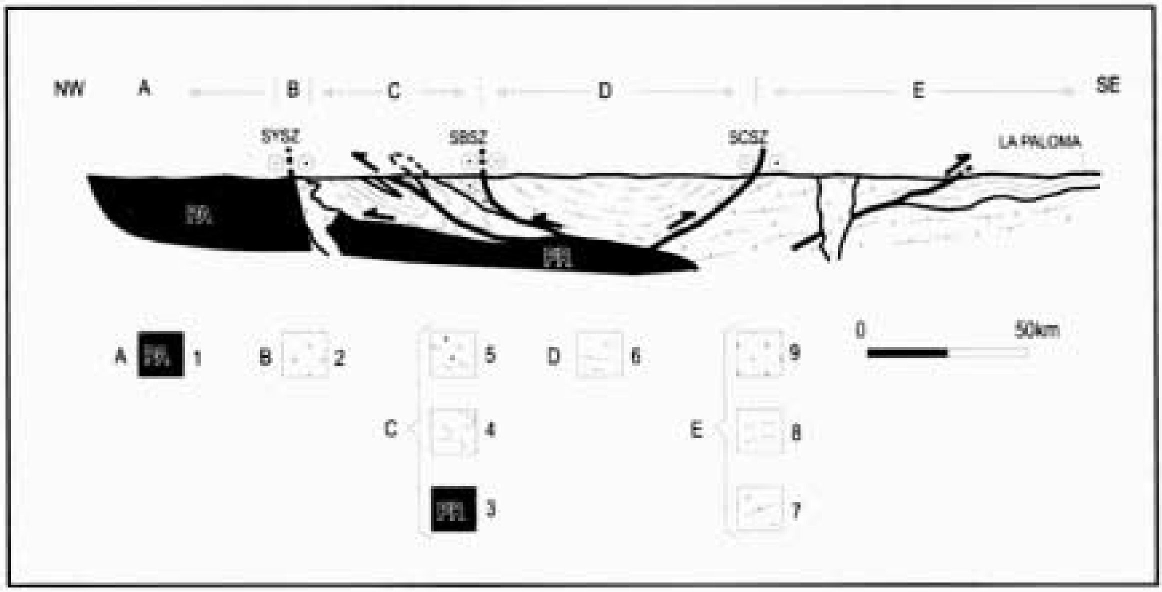


FIGURE 3 - Geological cross section through the São Gabriel Block and central part of the DFB. A - São Gabriel Block, 1 - Rio de La Plata Craton (basement); 2 - Vacacaí and Cambaí gneiss; 3 - São Sepé and Caçapava intrusive granite; 4 - Maricá and Santa Bárbara foreland basins; B - Basement inlier and cover, 5 - Encantadas gneiss; 6 - Early Paleozoic Guaritas Formation (Paraná Basin related sandstones); C - Porongos Schist belt, 7 - Cerro da Árvore metavolcano-sedimentary sequence; 8 - Encruzilhada gneiss; 9 - Serra dos Pereiras mica schist; 10 - Cordilheira Suite (peraluminous two-mica granite); 11 - Camaquã foreland basin; D - Granite Belt, 12 - "African affinity" basement; 13 - Pinheiro Machado Suite (deformed calc-alkaline granitoid); 14 - Viamão Suite (deformed, course grained biotite granitoid); 15 - Dom Feliciano Suite (non-deformed, leucocratic pink granite). DGSZ - Dorsal do Canguçu Shear Zone; CSZ - Cordilheira Shear Zone.

FIGURE 4 - Geological cross section throughout the southern part of the DFB and Punta del Este Terrane. A - Rio de La Plata Craton, 1 - Piedra Alta Terrane (Paleoproterozoic gneiss and low grade metavolcano-sedimentary belts); B - Foreland Domain, 2 - Sierra de Animas Suite (post-tectonic granite and volcanic rocks); C - Lavalleja Schist Belt, 3 - Punta Rasa gneiss (basement inlier); 4) - Metavolcano-sedimentary sequences (phyllite, marble and mica schist with basic and acid volcanic lenses); 5 - Carape Complex (isotropic and thrust-deformed granitoid); D - Granite Belt, 6 - Aiguá Batholith (deformed calc-alkaline to isotropic alkaline granite); E - Punta del Este Terrane, 7 - 1.0 Ga high grade gneiss migmatized during the Neoproterozoic-Cambrian transition; 8 - Rocha Group (low grade metasedimentary sequence); 9 - Post-tectonic granite. SCSZ - San Carlos Shear Zone; SBSZ - Sierra Ballena Shear Zone; SYSZ - Sarandy del Y Shear Zone.



(5 - 10%), generally with titanite, zircon, epidote and chlorite. Feldspar occurs as euhedral crystals with partially saussuritized, zoned plagioclase. The association of biotite with epidote indicates disequilibrium at low temperature; also suggested by granoblastic quartz with wavy extinction.

Occasionally dark-grey, medium to coarse-grained gabbro bodies with massive structure occur with pyroxene megacrysts reaching 1 cm. Mineralogically, they consist of pyroxene (50%), labradorite (45%), biotite+chlorite (5%) and traces of quartz, epidote and apatite. The plagioclase is inequigranular, locally saussuritized. The pyroxene is poikilitic with plagioclase inclusions. The chlorite-epidote association may indicate the role played by a late thermal event. The presence of biotite and quartz may suggest a basic-alkaline origin.

Two Rb/Sr isochrons were obtained for the São Pedro de Alcântara Suite, yielding 593 ± 24 Ma and 595 ± 11 Ma, respectively for the quartz-diorite and monzonite facies. Zircon extracted from the latter yielded an age 617 ± 38 Ma (Basei *et al.*, in preparation) in the Concordia diagram. Such values agree with the ages previously obtained by both methods. This magmatism is, from the geochemical and mineralogical point of view, the least evolved among the batholith granitoid rocks.

In the Santa Luzia region, a little farther N of Santo Amaro da Imperatriz, an isotropic, monzonitic, equi to inequigranular (white feldspar megacrysts up to 1.5 cm), medium-grey, biotite granitoid rock occurs. It is mainly composed of plagioclase, microcline, biotite and scarce quartz. Zircon, titanite, and epidote are abundant. This body should represent a São Pedro de Alcântara Suite facies. The value of 592 ± 15 Ma obtained from zircon and titanite (Basei, 1985) is a little younger, but concordant within the analytical error with the age obtained for the São Pedro de Alcântara Suite, maybe representing final stages of the emplacement of the granitoid rocks of this suite.

Pedras Grandes Suite

This suite consists of leucocratic, alaskitic, isotropic, red-rosy granitoid occurring as small stocks to large batholiths. They are the final expressions of granitic manifestations within the Florianópolis Batholith, being associated temporally and spatially with acid volcanic rocks. Locally, they contain xenoliths of deformed granitoid. The best examples of this unit occur in the southern part of this granitic domain, the best typical representatives being the granite of the Santa Catarina Island and Tabuleiro. Biotite monzogranite predominates with variations to syenogranite with alkaline affinities, composed of plagioclase (sodic oligoclase), potassic feldspar, quartz and biotite; zircon, allanite, apatite, opaque minerals and subordinately fluorite are accessories.

The best age presently available for this suite is 593 Ma, obtained from zircon by SHRIMP (Silva *et al.*, 1997). However, several Rb/Sr ages suggest the temporal continuity of this magmatism up to around 550 Ma (Basei, 1985; Soliani Jr., 1986; May, 1990).

Magmatism of the northern limit of the Florianópolis Batholith

The calc-alkaline magmatism associated with the Major Gercino Shear Zone is represented by the Rolador Granitoid (composed mainly of porphyritic to porphyroid monzogranite) and the Fernandes Syenogranite (coarse-grained, porphyroid amphibole-bearing rosy syenogranite), defined by Passarelli (1996). This lineament is very important to the framework of the Proterozoic terranes of southern Brazil, there being a discontinuity of lithosphere scale attributed to it (Basei *et al.*, 1992). It represents a suture between the Granite and Schist belts in the State of Santa Catarina.

Both associations make tectonic contact with the Schist Belt rocks (Brusque Group). However, they are petrographically and geochemically distinct from the granitic magmatism which occurs within the metasediments of this group. Isotope data for both associations present the same pattern observed in the Florianópolis Batholith, prevailing Mesoproterozoic (1.5 ± 0.2 Ga) Nd model ages and moderately negative ε_{Nd} values for 600 Ma (Mantovani *et al.*, 1987; Basei and Hawkesworth, 1993; Babinski *et al.*, 1997).

In general, they are metaluminous granitoid plutons that present elongated shapes with their more external parts involved in shearing processes. They are isotropic in the center, with development of magmatic or sub-magmatic flow structures. Geochronological studies using zircon extracted from granitoid of these two associations indicate an age of 644 ± 41 Ma for the Rolador Granitoid and 612 ± 19 Ma for the Fernandes Syenogranite (Passarelli *et al.*, 1997, recalculated with the inclusion of new fractions). Despite both associations are spatially related to the Major Gercino Shear Zone (MGSZ), it is probable that only the younger Fernandes Association would represent the magmatism genetically and structurally associated with MGSZ dextral development. This dextral development provided the structural control over the emplacement manifest in the elongated shape of the body, parallel to the shear zone.

In the Angelina region, at the western end of the batholith, an isotropic, rosy, equi to inequigranular hololeucocratic granitoid occurs, with potassic feldspar standing out from a medium-grained felsic matrix. Compositionally, it corresponds to a syenogranite with microcline (50%), quartz (35%), plagioclase (10%), muscovite (5%) and traces of biotite. It is possible that this granite corresponds to a southern extension of the associated Fernandes Granitoid the emplacement of which was conditioned by the development of the MGSZ. The Rb/Sr isochron age of 590 ± 12 Ma for the granitoid (Basei *et al.*, 1987) coincides with the value obtained by the same method for the Fernandes Granitoid rocks in the Major Gercino region (Passarelli *et al.*, 1997).

The zircon data available for the rocks that constitute the Florianópolis Batholith indicate that this unit was essentially formed in the Neoproterozoic III. The data also suggests that several granitoid generations, from rocks intensely deformed (considered as formed in the Archean or Paleoproterozoic) to late isotropic bodies, were generated during a maximum interval of 50 Ma. The oldest ages are about 640 Ma (maximum U/Pb age in zircon from calc-alkaline granitoid) and the youngest around 590 Ma (U/Pb in zircon from the late isotropic granitoid). It is possible that this interval is even smaller because the majority of and the most precise radiometric determinations tend to concentrate in the range of 625 and 590 Ma.

Pelotas Batholith

As observed in Santa Catarina, the eastern part of Rio Grande do Sul is mainly composed of Neoproterozoic igneous rocks. Several granitoid rock-types predominate, grouped in the Pelotas Batholith (Fragoso Cesar *et al.*, 1986). This 400 km long and 80 - 20 km wide batholith is composed of several suites of deformed to isotropic granitoid bodies that frequently present remnants of metamorphic rocks representative of their hosts. In a recent paper, Phillip (1998) summarized the geological data for the Pelotas Batholith presenting, besides petrographic and geochemical studies, a geological map where the main suites of the Rio Grande do Sul Coastal Granitoid Belt are represented. From the different suites proposed by Phillip (1998), only the Pinheiro Machado, Viamão and Dom Feliciano suites are considered in this paper as being part of the Pelotas Batholith.

Pinheiro Machado Suite

Represents the oldest and the main suite of the Pelotas Batholith, constituted by the most mafic lithologies found in this domain. Granodiorite associated with monzogranite, diorite and tonalite predominate, generally showing igneous flow defined by banding manifest in the form of schlieren-type structures as well as by the alignment of biotite and euhedral feldspar crystals. A metamorphic foliation is imprinted. This was generated by sub-horizontal shearing (Fernandes *et al.*, 1995 a, b; Phillip, 1998). The NE trending low-angle foliation contains stretched minerals (preferentially feldspar) with a conspicuous NW orientation. This deformation pattern is not observed in the other suites that constitute the batholith, exhibiting only the deformation developed in association with the high-angle shear zones. The granitoid rocks from the São Pedro de Alcântara and Santo Antônio suites may correspond, in the context of the Florianópolis Batholith, to the Pinheiro Machado Suite.

U/Pb ages in zircon (Babinsky *et al.*, 1997) are 610 ± 5 Ma (migmatitic gneiss of granodioritic composition) and 616 ± 2 Ma (biotite-rich migmatitic gneiss), values interpreted as indicative of the granitoid emplacement. Silva *et al.* (1997) presented the same interpretation for the 629 ± 8 Ma obtained by SHRIMP for the Pinheiro Machado Suite lithotypes. The authors also gave a 781 ± 5 Ma age for zircon extracted from an orthogneiss xenolith obtained from a granitoid from Pinheiro Machado Suite.

Viamão Suite

Several elongated bodies, concordant with the main NE-SW trending shear zones represent this suite. Monzogranite predominates over the granodiorite, locally there occurring syenogranite. These granitoid rocks are coarse-grained, porphyritic to equigranular, with texture defined by the alignment of potassic feldspar megacrysts and biotite, in general presenting a great quantity of mafic enclaves and significant occurrence of dioritic rocks. They may correspond, in the context of the Pelotas Batholith, to the Paulo Lopes-type granitoid discontinuously observed along the Florianópolis Batholith. The geochronological study for the Viamão Suite is precarious when compared with the Pinheiro Machado Suite. Values of 572 ± 22 Ma (Rb/Sr, Koester 1995) and 595 ± 1 Ma (U/Pb, Babinski *et al.*, 1997) confirm field observations that places this suite, in time, after the Pinheiro Machado Suite.

Dom Feliciano Suite

Defined by Tessari and Picada (1966), it represents the youngest magmatism in the Pelotas Batholith. Rosy leucogranite predominates. It is very homogeneous compositionally, structurally and petrographically with a marked lack of mafic enclaves and xenoliths of the host rocks. These granitoid rocks are isotropic or locally deformed under ductile-brittle conditions by late faulting. The relation of this magmatism with acid volcanic rocks and rhyolitic dykes is another point that stresses its relationship with the Pedras Grandes Granitoid of the Florianópolis Batholith. The bodies of this suite predominate in the northeastern part of the batholith

(Phillip, 1998) and their emplacement was not controlled by the main shear zones that affected the batholith. The lack of U/Pb ages for zircon makes difficult a more precise definition for the main period of this magmatism. However, it is likely that in like manner to the Pedras Grandes Suite in Santa Catarina, it is slightly older than the values indicated by the Rb/Sr isochrons, distributed between 570 and 550 Ma (Cordani *et al.*, 1974; Soliani Jr., 1986).

Considering that the Pelotas Batholith represents, in Rio Grande do Sul, the natural continuity of the Florianópolis Batholith occurring in Santa Catarina, it is here suggested that based on the similarities between them the western limit of the latter is defined by the Cordilheira Shear Zone.

This shear zone is associated with the alignment of homonymous granitoid plutons. It is thus proposed that the Cordilheira and Encruzilhada do Sul granitoid plutons do not belong to the Pelotas Batholith, but rather to the tectonic context of the Porongos Group, making contact with the metamorphites of the latter. The Encruzilhada do Sul Granitoid would in part correspond, in Santa Catarina, to the petrographically-similar bodies of the Valsungana Suite, having the same relations with the supracrustal host rocks. The southern limit of the Pelotas Batholith is represented by the tectonic alignment that separates the Pinheiro Machado and Erval suites (Phillip, 1998). It is also suggested here that the latter suite is not part of the Pelotas Batholith, but is related to the Aiguá Batholith that, in Uruguay, may represent another body correlated with other components of the Granite Belt.

In this paper the Pelotas Batholith is considered to be composed only of Phillip's (1998) Pinheiro Machado, Viamão and Dom Feliciano suites, the magmatic development of which resulted from the evolution of a single batholith. Nd isotopes confirm Sr isotope data that suggest the participation of the crust in the generation of these rocks, with predominant Mesoproterozoic model ages (Mantovani *et al.*, 1987; May, 1990; Babinski *et al.*, 1997). May (1990) presents two age intervals, 1.6 Ga for the Pinheiro Machado Suite granitoid and 1.2 Ga for the Dom Feliciano Suite granite.

ε_{Nd} values for the Pinheiro Machado Suite are distributed between - 4.4 and - 7.1 and for the Viamão Suite. They concentrate in the - 7.6 and 1.0 interval. More negative values, between - 8 and - 9 were obtained for gneissic xenoliths from the Pinheiro Machado Suite. The Dom Feliciano Suite presents similar behavior to that of the Viamão Suite with variation of the ε_{Nd} values between - 6.3 and 0.1.

According to the work of Mantovani *et al.* (1987), Sr and Nd isotopes for the Pelotas Batholith suggest derivation and magmatic evolution similar to that found in the Florianópolis Batholith in Santa Catarina (Basei, 1985; Basei and Hawkesworth, 1993). These results lend increased support to the hypothesis that places these two units in a single granitoid belt.

Aiguá Batholith

In like manner to southern Brazil, in the Uruguayan Shield all the region to the E of the Schist Belt (Lavalleja Group) is a domain of granitoid rocks of Neoproterozoic age that can be correlated with the Pelotas Batholith (Bossi and Campal, 1991; Bossi and Campal, 1992). Recently, Preciozzi *et al.* (1999a, b) have identified the Punta del Este Terrane in eastern Uruguay, in the region that had previously been assigned to the Aiguá Batholith. It consists of gneiss and migmatite (not recognized in either Brazil or Argentina) that show U/Pb ages around 1.0 Ga and that would be akin to the Namaqua terranes of southwestern Africa. The terranes that can be correlated with the granitoid rocks of the Florianópolis and Pelotas batholiths observed in Brazil would then here be constrained by the Lavalleja Belt and the Punta del Este Terrane.

As observed in the Florianópolis and Pelotas batholiths, in the Aiguá Batholith igneous rocks of different composition predominate where the main lithotypes are poly-intrusive, calc-alkaline granitoid, there occurring, subordinately, isotropic granitoid bodies of syenogranitic composition. This granitic belt is interpreted as representing the roots of a magmatic arc (Fragoso Cesar *et al.*, 1986).

Recently Preciozzi *et al.* (1999, in preparation) have presented U/Pb data for zircon obtained from two granitoid bodies of the Cañas region, both in the southern part of the batholith. These are rocks of monzogranitic composition with biotite and amphibole as the main mafic constituents in a quartz-rich matrix. White, sub to centimetric potassic feldspar megacrysts stand out in a medium to coarse-grained matrix. These are slightly or non-deformed rocks, intruded during the tardi-kinematic phase.

The ages 587 ± 16 Ma and 572.2 ± 2.5 Ma are attributed to the time of emplacement of these granitoid plutons, confirming the previous Neoproterozoic values inferred from the Rb/Sr method (Hart, 1966; Umpierre and Halpern, 1971; Fernandez and Preciozzi, 1974). These ages are younger than those representative of the Brasiliano Cycle (ages older than 600 Ma) available for the southern Brazil in similar terranes.

Additionally, Preciozzi *et al.* (1999, in preparation) present K/Ar results for the same samples from which zircon was extracted. The ages are *c.* 20 to 30 Ma younger than the U/Pb values, suggesting a long cooling period. These authors suggest that the mineralogical and geochemical characteristics of the majority of these granitoid plutons define an important crustal participation in their generation. This conclusion is confirmed by Sm/Nd data that show very negative ε_{Nd} between - 9.8 and - 12.6 and model ages (T_{DM}) between 2.1 and 1.4 Ga. Mesoproterozoic ages associated with ε_{Nd} negative values are a constant in the granitoid rocks of the Granite Belt of southern Brazil, confirming the correlation suggested in previous works.

Schist Belt

The Schist Belt contains supracrustal rocks distributed between the Granite Belt and the foreland basins. In this segment metasedimentary and metavolcano-sedimentary sequences predominate, occurring discontinuously along a narrow belt with average widths around 40 km. Three distinct metamorphic complexes are named, from N to S: Brusque (SC), Porongos (RS) and Lavalleja (UY). These are polydeformed sequences in which at least three fold phases are recognized, associated with a northwestwardly mass transport that evolved to a predominantly lateral movement. The preferential surface in the majority of the metamorphic rocks is transpositional S_2, accompanying the metamorphic peak. The regional metamorphism generated rocks of the greenschist facies and locally low-amphibolite facies.

The continuity of the areas of occurrence of the supracrustal sequences is suggested by the tectonic positioning of these segments, by their characteristics and metamorphic ages and also by the granitic magmatism intrusive in these supracrustal rocks. Despite these common points, some fundamental elements for a more effective correlation are still unknown, such as sedimentation age and detailed comparative studies between the respective lithostratigraphic columns of each belt. Any statement attesting to the continuity of these fold belts placing them in the context of the same sedimentary basin should be accepted with reservations.

The main critical point in any correlation refers to the sedimentation age of the units. Up to now consistent data to positioning any of them in time has not been obtained. All possess a Paleoproterozoic basement (2.3 - 2.0 Ga) and Neoproterozoic metamorphism and granite genesis (*c.* 760 - 600 Ma) that leaves an ample time interval for sedimentation.

Recent SHRIMP U/Pb data for zircon presented by Porcher *et al.* (1999) indicated for Rio Grande do Sul values between 783 - 766 Ma for the acid volcanism intercalated with the Porongos Complex. These authors interpret such ages as indicative of the time of the volcanic rock generation in the context of the opening of the Porongos paleobasin and consequently indicative of the time of the basin in-filling. This interpretation is questioned here. The alternative hypothesis proposes that the volcanism generated during the metamorphic climax that affected the Porongos Complex represents volcanism produced by melting of deep levels of the sedimentary pile and therefore is not representative of the time of deposition of the metasedimentary units. The peraluminous characteristic of the volcanism corroborates with the participation of upper crustal levels in the generation of this magmatism.

Therefore it is likely that a significant part of Brusque, Porongos and Lavalleja metavolcano-sedimentary sequences have been deposited in Mesoproterozoic pre-Brasiliano basins without any link with the generation and deformation of the granitic batholiths located to the E (Florianópolis, Pelotas and Aiguá).

Brusque Metamorphic Complex

In Santa Catarina the Brusque Group is composed of two metavolcano-sedimentary series separated by the Valsungana Batholith. In the northern series the sedimentary sequence begin with a pelitic-psammitic unit (garnet-rich mica quartz schist and quartzite) that grades to psammitic-pelitic units (homogeneous metarhythmite and sericite schist) overlain by a metavolcano-sedimentary unit (metamarlstone, calc-schist, metabasic and subordinately grey sericite schist; Basei, 1985, Caldasso *et al.*, 1988). The mafic rocks represent a syn-sedimentary basic magmatism with tholeiitic to alkaline affinities characterized as variolitic basalt where structures indicating liquid immiscibility are frequently observed (Basei 1985; Sander 1992). In the southern segment, the basal sequence is composed of a metavolcano-sedimentary unit possibly representing the rift phase of the Brusque paleobasin (Basei *et al.*, 1994). In this unit volcano-exhalative deposits are characteristically formed by a thick tourmalinite pile associated with meta-basalt, BIF units, orthoquartzite (chert?) and calc-silicate rock (Silva *et al.*, 1985; Basei *et al.*, 1994). This sequence is overlain in tectonic unconformity by a metasedimentary unit formed by a thick psammitic-pelitic pile where micaceous quartzite, quartz-sericite schist, pelitic sericite schist predominate. Acid metavolcanic rocks occur locally.

The Brusque Group magmatism consists of a series of slightly deformed, isotropic granitoid plutons, with metaluminous affinity, indicating strong crustal contribution. They can be grouped in three main series, all of them showing tardi-tectonic characteristics in relation to the main phases of metamorphism and deformation of the host supracrustal rocks. The oldest São João Batista Suite consists of two-mica leucogranite locally with garnet, tourmaline and fluorite, occurring as small stocks and lode bodies. Known W and Sn occurrences are related to these granitoid bodies. In an intermediate position, there is the Valsungana Suite composed of whitish-grey granitoid typically with centimetric microcline megacrysts in a coarse-grained biotite-rich matrix. In the Nova Trento Suite light grey to slightly rosy biotite granitoid occurs late in relation to the previous series. All the granitoid plutons develop a clear contact metamorphic rim reaching the pyroxene-hornfels facies (Basei, 1985; Caldasso *et al.*, 1988, 1995a, b; Castro *et al.*, 1999).

Porongos Metamorphic Complex

In Rio Grande do Sul part the Schist Belt is represented by the Porongos Complex with the exposures in the vicinity of Santana da Boa Vista, associated with the Santana-Canapé Dome and the Cerro da Árvore Nappe. In the nucleus of an antiformal structure that characterizes the complex there occurs the Encantadas Gneiss (basement of the Porongos Group), tectonically covered by the metamophites of the Cerro dos Madeiras Group (Jost, 1982). At the base a psammitic metasedimentary unit is composed mainly of meta-arkose and impure quartzite intercalated within metapelite and rare amphibolitic gneiss (Arroio dos Neves Formation). In an intermediate position there occur quartzitic metarhythmite beds (Olaria Formation). A pile *c.* 2000 m-thick of mica schist with marble and orthoquartzite intercalations occur at the top of the sequence (Irapuazinho Formation). The Cerro da Árvore Nappe consists mainly of a metavolcanic sequence with meta-andesite, metadacite and a variety of pyroclastic rocks. Subordinately, metachert, marble, metapelite, graphite schist and rare quartzite are observed (Jost, 1982). The vergence is northwestwards, and the structure overrides the eastern flank of the Santana Dome.

The metamorphic conditions that affected these units range from the chlorite zone in the greenschist facies up to the staurolite zone in the amphibolite facies. The paragenesis suggests low-pressure metamorphism (between 2.0 and 2.2

kbar) for Cerro da Árvore and medium pressure metamorphism (3.5 to 4.8 kbar) for the Cerro dos Madeiras Group (Jost, 1982).

For the region of the Canapé Antiform, Marques *et al.* (1998a, b) emphasize the magmatism generated previously by the metamorphic and deformational climax of the Porongos Complex. This magmatism, according to these authors, has geochemical characteristics indicative of several tectonic environments varying from the rift stage (alkaline gneiss) to the subduction phase (calc-alkaline acid volcanic rocks) that would be associated with the pre-existing oceanic crust consumption.

It is here suggested that the granitoid and gneissic rocks of the region known as Encruzilhada do Sul Block belong to the context of the Porongos Complex. In disagreement with the proposals of Fragoso Cesar (1991), Phillip (1998) and Silva *et al.* (1997), they should be excluded from the Pelotas Batholith, from which the Cordilheira Shear Zone separates them. This region is characterized by igneous and sedimentary rocks with part of the gneissic rocks representing the basement (Encantadas-type) of the Porongos Complex (Porcher *et al.*, 1999), or deep levels (amphibolite facies) of the supracrustal rocks of the Cerro dos Madeiras Group (Silva *et al.*, 1997). This region, differently from what occurs in other areas underlain by the Porongos Complex, presents a larger number and greater diversity of granitoid rocks represented by the Encruzilhada do Sul Batholith as well as by several igneous rocks including the Piqueri Syenite, Capivarita Anorthosite and Pitangueiras Granitoid.

The Campinas Granitic Suite best represents the S-type granitic magmatism that affected the supracrustal rocks of the Porongos Complex. Small-circumscribed stocks distributed in the eastern part of the Schist Belt characterize these rocks. These are preferentially equigranular, isotropic two-mica leucogranites with peraluminous affinity, suggesting interaction with the upper crust (Frantz and Jost, 1983). The Rio Grande do Sul tin occurrences are associated with this magmatism.

Lavalleja Metamorphic Complex

In Uruguay, the Lavalleja Group represents the southern segment of the Schist Belt. This belt is formed by three supracrustal units, namely, from E to W, Zanja del Tigre, Fuente del Puma and Minas Formations (Sanches Bettuci, 1998), accompanied by a decrease in the metamorphic grade from E to W from low amphibolite, greenschist to very low (anchimetamorphic) grade in the northwestern region.

The Zanja del Tigre Formation corresponds to a metavolcano-sedimentary sequence consisting of gabbro and (para and orthoderived) amphibolite hosted by mica schist, garnet-rich schist and a variety of types of marble. It is overlain by psamitic-pelitic (metaconglomerate, calcarenite, calc-dolomite and mica schist) metavolcano-sedimentary sequence (Fuente del Puma Formation) where, in relation to the basal unit, there is a clear increase of the volcanic contribution (gabbro, basalt, volcanic breccia and rhyolite). The upper Minas Formation consists exclusively of sedimentary rocks including metapelite, quartzite and arkose, besides limestone that include stromatolitic units.

Spatially associated with the Zanja del Tigre Formation, and with its western end overriding the Fuente del Puma Formation, there occur orthogneiss and migmatite of the Carapé Complex where deformed granitoid plutons of varied composition predominate with subordinately feldspatic mica schist and undifferentiated metabasic rocks. Towards the top, this complex grades from syn-tectonic muscovite and tourmaline-bearing pegmatite to anorogenic isotropic granitoid. Despite their tectonic contacts, Sanches Bettucci (1998) suggested that the Carapé Complex stratigraphically overlies the Lavalleja Group.

As observed in the Brusque Group in Santa Catarina, the granite genesis that affected the Lavalleja Group metamorphites is widespread and very diversified, from the compositional point of view. The largest bodies are the Maldonado Granitoid intrusive in the Fuente del Puma Formation, and the Penitente Granitoid intrusive in the Zanja del Tigre Formation.

Basement Nuclei within the Schist Belt

The Dom Feliciano Belt contains sporadic occurrences of gneiss-migmatite rock that comprise the basement of the metamorphic units of the Schist Belt supracrustal sequences. In Santa Catarina, a little S of the overthrusts that constitute the northern limit of the Brusque Complex, there occurs mylonitic and granulitic gneiss, partially covered by the sediments of the Paraná Basin that are tectonically overlain by Brusque metamorphites, yielding Pb/Pb ages of *c.* 2.4 Ga (Basei, 1985). Similar values were obtained by the Rb/Sr method for migmatite that occurs in the Camboriú region (Morro do Boi). In this case, the basement occurrences are found at the borders of the Valsungana Granitoid, which suggests that this occurrence represents mega-xenoliths from the basement, uplifted by the granitoid.

In Rio Grande do Sul State, the Schist Belt basement is well defined as several occurrences of gneissic rocks, the main lithotype of which is the Encantadas Orthogneiss that occupies the nucleus of the Santana Dome. Rb/Sr ages close to 2.2 - 2.1 Ga (Soliani Jr., 1986) were confirmed regionally by SHRIMP analyses in zircon (Porcher *et al.*, 1999; Leite *et al.*, 1999).

In Uruguay, as in the other parts of the Schist Belt, the basement of the Lavalleja Complex consists of gneissic rocks, best exposed in the locality called Punta Rasa (SE of Montevideo). These are finely foliated mylonitic garnet gneiss that indicate, by means of SHRIMP dating, ages around 2.1 Ga (U.G. Cordani, personal communication).

From the available data, and despite the fragmented nature of the information, it is possible to recognize the predominance of a Paleoproterozoic gneissic basement along the entire length of the Schist Belt.

Foreland basins

The foreland basins form a narrow belt parallel to the general DFB trend, segmented, in like manner to the other DFB units, into three main basins named, from N to S: Itajaí, Camaquã and El Soldado-Piriápolis. Their sedimentary characteristics suggest that they should have been interlinked during the Vendian-Cambrian boundary by a seaway open to the SW.

Itajaí Basin

The Itajaí Basin (Basei *et al.*, 1999a, b) occupies an area of more than 700 km^2 in the northeastern part of Santa Catarina, along the Itajaí River valley. It is elongated along N60°E and consists of a thick epiclastic sedimentary sequence with subordinate trachytic to rhyolitic volcanic rocks including pyroclastics.

Pioneer studies (Maack, 1947; Salamuni *et al.*, 1961) divided the Itajaí Series in two different units redefined by Silva and Dias (1981) as the basal Gaspar Formation, composed of psammitic sequences with minor conglomerate and volcanic rocks; and the Campo Alegre Formation, formed by pelitic and pelitic-psammitic rhythmite beds.

Basei (1985) and Basei *et al.* (1987) attributed a thickness of 7500 m to the Itajaí Group. They showed that it has been affected by two deformational phases also present in the fold belt formed by the Brusque Group, and interpreted the structure as monoclinal with NE vergence towards the granulite terrane. They proposed that it is composed of two main units, divided into four informal lithostratigraphic sub-units: (i) a lower psammitic unit, equivalent to the Gaspar Formation, with an arenaceous-conglomeratic sub-unit containing thick arkosic sandstone beds intercalated with lenses of polymictic conglomerate and volcanic tuff, overlain by a rhythmic sandstone-silt sub-unit with microconglomeratic layers; and (ii) an upper silty unit, with a silty-arenaceous sub-unit with predominant silt at the base, and a silty-pelitic sub-unit of homogeneous clay and siltstone, containing small lenses of coarser material at the top.

Several authors including Krebs *et al.* (1990), Appi and Cruz (1990); Rostirolla (1991) elaborated stratigraphic columns, consisting of a lower sequence and an upper transgressive unit represented by a condensed distal system of rhythmic slates. Four main sedimentary facies associations were recognized and classified according to their depositional environment: 1) - turbiditic associations; 1.1) sandstone-conglomerate density turbidites; 1.2) graded density turbidites; 1.3) classic medium density turbidites; 1.4) attenuated turbidites; 2) - basin associations 2.1) pelitic sediments; 2.2) subaqueous slides; 3) - Transitional associations; 3.1) coastal plain sandstone 4) - continental associations 4.1) ruditic alluvial fans; 4.2) intercalated alluvial sandstone).

Restricted thin (< 50 cm) levels or lenses of strongly recrystallized tuff are intercalated within the Itajaí basal formations. The tuff units are fine-grained light green coloured rocks composed of quartz and sericite. Predominantly acid volcanic and subvolcanic rocks are more abundant than the pyroclastic rocks, and are intercalated within the sediments. Clasts of acid volcanic rocks with diameters from 3 mm to 40 cm are observed in all conglomerate units. Subordinate basic and intermediate rocks also occur as late dykes.

Paim *et al.* (1998) indicated the occurrence of ichnofossils and fossils in low-density turbiditic levels of the lower part of the upper unit of the Itajaí Group. The fossil found is of the *Chancelloria* type representing a Cambrian taxon with worldwide occurrence restricted to the Lower to Middle Cambrian, which led these authors to attribute a maximum age of 540 Ma for the deposition of the Itajaí Group. Cambrian trace fossils corroborate with this suggestion.

However, considering that the rocks of the Itajaí Basin were affected by the Subida Granite and by several domes of rhyolitic rocks, Basei *et al.* (1999a, b) suggested ages around 560 Ma for this magmatism, and concluded that the deposition of the Itajaí Group must have begun before 560 Ma. As a consequence, and considering that *Chancelloria* restricted these rocks to the Cambrian, these authors suggested that the age of the Proterozoic-Phanerozoic boundary in southern Brazil was older than 540 Ma presently accepted by the international community.

Camaquã Basin

In this paper the segmentation recommended by Fragoso Cesar (1991) is adopted, which associates only the Camaquã Basin with the development of the DFB. The other Eopaleozoic basins of Rio Grande do Sul, the largest being Santa Bárbara, would have their evolution related to the São Gabriel Block that would represent another Neoproterozoic belt developed in the western part of the basement domain in Rio Grande do Sul.

The Camaquã Basin starts with the Arroio dos Nobres Formation that presents, in its basal part, a sequence of deltaic fans consisting of sandy-pelitic rhythmite beds which grade towards the top to conglomerate and conglomeratic sandstone. In this group, or intercalated with it, there is a thick rudite sequence including conglomerate beds of variable composition with intercalations of coarse-grained sandstone. The upper unit of the Arroio dos Nobres Formation consists mainly of sandstone of variable composition with subordinate rudite wedges.

The Arroio dos Nobres Formation is separated from the overlying uppermost unit of the Camaquã Basin, the Guaritas Group, by an angular unconformity. The Guaritas Group is a sequence of sub-horizontal continental deposits consisting of alluvial fan conglomerate and arkosean sandstone at the base, grading to eolian arkosean sandstone interfingered with breccia, conglomerate and fluvial sandstone at the margins of the basin. This group is overlain, in erosional contact, by deposits of conglomerate, sandstone and fluvial-deltaic pelite (Fragoso Cesar *et al.*, 1999).

A major volcanic manifestation is intercalated within the Guaritas sedimentary sequence. It consists of lenticular flows similar to pahoehoe-type lava (Lopes *et al.*, 1999). These rocks, with basic alkaline affinities, were dated by U/Pb SHRIMP (Remus *et al.*, 1999; Hartmann et *al.*, 1998) at 470 Ma, placing the Guaritas Group in the Ordovician.

In agreement with the suggestions to exclude the Guaritas Group from the context of the foreland basins in the Proterozoic-Phanerozoic transition and associating it with previous rift phases of the Paraná Basin (Fragoso Cesar *et al.*, 1999; Lopes *et al.*, 1999) it is here pointed out that this unit covers, unconformably, those units assigned to the

Santa Bárbara and Camaquã basins. Each of these units is associated with one or two Neoproterozoic belts known in the Precambrian of Rio Grande do Sul.

Arroyo del Soldado-Piriápolis Basins

Differently from what is observed in the states of Santa Catarina and Rio Grande do Sul, in Uruguay there is not a single basin that can be referred to as typical of the transition phase between the metamorphic-deformational climax of the adjacent metasedimentary belts and the installation of large Paleozoic intracratonic basins. On the other hand, there are several occurrences that fulfill the above-mentioned role.

The Arroyo del Soldado Group is the most significant of all basin sequences deposited in the Proterozoic-Phanerozoic transition in Uruguay. The sequence is about 5000 m thick (Gaucher and Sprechmann, 1998), and consists of shallow marine deposits in angular unconformity on a metasedimentary basement of probable Mesoproterozoic age (Gaucher *et al.*, 1996). The diversity of its fossil content permits its positioning between the end of the Vendian and the Lower Cambrian. Among the fossils found *Cloudina riemkeae* is prevalent in the lower levels of the group, permitting correlation with the Jacadigo and Corumbá groups in western Brazil and the Nama Group in southwestern Africa.

The Arroyo del Soldado Group is about 250 m thick and consists of four formations that from base to top correspond to a sandy-pelitic sequence at the base that grades into carbonate beds of probable biomicritic origin, covered by a pelitic sequence rich in organic matter, banded iron formation units and chert. The BIF units corroborate with the suggestion of a possible worldwide glaciation that occurred around 550 Ma, close to the Precambrian-Cambrian boundary. Overlying these rocks there lies a thick sequence of mature quartzose to arkosean sandstone beds. In the upper part, carbonate sediments once again predominate. These include beds of oolitic calcarenite with intercalations of micritic limestone becoming stromatolitic towards the top.

An E-W compression caused the development of open folds in the Arroyo del Soldado Group, with subhorizontal axes trending from N15°W to N20°E. The folds generated are isopachous and do not present axial plane schistosity (Gaucher *et al.*, 1996).

Masquelin and Sanches Bettuci (1993) described another important occurrence that can be temporally correlated in detail with the Arroyo del Soldado Group in the Piriápolis region (E of Montevideo). This is the Playa Hermosa sub-basin, the most conspicuous of several deposits observed in this region that may contain some 3700 m of sediments consisting mainly of siliciclastic beds including graded sandstone, rhythmic sandstone (alternation of sandstone and pelite) and an important contribution of zones of matrix-supported polymictic conglomerate. Towards the top, cross-bedded sandstone units tend to predominate with intercalations of conglomerate, frequently with centimetric clasts of volcanic rocks and microsyenite. However, the presence of ignimbrite and trachyte (probably associated with the Sierra de Animas Formation magmatism), intrusive in the sedimentary sequence, and locally with centimetric meta-arenite xenoliths, suggests a contemporaneity between the volcanism and the upper part of the sedimentary sequence.

The Playa Hermosa sedimentary sequence was deposited after the metamorphic-depositional climax of the Lavalleja Complex, only presenting brittle deformation manifest by tilted beds dipping NW, N20° - 60°E, thrust faults and cataclastic zones. Locally, as in the Cañada Azucarera region, a S_1 cleavage is developed in axial zones of folds with N30°E-trending axes (Masquelin and Sanches Bettucci, 1993).

Foreland domain

The crystalline terranes observed in the northwestern and western regions of the Dom Feliciano Belt mainly consist of gneiss-migmatite rocks that served as the foreland domain for the belt development during the Neoproterozoic-Cambrian. These continental masses, even those for which the term craton is applied, represent pre-existing continental fragments totally involved by the processes associated with ocean closing between Africa and South America during the formation of the Gondwana Supercontinent. The destruction of the oceanic crust has as expressive counterpart in generation of the Granite Belt and in the development of the Kaoko-Gariep fold belts.

As observed in Figure 1, the foreland terranes can be divided into two main groups: the Luís Alves Microplate in the northern region, and the Rio de la Plata Craton in the central-southern region, partially covered by the Paleozoic sediments of the Paraná Basin. Geological, geochronological and geophysical features do not favor a possible continuity between these two units, being likely that they represent two independent crustal fragments.

Luís Alves and Curitiba Microplates

The terranes that constitute the basement of the northern part of southern Brazil, from Apiaí (Ribeira) to Brusque (Dom Feliciano) fold belts, can be grouped in two distinct geotectonic units: Luís Alves and Curitiba microplates (Basei *et al.*, 1992). The NE trending Pien Suture Zone, extending along the southern limit of a Neoproterozoic calc-alkaline granitoid belt separates these domains. This calc-alkaline granitoid belt occurs along the southern border of the Curitiba Microplate, as a consequence of the northwestward subduction of the oceanic crust between these geotectonic units (Machiavelli *et al.*, 1993; Harara, 1996). The microplates in question are shown in Figure 5 that indicates the continuation to the N of the geological map presented in Figure 1.

An important landmark in the evolution of the Luís Alves and Curitiba domains is represented by expressive non-deformed granite genesis of alkaline-peralkaline nature (Serra do Mar Suite) as well as by intense volcanism, related to the development of several extensional basins (Corupá, Campo Alegre and Guaratubinha). The main time interval (600 ± 10 Ma) of the formation of these rocks is restricted

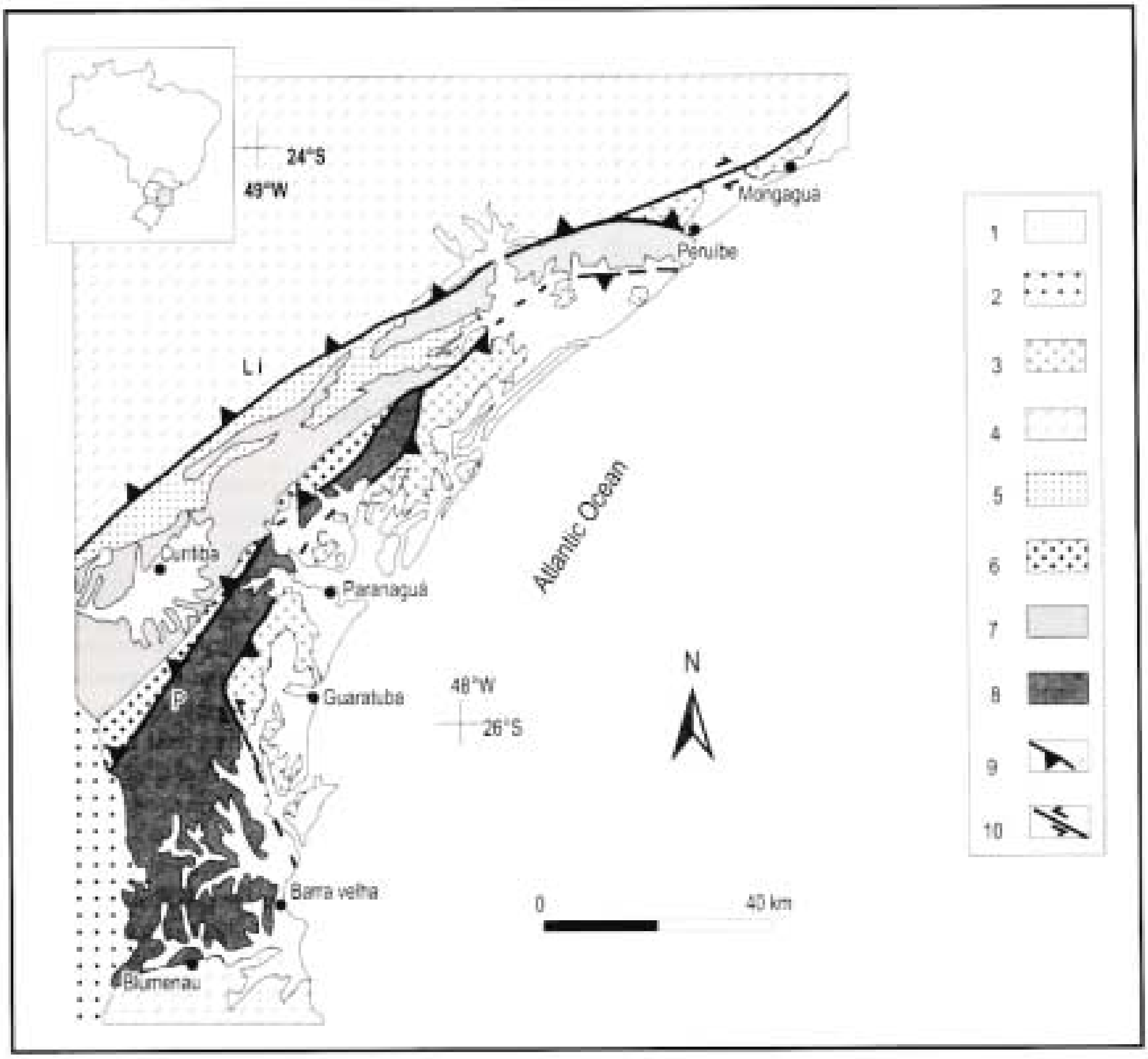


FIGURE 5 - Simplified geological map of the northern part of southern Brazil. 1 - Quaternary sedimentary cover; 2 - Paraná Basin; 3 - Costeiro Granite Belt (c. 620 Ma Paranaguá Batholith); 4 - Neoproterozoic fold belts (Ribeira Apiaí-N and Dom Feliciano-S); Curitiba Microplate. 5 - Capiru/Setuva metasedimentary sequences; 6 - Rio Pien Batholith (arc-related granitoid); 7 - Atuba Complex; Luis Alves Microplate. 8 - Santa Catarina Granulite Complex; 9 - Major Neoproterozoic Suture Zones, LI - Lancinha/Itariri, P - Pien, C - Costeiro, 10 - Trancurrent Shear Zones. Post-orogenic granitoid and basins were not represented.

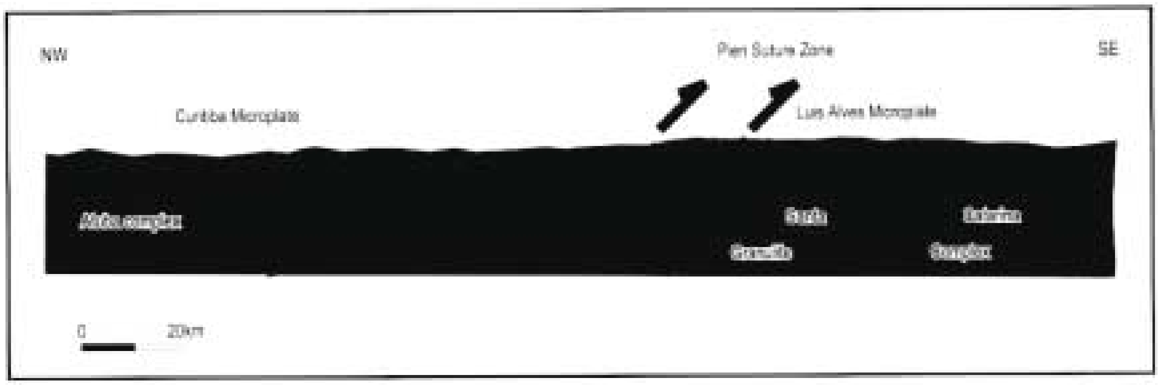


FIGURE 6 - Pien Suture Zone: geological cross section throughout the collisional limit between Curitiba and Luis Alves microplates. Mafic-ultramafic rocks (black pods) are interpreted as remnants of the Neoproterozoic oceanic crust. The age of the collision process can be estimated around 600 Ma, based on the ages of the post-collision intrusive granitoid and the Campo Alegre volcano-sedimentary basin (dated by zircon c. 595 Ma) and the deformed calc-alkaline I-type magmatic epidote bearing granitoid of the Rio Pien Batholith (c. 615 Ma).

to the final stages of the Brasiliano Orogeny and the beginning of Rio Doce Orogenesis. These granitoid plutons and basins are shown only in the geological profile (Fig. 6).

Curitiba Microplate

In its northern part the Curitiba Microplate is overlain by thick metasedimentary sequences (Capiru and Setuva formations), which have paleogeographic affinities with this microplate. These metasediments were not deposited in the same basin of the metasediments of the Apiaí (Ribeira) Fold Belt (Votuverava and akin), which occurs exclusively to the N of the Lancinha Shear Zone. The Lancinha-Itariri Shear Zone would represent, at the surface, the trace of an important suture that separates the Curitiba Microplate from the terranes situated to the N of it. The gneiss-migmatite rocks that occur in this microplate were called the Atuba Complex (Siga Jr. *et al.*, 1995) being mainly composed of banded migmatitic amphibole gneiss of Paleoproterozoic age, which underwent intense deformation and migmatization during the Neoproterozoic. Within the gneiss, which is regionally metamorphosed in the amphibolite facies, nuclei of high-grade metamorphic rocks of predominantly charno-enderbitic composition may be observed. These rocks are preferentially distributed in the northern part of these terranes, the best exposures of which are found in the Peruíbe region (Itatins Massif) in the southeastern part of the State of São Paulo. The Paulo Leminski Quarry in the City of Curitiba, is another site of excellent exposure.

For the Atuba Complex (Curitiba Microplate) previous U/Pb data obtained from zircon of the gneiss-granulite rocks show ages of 2.1 Ga (Siga Jr., 1995; Siga Jr. *et al.*, 1995; Picanço *et al.*, 1998). An additional determination for the Paulo Leminski Quarry resulted in 2.105 ± 0.003 Ga (Basei *et al.*, 1999b), confirming previous data. This age was obtained for zircon extracted from charno-enderbite, locally pegmatoid, which occurs within tonalitic, medium to dark-grey granulitic gneiss. Whole-rock Rb/Sr isochrons and K/Ar cooling ages are always Neoproterozoic. On the other hand, Nd model ages for metamorphic rocks are predominantly Archean.

Luís Alves Microplate

This microplate is considered to be a crustal segment where high-grade rocks of regional expression form the Santa Catarina Granulitic Complex (Hartmann *et al.*, 1979). The Pien Suture Zone defines its north-northwestern limit. It extends a little below the City of Blumenau, where it is overlain by anchimetamorphic sediments of the Itajaí Foreland Basin (DFB). It is the only crustal block of the Brazilian southeastern sector that had already cooled (below the 300° C isotherm) before the Paleoproterozoic, not undergoing Neoproterozoic tectono-thermal superposition. The high-grade rocks are predominantly orthoderived with a subordinate contribution of kinzigitic gneiss, quartzite and calc-silicate gneiss, besides the local occurrence of BIF units.

Compositionally they are charno-enderbite, predominating depleted types that are very similar to the granulite of the Lewisian Complex of Scotland.

The presence of mafic-ultramafic rocks is a constant displaying different dimensions, shapes and generations. The Barra Velha Mafic-Ultramafic Complex (without peridotite) represents the major occurrence of these rocks. Recent studies involving 1:50 000 scale mapping, permitted the definition of the complex that consists of expressive areas underlain by biotite gneiss, amphibole gneiss and amphibolite-facies migmatite that do not show any trace or even relicts of high-grade paragenesis. Associated with these rocks and passing to migmatitic gneiss, there are an important magmatic phase represented by hololeucocratic, quartz-feldspatic pink granitoid rocks. These granitoid plutons, despite being strongly deformed, do not show the structural complexity normally found in the majority of the high-grade rocks.

For the Santa Catarina Granulite Complex, reworked older data, plus new age determinations, permitted the definition of an age of 2.35 ± 0.03 Ga as representative of the main period of high-grade metamorphism. These ages were obtained from the analysis of several populations of potato shape-type zircon (rounded, brown, and translucent). In the same rocks, a second population of zircon crystals was also identified, consisting of slightly rosy, transparent, biterminated prismatic crystals (2x1 to 3x1), which yielded ages between 2.3 and 2.2 Ga (Basei *et al.*, 1999b).

Still in the Luís Alves Domain, in the Ibirama region, dating of a deformed rosy leucogranite revealed an age of 2.012 ± 0.021 Ga (Basei *et al.*, 1999b). This age is interpreted as representing the time of the granitoid emplacement, and as it is closely associated with the migmatite and gneiss that lack high-grade paragenesis, this age indicates the time of amphibolite-facies metamorphism, which regionally affected the complex.

In Figure 5 the existence of two Paleoproterozoic high-grade belts is suggested. The first crops out along the northern part of the Atuba Complex (Curitiba Microplate) being represented by a discontinuous orthogranulitic belt (mangerite to charnockite). The second, with predominant distribution in the context of the Luís Alves Microplate, is the main constituent of the Santa Catarina Granulite Complex.

The agreement between U/Pb ages obtained for three different localities of the Curitiba Microplate allows us to state, confidently, that the Atuba Complex high-grade metamorphism occurred at 2.1 ± 0.01 Ga. The zircon of this domain, even with SHRIMP analyses, lack evidence of a preterit crustal history, suggesting (despite the Archean Nd model ages) that the generation and high-grade metamorphism (including deformation and migmatization) of these igneous rocks were practically simultaneous. Subsequently, in Neoproterozoic III, these rocks were affected, to different degrees, by deformation, migmatization and amphibolite-facies metamorphism. Therefore, the Atuba Complex is a Neoproterozoic unit that had a polycyclic evolution starting at the Paleoproterozoic.

The geological history of the Santa Catarina Granulite Complex differs greatly from that suggested by the zircon of the Curitiba Microplate. The highly discordant behavior of its zircon can be mainly attributed to crustal heritage that was not recognized in the Atuba Complex granulite. This suggests that, in this case, the magmatic and metamorphic episodes were separated by a conspicuous time gap. Another very important difference between these complexes refers to the lack of Neoproterozoic superposition in the terranes of the Luís Alves Microplate. The gneiss-migmatite rocks that constitute this domain regionally present K/Ar ages older than 1.7 Ga. The Santa Catarina Granulite Complex can be considered as a Paleoproterozoic complex that underwent polycyclic evolution, where the main metamorphic and deformational events occurred at 2.35 and 2.1 Ga.

The first granulite facies metamorphism of the Santa Catarina Granulite Complex (*c.* 2.35 Ga) precedes *c.* 250 Ma its equivalent in the Atuba Complex (*c.* 2.1 Ga). However, the fact that the ages around 2.1 Ga occur in a conspicuous manner in both microplates allows us to suppose that these entities could have formed a single block during the Paleoproterozoic that were separated (*c.* 800 Ma) and reunited again (*c.* 610 Ma) in the Neoproterozoic.

Rio de La Plata Craton

The Rio de La Plata Craton (Almeida *et al.*, 1973; Fragoso Cesar and Soliani Jr., 1984; Dalla Salda *et al.*, 1988), groups the majority of the pre-Neoproterozoic terranes of southeastern Brazil and Uruguay, as well as northeastern Argentina. It is a geotectonic unit mostly covered by the Paleozoic-Mesozoic sedimentary successions of the Paraná Basin that served as foreland for the development of the southern part of the DFB. Certain geological characteristics contribute to further subdivisions giving way to regional names. In Argentina, the occurrence of Precambrian-Eopaleozoic basement gneiss is rare, being the main exposure area restricted to the Tandilia region.

In this paper, the different segments that constitute the Rio de La Plata Craton are grouped in two main units: Taquarembó (RS) - Rivera (UY) - Nico Perez (UY), and Piedras Altas (UY). The main differences that allow such distinction are based on the fact that the first unit presents Archean Nd model ages and that it was differently affected by Neoproterozoic-Cambrian thermal-tectonic events characteristic of the DFB. On the other hand, the Piedras Altas Terrane represents a juvenile crustal segment of Paleoproterozoic age that had not been internally heated since 1.75 Ga ago.

Taquarembó-Rivera-Nico Perez Blocks

Due to geological and geochronological similarities the Taquarembó, Rivera and Nico Perez occurrences can be placed in a single block where granulitic rocks occur within a gneiss-migmatitic unit. The granulitic rocks predominate in the northern part of this segment (Taquarembó Block, RS) and occur subordinately in the Rivera-Valentines region (UY), whereas in the Nico Perez Terrane gneiss and migmatite of amphibolite facies predominate.

The old terranes of the western part of the Rio Grande do Sul Shield occur S of the Ibaré Shear Zone, which separates this segment from the Neoproterozoic São Gabriel Block. In the Taquarembó Block the Santa Maria Chico Complex best represents the granulitic rocks. In this Paleoproterozoic complex containing Archean protoliths (Hartmann et al., 1999) orthoderived bimodal rocks, depleted in lithophile elements, varying between acid to basic composition predominate. Subordinately, sillimanite gneiss, anorthosite and pyroxenites occur, metamorphosed in the granulite facies reaching pressures of 10 kbar and 800°C (Hartmann, 1998). Sedimentary cover, acid volcanics and post-tectonic, Neoproterozoic to Cambrian granitoid plutons are observed frequently in this gneissic domain (Soliani Jr., 1986).

Whole-rock Rb/Sr and Pb/Pb isochrons, distributed between 2.5 and 2.03 Ga and obtained from analyses of different metagranitoid bodies, mark the characteristic geochronological pattern for this segment of the basement. The few Nd model ages indicate Archean values with strongly negative $\varepsilon_{Nd(0)}$. In a recent paper, Hartmann *et al.* (1999) presented SHRIMP results for a metabasalt and a metatrondhjemite of the Santa Maria Chico Complex. Both rocks presented very similar radiometric patterns, with inherited zircon indicating 2.55 and 2.13 Ga as the age of the igneous protolith; and a smaller number of zircon crystals, considered as metamorphic by the authors, yielding 2.05 and 2.01 Ga, interpreted as representing of the age of the high-grade metamorphism that affected this complex.

In Uruguay, the Rivera and Nico Peres domains are geologically very similar to the Taquarembó domain with several gneiss and migmatite units occurring close to the granulitic rocks. Long mylonitic belts and relicts of supracrustal rocks of medium to high metamorphic grade, as well as granitic intrusions of different ages and composition attest the geological complexity of this crustal segment. In its eastern end the Nico-Perez Terrane presents a metavolcano-sedimentary unit (Pavas Formation, Preciozzi et al., 1985) composed of quartzite, conglomerate, fuchsite-bearing mica schist, marble and ortho-amphibolite, the metamorphism reaching the high amphibolite facies (garnet-sillimanite).

The long magmatic and metamorphic history of this domain can be built with the Rb/Sr ages for metagranitoid of the Corrales region (Cristalina de Rivera Island) that yielded values around 2.272 Ga in Rb/Sr isochrons (Soliani Jr., 1986; Cordani and Soliani Jr., 1990). This indicates that the Transamazonian Cycle (Paleoproterozoic) was responsible for the metamorphism of the basement gneiss. The 1.76 Ga old Illescas rapakivi granite (Bossi and Campal, 1992), as well as K/Ar ages of 1.253 Ga for neoformed muscovite from mylonite of its eastern border (Campal et al., 1995) strongly suggest that Mesoproterozoic thermal-tectonic events took place in the Nico Perez Terrane. Additionally, Neoproterozoic shear zones affected this domain manifest by the intrusion of porphyritic granite and granodiorite with Rb/Sr ages distributed between 689 and 664 Ma. Isotropic K-feldspar granites are much younger, around 580 Ma old. Initial $^{87}Sr/^{86}Sr$ ratios clearly show that these granitoid rocks did not originate from melting of basement gneiss (Soliani Jr., 1986). The lack of radiometric determinations in zircon, together with the Archean record of the Nd model ages, allows us to suggest a probable evolution from the reworking of Archean protoliths for the Rivera-Nico Perez Block.

In Argentina, the geological and radiometric information related to the basement of the Rio de La Plata

Craton is scarce and concentrated in the Serra de Tandil region, in the Province of Buenos Aires. In this area gneiss and migmatite occur with different metagranitoid bodies, including data for mylonitic anatexite, which present a Paleoproterozoic geochronological pattern. Rb/Sr isochron ages are distributed between 2.15 and 1.77 Ga, including grey gneiss to rosy leucogranite (Varela et al., 1988; Ramos et al., 1990; Dalla Salda et al., 1992). Acid and basic metavolcanic rocks (Lema and Cucchi, 1981; Teruggi et al., 1988), later to the main deformation, are recognized within the regional gneiss.

The sedimentary cover of the Tandil region corresponds to a group of NW-SE trending lithostratigraphic units that have received several local names (Tandil Group, La Tinta Formation, Loma Negra Formation, and Sierras Bayas Formation, among others). It consists of a non-metamorphosed siliciclastic to carbonate sequence. The interval between deposition and diagenesis is between 800 Ma (Pothe de Baldis et al., 1983; Poiré, 1993) and 720 Ma (Bonhomme and Cingolani, 1980) estimated, respectively, from microfossil, stromatolite associations and radiometric data.

Some basement granitoid presented K/Ar ages between 870 and 790 Ma (Linares, 1977) showing the Neoproterozoic superposition on this domain permitting its correlation with the other reactivated segments of the Rio de La Plata Craton. *However, it is interesting to note that despite being situated* close to the eastern border of this cratonic domain (which could suggest its relationship with the DFB magmatism), the ages obtained for this granitoid are very similar to those obtained for the magmatism of the São Gabriel Block, which is characteristic of the western part of Rio Grande do Sul, Brazil.

Piedra Alta Terrane

The Piedra Alta Terrane (PAT) occupies the southwestern end of the Uruguayan Shield being separated from the Nico Perez Terrane by the Sarandy del Y Shear Zone. It consists of a group of orthoderived rocks where gneiss and migmatite of different compositions predominate. Anatectic granitoid bodies are frequently observed (Cerro Colorado), generally associated with basic magma, magma mixing and mingling phenomena (Fernández and Preciozzi, 1974; Bossi and Navarro, 1988; Dalla Salda *et al.*, 1988; Bossi *et al.*, 1993).

Three E-W-trending segments of metavolcano-sedimentary rocks, folded and metamorphosed to varying degrees (Arroyo Grande, San Jose and Montevideo belts) can be observed within a gneiss-migmatite unit. As before, the term San José Belt, stretching out between Cerro de Sán Juan (W) and Cerro Fray Marcos (E), was proposed with the intention of naming several metamorphic units represented by the Paso Severino, San Jose and Cerros de San Juan formations. Variations along the trend, in the metamorphic grade, and in the lithostratigraphic organization are frequent. In the San Jose Formation carbonate rocks, sericitic phyllite, quartzite and important lenses of basic metavolcanic rocks are observed.

PAT shows no evidence of Neoproterozoic orogenies (deformation, metamorphism or granite genesis). The available Rb/Sr and K/Ar cooling ages are restrict to the 2.2 and 1.9 Ga interval (Hart 1966; Umpierre and Halpern 1971; Cingolani *et al.*, 1990; Cingolani *et al.*, 1997; Preciozzi and Bourne 1992), being grouped in three major intervals: 1.9 to 1.7 Ga, 2.2 to 1.9 Ga and up to 2.2 Ga. A non-deformed mafic dyke swarm is the last manifestation of magmatic activity in the PAT with K/Ar and Ar/Ar values around 1.78 ± 0.03 Ga (Bossi *et al.*, 1993).

Recent data including Rb/Sr (2.094 ± 0.028 Ga, Ro = 0.70174) and Sm/Nd (1.949 ± 0.024 Ga, Ro = 0.51003) whole-rock isochrons were interpreted (Preciozzi *et al.*, 1999b) as the age of metamorphic episode that can be recognized in the entire Rio de La Plata Craton. The same authors present for the Isla Mala Granodiorite (intrusive in the San José supracrustal rocks) a U/Pb zircon age of 2.088 ± 0.012 Ga from three fractions of prismatic, well-formed zircon crystals. This age is interpreted as related to zircon crystallization, close to the granodioritic intrusion. SHRIMP ages around 2.07 Ga obtained by Bossi *et al.* (1999) for two granitoid plutons, including the Isla Mala, confirmed the Paleoproterozoic pattern previously determined from other radiometric data.

Based on geochronological and isotope data it can be observed that the geological evolution of the PAT occurred in a time interval no longer than 300 Ma, between 2.4 and 2.3 Ga (main accretion of its protoliths) and 2.1 and 2.0 Ga (magmatism, metamorphism and deformation). This *makes the PAT the only juvenile Paleoproterozoic geotectonic* unit observed in southern Brazil and Uruguay. Additionally, this unit was not internally affected by the Brasiliano Cycle being already stable and cold around 1.7 Ga which clearly differentiates it from the other units that constitute the Rio de La Plata Craton.

São Gabriel Block

This block occurs exclusively in the northwestern part of the basement in the Rio Grande do Sul State. The São Gabriel Block (SGB) is the only tectonic domain formed by rocks generated in the Brasiliano Cycle in southern Brazil and Uruguay that do not belong to the Dom Feliciano Belt.

Soliani Jr. (1986) showed that, contrary to what was considered at the time, the majority of the gneissic rocks that constitute the block were not as old as supposed, but were formed in the Neoproterozoic during the Brasiliano Orogeny. These rocks may have developed in a marginal basin situated along the eastern border of the Rio de La Plata Craton. Subsequent studies were carried out by several authors such as Fragoso Cesar (1991); Babinski *et al.* (1996); Leite (1997); Leite and Hartmann (1997) and more recently by Remus *et al.* (1999). These studies demonstrated that a great part of the material forming orthoderived rocks was also incorporated in the crust in the Neoproterozoic. This fact differentiates these rocks from those that formed the terranes of the DFB that are characterized by crustal reworking.

The basement of the São Gabriel Block consists of orthoderived gneiss varying from diorite, tonalite to trondhjemite possibly generated in an island arc context (Leite and Hartmann, 1997). Within the gneiss, grouped in the Cambaí Complex, harzburgite, serpentinite and basalt represent remnants of ophiolitic sequences as well as intensely deformed dioritic rocks, observed locally.

The main deformation in the gneiss is marked by the development of a shear foliation associated with intense transposition and generation of a metamorphic banding in amphibolite facies conditions. Kinematic indicators suggest a main eastward transport of the whole set. This nappe-associated transport direction was defined by several studies such as Tommasi *et al.* (1994); Fragoso Cesar *et al.* (1998); Leite (1997).

Another important metamorphic unit is formed by volcano-sedimentary rocks of the Vacacaí Complex that include phyllite, pelitic schist, marble, quartzite, metaconglomerate, amphibolite and calc-silicate rocks, and subordinately metamorphosed tholeiitic to alkaline basalt and komatiitic magnesian schist (Bittencourt and Hartmann, 1984). The regional metamorphism that affected these rocks varies from low greenschist to medium amphibolite facies (Hartmann *et al.*, 1990).

U/Pb, Rb/Sr and K/Ar ages for the metamorphic rocks cluster between 750 and 620 Ma (Soliani Jr., 1986; Babinski *et al.*, 1995; Leite *et al.*, 1999). Detrital zircon extracted from albite mica schist from the Caçapava Granite yielded values that suggest an important source with age between 900 and 800 Ma. This age, older than that of the regional orthogneiss, was only found in the Passinho Diorite (Leite *et al.* 1999; Remus *et al.*, submitted; Hartmann *et al.*, 1998), representing intensely deformed, discontinuous igneous bodies recognized among other orthogneissic rocks.

Volcano-sedimentary basins of the Maricá and Santa Bárbara types are distributed along the whole block, unconformably resting on regional gneiss. Locally, they present a gentle deformation related to block tilting or open isopachous folding that can represent a fine foliation of very low metamorphic grade. Rhyolitic volcanic rocks of the Tupanci and Acampamento Velho types are abundant, with the Hilário volcanic rocks, which occur within the Santa Bárbara unit presenting shoshonitic characteristics.

Varied granitoid plutons with tardi to post-tectonic characteristics and alkaline affinities, such as Caçapava, Lavras do Sul and São Sepé granites intrude all other rocks, often producing contact metamorphic aureoles. The U/Pb ages by SHRIMP obtained at the rims of the zircon crystals from the Lavras and Caçapava granitoid rocks indicated, respectively, 580 and 541 Ma, representing their emplacement time (Leite, 1995). The Lavras Granite is intrusive in the Maricá Group volcano-sedimentary units (foreland basin/internal molasse).

The ages observed for the granitoid rocks with magmatic arc affinities, the geological context of the observed petro-tectonic associations, as well as the tectonic vergence opposite to the preferential DFB transport sense, support the suggestion of an distinctive evolution for the São Gabriel Block in relation to the DFB. The magmatic history of this crustal segment began 150 to 200 Ma before the beginning of the DFB granitic magmatism. The generation of the São Gabriel Terrane is linked to the consumption of an ocean that existed 900 - 800 Ma ago between the Rio de La Plata Craton and the gneiss-migmatite rocks of the Encantadas type, which represent the basement of the metavolcano-sedimentary sequences of the Schist Belt (Porongos Complex). During the Neoproterozoic III this region, as a consequence of the Granite Belt collision underwent intense tectonic reactivation. The section in Figure 4 illustrates the present situation.

Punta del Este Terrane

The Punta del Este Terrane (PET) was defined by Preciozzi *et al.* (1999a, b) as gneiss and migmatite formed in the 1.0 Ga - 900 Ma interval and that was intensely reworked during the Rio Doce Orogeny (600 - 500 Ma). Therefore, mainly based on U/Pb radiometric ages, these authors proposed that the terranes situated to the E of the Alféres-Cordilheira Shear Zone should not be included in the Granite Belt. This implies that this crustal segment represents, in South America, terranes related to the Namaqua Belt of the southwestern region of the African continent. Similar terranes have not been recognized in Brazil and Argentina.

In the context of PET, three major groups can be defined: a mainly orthoderived, gneiss-migmatite basement, a low-grade metasedimentary cover, the Rocha Group, and a post-tectonic series that includes alkaline granitoid and acid volcanic rocks. This compartmentation is shown in Figures 1 and 4.

The basement consists of gneiss, rich in quartz, biotite and andesine, also containing garnet and muscovite. They correspond, in general terms, to high-metamorphic grade rocks with paragenesis including almandine, quartz, cordierite, and sillimanite (Masquelin, 1990). Compositionally, tonalitic gneiss predominates, exhibiting intense NE-SW foliation. Mafic rocks and varied paraderived gneiss occur as enclaves and large roof pendants within the orthoderived rocks. Still in the basement domain, lenses of protomylonitic rocks can be found where ocellar granitoid and migmatitic areas stand out, being the predominant leucosome composed of biotite-muscovite leucogranitoid.

U/Pb ages in zircon from tonalitic granitoid indicate values between 1.0 Ga and 900 Ma that were interpreted as representing the generation time for these rocks. This must also have been the age of the high-grade metamorphism that affected most of the regional gneiss. On the other hand, anatectic mobilizates related to migmatite leucosomes yield Cambrian ages around 540 - 520 Ma (Preciozzi *et al.*, 1999a, b; Preciozzi *et al.*, in preparation), indicated that the (superposed) metamorphic conditions during the Rio Doce Orogeny reached at least the amphibolite facies.

PET metasedimentary cover mainly occurs in the proximity of the towns of La Paloma and Rocha, and consists of an essentially metasedimentary siliciclastic sequence represented by the Rocha Group. Despite polyphasic deformation and low to medium metamorphic grade affecting these turbidite beds, preserved primary structures are frequently observed. These structures include plane-parallel stratification, cross-bedding, hummocky stratification and thick zones with graded bedding (Fragoso Cesar *et al.*, 1987; Sanches Bettucci and Mezzano, 1993). Considering that the PET basement correlates with the Namaqua Complex gneiss, the Rocha Group is tentatively correlated with the Gariep Group supracrustal units.

The Cerro de Aguirre Formation (Campal and Gancio,

1993) represents a volcanic rock pile of intermediate to acid composition. The rocks are displayed as open folds with N30°-40°E axial orientation, where locally the development of a plane axial cleavage can occur. Several isotropic, circumscribed granitoid bodies represent the last important magmatic manifestations affecting the PET. The Santa Tereza represents this alkaline-affiliated granitoid and José Ignácio bodies. Rb/Sr ages given by isochrons combining minerals and whole-rock data, as well as ages obtained exclusively for minerals, show ages between 611 - 590 Ma and 550 - 537 Ma for the José Ignácio and Santa Tereza granites, respectively.

It is here suggested that the Gariep-Rocha Basin would not have undergone any significant degree of oceanic opening, although the local development of oceanic crust did occur (Marmora Terrane). Therefore, the main branch of the Adamastor Ocean (Hartnady *et al.*, 1985) was developed W of the Gariep-Rocha Belt and its consumption generated the Granite Belt. The closing of the Gariep-Rocha Basin and the following deformation of the supracrustal pile occurred around 545 Ma as a consequence of the eastward transport of the Granite Belt against the African terranes. At this time there occurred the obduction of the volcanic rocks (oceanic crust) of the Marmora Terrane over the metasedimentary units of the Port Nolloth Zone (Frimmel *et al.*, 1998). It is possible that the separation of the PET and its African equivalents occurred only during the break-up of Pangea and the opening of the Atlantic Ocean.

Juxtaposition of the DFB different units

As previously stated, the different segments of the Dom Feliciano Belt and the adjacent terranes have distinct geological, structural characteristics and ages. These differences are interpreted as resulting from geological histories specific to each segment.

The Schist Belt presents a polyphase, metamorphic-deformational evolution with the late deformational phases associated with re-folding of the metamorphic foliation that characterizes it. The main deformational phase is manifest by folds with NE axes and low-dipping SE axial planes, and development of a S_2 transpositional foliation. This framework was generated between 750 and 640 Ma during the Brasiliano Orogeny, and was associated with nappes with NW transport (Basei, 1985).

Most of the magmatic evolution of the Granite Belt took place between 620 and 590 Ma. The generation of the older calc-alkaline suites is associated with the subduction phase of part of the Adamastor Ocean and its deformation was generated with the collision between the Granite and Schist belts that occurred around 600 Ma. Soon after this collisional phase, the intrusion of late, isotropic, alkaline granitoid plutons related to the Pedras Grandes and Dom Feliciano suites took place, culminating around 590 Ma.

Reactivation of the main mylonitic transcurrent belts with sinistral characteristics and higher temperature in the southern part of the belt (RS and UY), and dextral characteristics and low temperature in the northern part (SC), took place as a result of important directional transpressional movement of the belt. The best estimation for this phase is represented by Ar/Ar ages around 534 ± 3 Ma (Phillip, 1998) for high-angle shear zones of the Canguçu and Pinheiro Machado regions (RS). In Santa Catarina, this process was older, as indicated by K/Ar ages for neoformed mica from the mylonitic rocks of the Major Gercino Shear Zone, with preferential values around 569 Ma (Passarelli, 1996).

As a result of the convergence of the Schist and Granite belts, the flexural basins were generated in the foreland, the diachronic filling of which took place from the Vendian in Uruguay to *c.* 560 Ma in Santa Catarina. The collisional phase deformation that began around 600 Ma in most regions of the Granite Belt, caused in the Schist Belt, reactivation and development of thrust faults and re-folding with SE-dipping axial planes, reaching the foreland basins only around 535 Ma. In these basins, this deformation consisted of thrust faults and gentle folds with NW vergence that locally developed plane-axial cleavage in the low greenschist facies.

In summary, the metamorphic and deformational history of the three DFB segments falls within the 750 and 530 Ma interval. Initially, in a not well established tectonic context, the main phase of deformation and metamorphism of the Schist Belt took place around 750 Ma, and at about 640 Ma it was transported against its foreland situated to the W (Luís Alves Microplate and Rio de La Plata Craton). Around 620 Ma, easterly subduction of the existing oceanic crust (Adamastor) and the generation of the Granite Belt took place in a magmatic arc context. Around 600 Ma the collision between the Granite Belt and the Schist Belt took place and the foreland basins started to form. Around 570 Ma an important mass transport occurred parallel to the direction of the belt and the last post-collisional granitoid plutons were generated. Close to 535 Ma the belt deformation reached the foreland basins and reactivated several high angle shear zones.

Correlations with the African counterpart

The interest in correlating terranes on opposite margins of the South Atlantic Ocean reflects a natural curiosity of both the researchers who work in the eastern South America and of those who study southwestern Africa. Porada (1989), Hartnady *et al.* (1985), Trompette (1994), Hanson *et al.* (1998), Hoffman (1999) are among the most recent works.

The major problem of past reconstruction stems from the fact that they focused on the direct correlation between the metasedimentary belts observed on both sides of the Atlantic. Therefore, several papers proposed, in our understanding incorrectly, emphasize the correlation between the South-African Kaoko/Gariep belts with their possible South-American Brusque/Porongos/Lavalleja equivalents. Despite presenting similar ages (Pan-African/Brasiliano), a closer examination, and mainly in the light of recent inflow of much radiometric data, these fold belts are seen as being significantly different. This rules out tectonic models that state that they formed a large and single sedimentary basin at the beginning of their geological histories.

Considering that the younger Granite Belt cannot be the magmatic arc associated with the development of the Dom Feliciano Schist Belt, the sediments of the probable basin associated with this arc would be lacking on the South American side. In an analogous way, on the African side, the rocks representative of the magmatic arc that must have existed during the development of the Kaoko/Gariep paleobasin are absent.

The tectonic model presented in this paper (Fig. 7), based on radiometric ages and isotope signatures, suggests that the magmatic arc active during the deposition of the Kaoko (Coastal Damara) / Gariep units in a possible back-arc basin, would be represented in the eastern parts of the South American Continent. It is here proposed that this magmatic arc developed in the 620 - 590 Ma period was generated from the subduction of an E-trending oceanic crust under the old terranes of the African side, with the DFB Granite Belt representing its roots.

The main argument suggesting that the plunge of the oceanic crust must have been to the E and not to the W as stated by previous workers is the Granite Belt Sm/Nd signature, which is totally different from the pattern observed in the other South-American tectonic domains. Model ages (T_{DM}) for the Granite Belt are systematically younger (between 1.7 - 1.2 Ga, clustering in the 1.6 - 1.4 Ga interval) than those for the Schist Belt units (granitic and metasedimentary rocks with ages around 2.0 Ga) and much younger than those found in the Luís Alves, or even Rio de La Plata-type basement units (> 2.1 and mainly clustering around 3.2 - 2.7 Ga).

The geochemical affinity of the arc magmatism may have been much higher than the corresponding material in the overriding plate (active margin), where the arc was installed, than in the material of the opposite plate (passive margin). It is here suggested that this should be the explanation for both the isotope differences between the arc and the terranes situated to the W, and the similarities between the Granite Belt and the African magmatism of southwestern Africa. Therefore, this suggests that the lithosphere above the subducting plate, melted for the generation of the Granite Belt, differs from the one that continues westwards, since the granitoid bodies of similar age have a different isotope signature.

Therefore, if on the South-American side model Nd ages between 1.6 and 1.2 Ga are characteristics of the Granite Belt, values in the same interval are common in several regions of the South African portion. Mesoproterozoic ages predominate for the rocks of the western part of Damara (region between the Walvis Bay/Karibib/Huab River) with the Palmental-type calc-alkaline granitoid and intra-plate syenite to granite presenting model ages between 1.5 - 1.1 Ga (McDermott 1986; McDermott and Hawkesworth, 1990). A very similar pattern was also found for metaluminous A-type granitoid rocks of the Damara central region (Jung *et al.*, 1998). Values in the same interval are equally common in the Neoproterozoic Mozambique Belt in Tanzania, where several metapelite and charnockite bodies occurring in the eastern part of the belt present Nd model ages between 1.5 and 1.1 Ga (Möller *et al.*, 1998). On the other hand, Archean ages (Limpopo Belt, xenoliths in Kaapvaal and some of the basement nuclei within the Damara metasediments) or Paleoproterozoic (Namaqua and xenoliths within the deformed granitoid of the Namaqua Belt) are characteristic of the rocks associated with the basement of the Neoproterozoic cover (Harris *et al.*, 1987; Reid 1997; Möller *et al.*,1998).

Part of the metasediments of the Damara (notably Rossing and Kuiseb) and Nama (mainly Kuibis and Schwarzrand) show Nd model ages similar to those obtained for the Granite Belt. This suggests that the Granite Belt must have been an important source for these metasediments, reinforcing the model of an evolution in a back-arc basin for the Kaoko and Gariep units and consequently of hinterland for the Nama.

Figure 8 shows a hypothetical section of the DFB and Kaoko/Gariep *c.* 500 Ma ago, soon after the building of western Gondwana. The double vergence of the thrusts should have resulted from the several collisions involved in the formation of both belts. The westward collisional tectonics involving the Granite Belt must have developed on the South-American side around 600 Ma. However, only after 545 Ma ago did the eastward thrusts placed these granitoid plutons on the supracrustal units of the African side in a synchronous manner to their metamorphic peak (Frimmel and Frank, 1998) and reactivated the structures of the South-American side deforming the foreland basins. On the African side, basins of similar ages (Nama) underwent deformation only after 506 Ma ago (Frimmel and Frank, 1998).

Final Remarks

As stated above, we do not accept a tectonic model involving northwestward subduction of an oceanic crust producing the Granite Belt and the Schist Belt, in back-arc conditions. When examined in detail such a model presents several problems that invalidate it, despite the fact that it represents a logical reading of the petrotectonic zoning of the DFB.

Within the general picture, all western parts of southeastern Brazil and eastern Uruguay are characterized by old domains, grouped in the Rio de La Plata Craton (RS and UY) and Luís Alves Microplate (SC), over which the main mass transport occurred, generating the Dom Feliciano Belt.

In the western region, the Luís Alves Microplate and the Piedras Altas Terrane (part of the Uruguayan Rio de La Plata Craton) were preserved from the Neoproterozoic tectonics. These tectonic episodes affected in a significant way the other old domains (Taquarembó Block, RS, and the Nico Perez-Valentine Terrane, Uruguay), causing general heating (rejuvenation of the K/Ar ages in biotite and amphibole) and generating several granitoid bodies.

In the Dom Feliciano Belt domain, the Schist Belt presents metamorphic ages around 120 Ma older than those obtained for the formation and emplacement of the older granitoid bodies recognized in the adjacent Granite Belt. Therefore, it is necessary to dissociate the main metamorphoic phase of the Schist Belt from the Granite Belt.

FIGURE 7 - Hypothetical sketch for the evolution of Kaoko/Gariep - Dom Feliciano belts from subduction (620 Ma) to final deformation (500 Ma) of late volcano-sedimentary basins.
A - Foreland (Luis Alves Microplate and Rio de La Plata Craton);
B - Foreland basins (Itajaí, Camaquã and El Soldado); C - South-American schist belts and intrusive granitoid (Brusque/Porongos/Lavalleja);
D - Granite Belt (Florianópolis, Pelotas and Aiguá arc-related batholiths);
E - South African schist belts and granitoid intrusives (Kaoko/Gariep + Rocha); F - Nama Basin;
G - Kalahari Craton. Filled triangles: thrust faults; open triangles: subduction zones; dashed line: inferred place where the opening of South Atlantic might had occurred. In the present tectonic interpretation it is considered that the basic volcanic rocks of tholeiitic affinity recognized in the Gariep Belt (Marmora Terrane) was not related to a large ocean but to a narrow internal sea developed during extension conditions occurred in a back-arc basin environment.

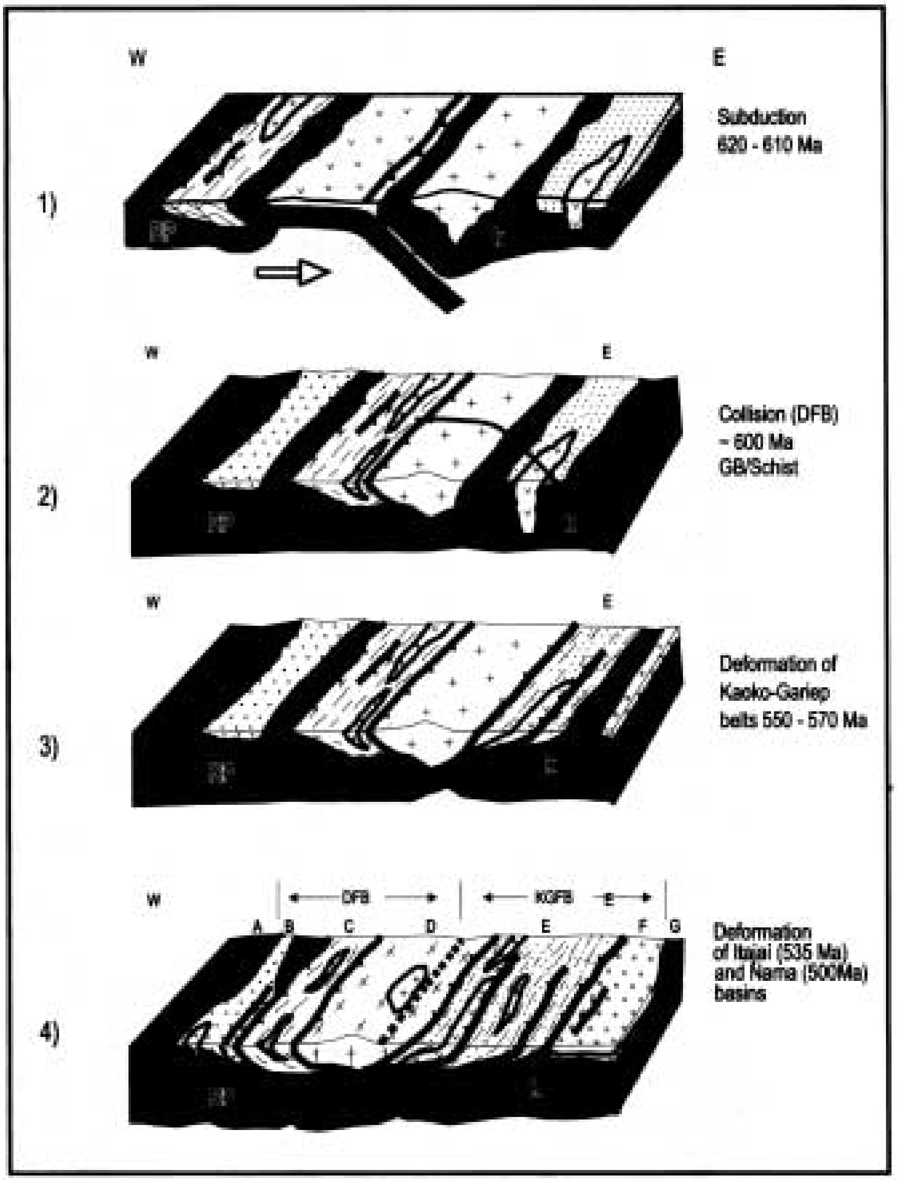
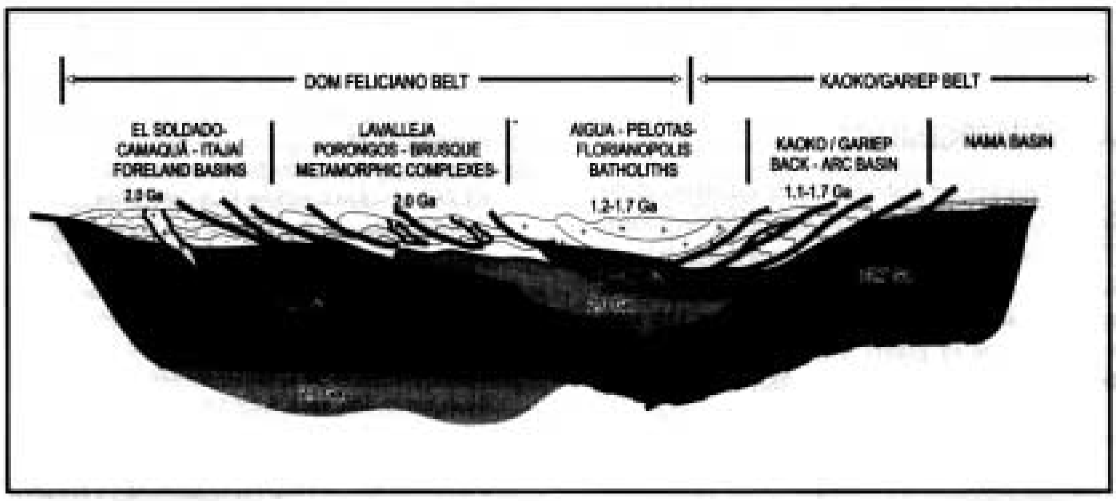


FIGURE 8 - 500 Ma schematic reconstruction of Kaoko/Gariep - Dom Feliciano belts. The proposed model considers the generation on the African side of a magmatic arc (Granite Belt) produced by the subduction of oceanic crust eastwards. The tectonic vergences were imposed on fold belts by their collisions with the Granite Belt. On the South American side the collision took place around 600 Ma and only c. 50 Ma later in southwestern Africa.
The 1.2 - 1.7 Ga Nd model age pattern recognized for the Granite Belt and its similarity with the Sm/Nd signature of the Neoproterozoic magmatism observed in southwestern Africa was a very important element in defining the model proposed.



At the time of the beginning of its evolution (*c.* 620 Ma) the supracrustal rocks had already reached their metamorphic climax and even undergone the diastrophism that placed them on their foreland by means of NW nappes that characterize the Schist Belt tardi-metamorphic deformation (*c.* 640 Ma). The biotite granitoid intrusives in the supracrustal rocks, despite presenting ages comparable with the Granite Belt magmatism, have petrological, geochemical and isotope characteristics that differentiate them from the latter. The development of an intense contact metamorphic aureole between these granitoid intrusives and the main foliation of the metamorphic rocks of the Schist belt is common.

The difficulty in dating precisely the time of sedimentation of the Schist Belt units and the confirmation of this as a function of the isotope signature has led several authors to classify the Schist Belt as a suspected terrane (Basei and Hawkesworth, 1993; Fragoso Cesar *et al.*, 1998). The provenance of the material that constitute the supracrustal rocks (at least for the state of Santa Catarina for which better information is available) does not seem to have been either old terranes situated to the W or to the younger rocks of the Granite Belt. Therefore, despite the Schist Belt being better known, the geological uncertainties still make a precise tectonic positioning of this crustal segment impossible.

The Granite Belt magmatic development (between 620 and 590 Ma) could be associated with the generation of a mature magmatic arc with important crustal participation, generated from eastward subduction, evolving in a distinct geographical position and in a dissociated manner from the terranes situated to the W. The probable back-arc basin for this belt could have been situated on the African side (part of the Coastal Damara/Kaoko/Gariep belts). In this context the Nama Basin would be a hinterland and not foreland basin, as suggested in previous works.

The geometric distribution and most of the structural features presently observed in the units that comprise the Dom Feliciano Belt would have evolved in the Proterozoic-Cambrian transition, after the collisional phase that juxtaposed the Granite Belt to the Schist Belt. The Foreland basins (Itajaí, Camaquã, Arroyo del Soldado-Piriápolis) were produced in a foreland in response to the approximation of the Granite Belt (600 - 560 Ma) and not, as has always been the consensus, generated during the deformational tectonics of the older (750 - 640 Ma) Schist Belt. Therefore, these basins must be understood as syn to tardi-collisional foreland basins of the Rio Doce Orogeny.

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REFERENCES

Almeida, F.F.M. de, Amaral, G., Cordani, U.G. and Kawashita, K. (1973). The precambrian evolution of the South American Craton margin south of the Amazonas river. In: *The ocean basins and margins*, eds. Nairn, A.E. and Stehli, F.G., pp. 1411-1446. New York: Plenum Press.

Almeida, F.F.M de, Hasui, Y., Brito Neves, B.B. de and Fuck, R.A. (1981). Brazilian Structural Provinces: an introduction. *Earth Sciences Review, Amsterdam*, **17**(1/2), 1-29

Appi C.J. and Cruz, C.E.S. (1990). Estratigrafia de seqüências na Bacia do Itajaí. In: *Congr. Bras. Geol., 36, Natal, 1990. Anais, Natal, SBG*, **1**, 93-106.

Babinski, M., Chemale, F., Hartmann, L.A., Van Schmus, W.R. and Silva, L.C. (1995). Acresção juvenil no bloco São Gabriel e retrabalhamento crustal no cinturão Dom Feliciano durante o ciclo Brasiliano: evidências isotópicas. *VI Simpósio Sul Brasileiro de Geologia, Porto Alegre, RS, Brazil, Extended Abstracts*, 105-107.

Babinski, M., Chemale, F., Hartmann, L.A., Van Schmus, W.R. and Silva, L.C. (1996). Juvenile Accretion at 750-700 Ma in Southern Brazil. Geology, **24**(5), 439-442.

Babinski, M., Chemale, F., Hartmann, L.A., Van Schmus, W.R. and Silva, L.C. (1997). U-Pb and Sm-Nd geochronology of the Neoproterozoic Granitic-Gneissic Dom Feliciano Belt, Southern Brazil. *Journal of South American Earth Sciences*, **10**(3-4), 263-274.

Basei, M.A.S. (1985). *O Cinturão Dom Feliciano em Santa Catarina.* Thesis submitted for the degree of Doctor of Philosophy, University of São Paulo, São Paulo, Brazil, 191 p. (unpublished).

Basei, M.A.S., Kawashita, K. and Siga Jr., O. (1987). Idade, características litoestratigráficas e estruturais do Grupo Itajaí, SC. In: *Simpósio Sul-Brasileiro de Geologia, 3, Curitiba, PR*, **1**, 93-106.

Basei, M.A.S., Siga Jr., O., Machiavelli, A. and Mancini, F. (1992) Evolução tectônica dos terrenos entre os cinturões Ribeira e Dom Feliciano. *Revista Brasileira de Geociências*, **22**(4), 216-221.

Basei, M.A.S. and Hawkesworth, C.J. (1993). O Magmatismo do Cinturão Dom Feliciano (SC) e sua importância no estabelecimento das principais descontinuidades crustais da região sul-brasileira. In: *Primer Simposio Internacional del Neoproterozoico-Cambrico de la Cuenca del Plata - La Paloma-Minas, Uruguai. Bol. Res. Ext.*, Tomo II, 41.

Basei, M.A.S. and Hawkesworth, C.J. (1994). The Granite Belt on the Eastern Part of Southern Brazil and Uruguay. In: *Eighth International Conference on Geochronology, Cosmochronology and Isotope Geology, Berkeley, USA. Abstracts, U.S. Geological Survey Circular*, **1107**, 22.

Basei, M.A.S., Campos Neto, M.C. and Siga Jr., O. (1994). Geologia do Grupo Brusque na Região de Canelinhas, SC. *Resumos do XXXVIII Congresso Brasileiro de Geologia, SBG, Camboriú-SC.*

Basei, M.A.S., Siga Jr., O., Cordani, U.G., Sato, K. and Lima, P.S. de (1999a). The Magmatism of the Itajaí Basin, SC, Southern Brazil, and its importance to define the Proteozoic - Phanerozoic limit. *II South American Symposium on Isotope Geology, Actas, Cordoba, Argentine*, 287-290.

Basei, M.A.S., Siga Jr, O., Reis Neto, J.M. dos, Passarelli, C.R., Prazeres, H.J., Kaulfuss, G., Sato, K. and Lima, P.S. de (1999b). Paleoproterozoic Granulitic Belts of the Brazilian Southern Region (PR-SC). *II South American Symposium on Isotope Geology, Actas, Cordoba, Argentine*, 291-294.

Bitencourt, M.F. and Hartmann, L.A. (1984). Reconhecimento geoquímico dos xistos magnesianos da região do Passo Feio, Caçapava do Sul, RS. In: *Congresso Brasileiro de Geologia, 33, Rio de Janeiro, SBG, Anais*, **6**, 2607-2614.

Bonhomme, M. and Cingolani, C. (1980). Mineralogia y geocronologia Rb-Sr y K-Ar de fracciones finas de la Formacion La Tinta, Provincia de Buenos Aires. *Asociación Geológica Argentina*

Revista, XXXV(4), 519-538.

Bossi, J. and Navarro, R. (1993). *Geologia del Uruguay*. Universidad de la República, Montevideo, 970 p.

Bossi, J. and Campal, N. (1992). Magmatismo y tectónica transcurrente durante el Paleozoico Inferior en Uruguay. In: *Paleozoico Inferior de Iberoamérica*, eds. Gutierrez-Marco, J. G., Saavedra, J. and Rabano, I., pp. 343-356. Universidad de Extremadura, España.

Bossi, J., Campal, N., Civetta, L.; Girardi, V.A., Mazzuchelli, M., Negrini, L., Rivalenti, G., Fragoso César, A.R.S., Sinigoi, S., Teixeira, W. and Piccirillo, E. (1993). Early Proterozoic dike swarms from western Uruguay: Geochemistry, Sm-Nd isotopes and petrogenesis. *Chemical Geology*, 106, 263-277.

Bossi, J. and Navarro, R. (1988). *Geología del Uruguay*. Depto. Publ. Universidad República, Montevideo, t. I, 453 p.

Bossi, J., Hartmann, L.A., Piñeyro, D., Leite, J.A.D., McNaughton, N.J. (1999). Geological and Geochronological Constraints on the Paleoproterozoic Evolution of the Isla Mala Granitic Suite, Uruguay. *II South American Symposium on Isotope Geology, Actas, Cordoba, Argentine*, 295-299.

Caldasso, A.L. da S., Camozzato, E., Ramgrab, G.E. and Silva, M.A. S. da (1988). Os granitóides Valsunganaa, Guabiruba e Faxinal no contextodos metamorfitos do Complexo Brusque, SC. In: *Cong. Bras. Geol., 35, Belem, SBG*, 3, 1104-1116.

Caldasso, A.L. da S., Krebs, A.S.J., Silva, M.A.S. da, Camozzato, E. and Ramgrab, G.E. (1995a). Programa de *Levantamentos Geológicos Básicos 1. 100.000; Folha Brusque (SG-22-Z-D-II-1), SC*. Brasília, CPRM, 243 p.

Caldasso, A.L. da S., Krebs, A.S.J., Silva, M.A.S. da, Camozzato, E. and Ramgrab, G.E. (1995b). *Programa de Levantamentos Geológicos Básicos 1. 100.000; Folha Botuverá (SG-22-Z-D-I-2), SC*. Brasília, CPRM, 303 p.

Campal, N. and Gancio F. (1993). Asociación volcanitas-piroclastitas de los Cerros Aguirre (dpto. Rocha): una nueva Formación y sus implicancias en la evolución del brasiliano en Uruguay. *I Simposio Internacional del Neoproterozóico-Cambrico de la Cuenca del Plata, La Paloma, Uruguay*, 2(44).

Campal, N., Gaucher, C., Schipilov, A. and Bossi, J. (1995). El Uruaçuano en el Uruguay: evidencias geológicas paleontológicas y radiométricas. In: *6 Simp. Sul-brasileiro de Geologia. Bol. Res. Expandidos. Porto Alegre, Brazil*, 97-99.

Campos Neto, M.C., and Figueiredo, M.C.H. (1995). The Rio Doce Orogeny, southeastern Brazil. *Journal of South American Earth Sciences*, 8, 143-162.

Castro, N. A. de; Basei, M.A.S. and Crósta, A. P. (1999). Caatinga Suite: W (Sn-Mo) specialized and other intrusive granitoids in the Brusque Group, Neoproterozoic of Santa Catarina, Southern Brazil. *Revista Brasileira Geociências* (in press).

Chemale Jr., Babinski, M., Van Schmus, W.R., Wildner, W. and Lima, E.F. (1997). U-Pb and Sm-Nd isotopic studies of Neoproterozoic to early Paleozoic belts in southern Brazil. I South American *Symposium on Isotope Geology, Campos do Jordão, SP, Brazil, Extended Abstract*, p. 71.

Cingolani, C., Spoturno, C. and Bonhomme, M. (1990). Resultados mineralógicos y geocrono-lógicos sobre las unidades Piedras de Afilar, Lavalleja y Barriga Negra, Rep. Oriental del Uruguay. In: *Primer Congr. Urug. Geol., Montevideo*, 1, 11-24.

Cingolani, C., Varela, R., Dalla Salda, L., Bossi, J., Ferrando, L., Piñeyro, D. and Schipilov, A. (1997). Rb/Sr geochronology from the Rio de la Plata Craton of Uruguay. In: *South American Symp. on Isotope Geology, São Paulo*, 1, 73-75.

Cordani, U.G., Halpern, M. and Berenholc, M. (1974). Comentários sobre as determinações geocronológicas da Folha Porto Alegre. In: *Carta Geológica do Brasil ao milionésimo, folha Porto Alegre (SH-22) e folha Lagoa Mirim (SI-22)*. DNPM, Ministério de Minas e Energia, Brasília, 70-84.

Cordani U.G. and Soliani Jr., E. (1990). Idades K/Ar e Rb/Sr das Ilhas Cristalinas de Rivera e Acegua (Uruguai e Rio Grand e do Sul) e seu enquadramento geotectônico regional. *Anais da Academia Brasileira de Ciências*, 62, 145-156.

Dalla Salda, L.H., Bossi, J. and Cingolani, C. (1988). The Rio de la Plata Cratonic Region of Southwestern Gondwanaland. *Episodes*, 11(4), 263-269.

Dalla Salda, L.H., Franzese, J.R. and Posadas, V.G. (1992). The 1,800 Ma mylonite-anatectic granitoid association in Tand ilia, Argentina. In: R. Mason (Editor), *Basement Tectonics 7*, pp. 161-174. Netherlands.

Fernández A. and Preciozzi F. (1974). La formación Arroyo Grand e y los granitoides asociados. *Cong. Bras. Geol., Anais, SBG, Porto Alegre, Brazil*, 28, 213-216.

Fernandes, L.A.D., Menegat, R., Costa, A.F.U., Koster, E., Porcher, C.C., Tommasi, A., Kraemer, G., Ramgrab, G.E. and Camozzato, E. (1995a). Evolução Tectônica do cinturão Dom Feliciano no Escudo sul-riogrand ense: Parte I – uma contribuição a partir do registro geológico. *Revista Brasileira de Geociências*, 25(4), 351-374.

Fernandes, L.A.D., Menegat, R., Costa, A.F.U., Koster, E., Porcher, C.C., Tommasi, A., Kraemer, G., Ramgrab, G.E. and Camozzato, E. (1995b). Evolução Tectônica do cinturão Dom Feliciano no Escudo sul-riogrand ense: Parte II – uma contribuição a partir das assinaturas geofísicas. *Revista Brasileira de Geociências*, 25(4), 375-384.

Fragoso Cesar, A.R.S. (1980). O Craton do Rio de La Plata e o Cinturão Dom Feliciano no Escudo Uruguaio-Sul Riogrand ense. *Congr. Bras. de Geol., Anais, SBG, 31, Camboriú*, 5, 2879-2891.

Fragoso Cesar, A.R.S. and Soliani Jr., E. (1984). Compartimentação Tectonica do Craton Rio de La Plata. *XXXIII Congr. Bras. Geol., Rio de Janeiro, Anais*, 2426-2434.

Fragoso Cesar, A.R.S., Figueiredo, M.C.H. de, Soliani Jr., E. and Faccini, U.F. (1986). O Batolito Pelotas (Proterozóico Superior/ Eopaleozóico) no Escudo do Rio Grande do Sul. *Congr. Bras. de Geol., Anais, SBG, 34, Goiânia*, 3, 1322-1343.

Fragoso Cesar, A.R. (1991). Tectônica de Placas no Ciclo Brasiliano: As orogenias dos cinturões Dom Feliciano e Ribeira no Rio Grand e do Sul. *Thesis submitted for the degree of Doctor of Philosophy, University of São Paulo, São Paulo, Brazil*, 362 p. (unpublished).

Fragoso Cesar, A.R.S., Machado, R. and Gomez Rifas, C. (1987). Observações sobre o Cinturão Dom Feliciano no escudo uruguaio e correlações com o escudo do Rio Grand e do Sul. In: *Simp. Sul-Bras. Geol., 3. Curitiba, SBG*, 2, 791-809.

Fragoso Cesar, A.R. and Machado, R. (1997). Neoproterozoic terranes of the Gaucho shield (southern Brazil and Uruguay). *I South American Symposium on Isotope Geology, Campos do Jordão, SP, Brazil, Extended Abstract*, 65-67.

Fragoso Cesar, A.R.S., Machado, R., Philipp, R., Mello, F.M., Endo, I., Nimmer, A.R., Preciozzi, F.P., Fambrini, G.L. and Sayeg, H.S. (1998). Terrenos Suspeitos a sul da Placa Sanfranciscana (SE da Plataforma Sul-Americana). *Actas del II Gongreso Uruguayo de Geologia, Punta del Este, Uruguay*, 1-6.

Fragoso Cesar, A.R., Ferreira Filho, W. da S., Fambrini, G.L., Machado, R., Riccomini, C., Almeida, R.P. de, Pelosi, A.P.M.R. and Janikian, L. (1999). Significado Tectônico do Magmatismo Rodeio Velho no Rift Guaritas (Eopaleozóico do Rio Grand e do Sul, Brasil). *I Simpósio sobre Vulcanismo e Ambientes Associados, Abstracts, Gramado*, RS, p. 16.

Frantz, J.C. and Jost, H. (1983). Petrologia dos granitos estaníferos do Rio Grande do Sul. *Actas I Simp. Sul-Bras. de Geol. Porto Alegre, Brazil*, 49-67.

Frimmel, H.E. (1995). Metamorphic evolution of the Gariep Belt. *South African Journal of Geology*, 98(2), 176-190.

Frimmel, H.E., Almond, J. and Gresse, P.G. (1998). *Birth of Gondwana*. Gariep Belt Field Trip (Gondwana 10 field guide), 75 p.

Frimmel, H.E. and Frank, W. (1998). Neoproterozoic tectono-thermal evolution of the Gariep Belt and its basement, Namibia/South Africa. *Precambrian Research*, 90(1998), 1-28.

Gaucher, C. and Sprechmann, P. (1998). Grupo Arroyo del Soldado: Paleontologia, Edad y Correlaciones (Vendiano-Cambrico Inferior, Uruguay). *Actas del II Gongreso Uruguayo de Geologia, Punta del Este, Uruguay*, 183-187.

Guacher, C., Sprechmann, P. and Schipilov, A. (1996). Upper and Middle Proterozoic fossiliferous sedimentary sequences of the Nico Perez Terrane of Uruguay: Lithostratigraphic units, paleontology, depositional environments and correlations. *Neues Jahrbuch für Geologie und Paläontologie. Abhandlungen*, 199(3), 339-367.

Halinann, S.E. and Mantovani, S.M.M. (1993). Structural Framework of the Southern Brazilian Shield: the perspective from the Gravity models. *III Intern. Geoph. Congress, Anais, SBGF, Rio de Janeiro*, 2, 1078, 1083 p.

Hanson, R.E., Martin, M.W., Bowring, S.A. and Munyanyiawa, H. (1998). U-Pb zircon age for the Umkondo dolerites, eastern Zimbabwe: 1.1 Ga large igneous province in southern Africa-east Antartica and possible Rodinia correlations. *Geology*, 26(12), 1143-1146.

Harara, M. (1996). *Análise estrutural, petrológica e geocronológica dos litotipos da região de Piên (PR) e adjacências.* Master thesis presented to the Institute of Geosciences - USP, São Paulo, Brazil, 196 p. (unplublished).

Harris, N.B.W., Hawkesworth, C.J., Van Calsteren, P. and McDermott, P.F. (1987). Evolution of continental crust in southern Africa. *Earth and Planetary Science Letters*, 83, 85-93.

Hart S. (1966). *Radimetric ages in Uruguay and Argentina and their implications concerning continental drift.* Geological Soc. Am. Annual Meeting, San Francisco, 86 p.

Hartmman, L. A., Silva, L.C. and Orlandi, V. (1979). O Complexo Granulítico de Santa Catarina. *Acta Geológica Leopoldensia*, 6, 94-112.

Hartmann, L.A., Tindle, A. and Bitencourt, M.F. (1990). O metamorfismo de fácies anfibolito no Complexo Passo Feio, RS, com base em química dos minerais. *Pesquisas*, 17(1-2), 62-71.

Hartmann, L.A. (1998). Deepest exposed crust of Brazil – Geochemistry of Paleoproterozoic depleted Santa Maria Chico granulites. *Gondwana Research*, 1, 331-341.

Hartmann, L.A., Silva, L.C., Remus, M.V.D., Leite, J.A.D. and Philpp, R.P. (1998). Evolução Geotectônica do Sul do Brasil e Uruguai entre 3.3Ga e 470 Ma. *Actas del II Gongreso Uruguayo de Geologia, Punta del Este, Uruguay*, 277-284.

Hartmann, L.A., Leite, J.A.D., McNaughton, N.J., Santos, J.O.S. (1999). Deepest exposed crust of Brazil-SHRIMP establishes three events. *Geology*, 27(10), 947-950.

Hartnady, C., Joubert, P. and Stowe, C. (1985). Proterozoic Crustal Evolution in Southwestern Africa. *Episodes*, 8(4), 236-244.

Hasui Y., Carneiro C. dal R. and Coimbra A.M. (1975). The Ribera Folded Belt. *Revista Brasileira de Geociências*, 5, 257-266.

Hoffman, P.F. (1999). The brake-up of Rodinia, birth of Gondwana, true polar wander and the snowball Earth. *Journal of African Earth Sciences*, 28(1), 17-33.

Jost, H. (1981). *Geology and Metallogeny of the Santana da Boa Vista region, South Brazil.* Thesis submitted for the degree of Doctor of Philosophy, University of Athens, Georgia, USA, 208 p. (unpublished).

Jost, H. (1982). Condições de metamorfismo de uma fração da faixa de Dobramentos Tijucas no Rio Grande do Sul. *Acta Geológica Leopoldênsia*, 4, 27-60.

Jung, S., Mezger, K. and Hoernes, S. (1998). Petrology and geochemistry of syn- to post-collisional metaluminous A-type granites – a major and trace element and Nd-Sr-Pb-O-isotope study from the Proterozoic Damara Belt, Namibia. *Lithos*, 45, 147-175.

Krebs, A.S.J., Lopes, R. da C. and Camozzato, E. (1990). Caracterização faciológica do Grupo Itajaí na Folha Botuverá (SC). In: *Congr. Bras. Geol., 36, Natal, 1990. Anais, Natal, SBG*, 1, 82-92.

Koester, E. (1995). *Evolução geológica do magmatismo sintectônico à Zona de Cisalhamento Transcorrente Dorsal do Canguçu, região de Encruzilhada do Sul (RS).* Thesis submitted for the degree of Master of Philosophy, University of Rio Grande do Sul, RS, Brazil, 240 p. (unpublished).

Leite, J.A.D. (1995). Datação SHRIMP U-Pb em zircões e o exemplo de dois corpos graníticos contrastantes no escudo sul-riograndense. *VI Simpósio Sul Brasileiro de Geologia, Porto Alegre, RS, Brazil, Extended Abstracts*, 5-12.

Leite, J.A.D. (1997). *A origem dos harzburgitos da Sequência Cerro Mantiqueiras e implicações tectônicas para o desenvolvimento do Neoproterozóico na porção oeste do Escudo Sul-Riograndense.* Thesis submitted for the degree of Doctor of Philosophy, Federal University of Rio Grande do Sul, RS, Brazil, 224 p. (unpublished).

Leite, J.A.D. and Hartmann, L.A. (1997). Neoproterozoic evolution of Cerro Mantiqueiras region in southern most Brazil: evidence from coupled SHRIMP zircon dating and Nd, Sr isotopic data. *I South American Symposium on Isotope Geology, Campos do Jordão, SP, Brazil, Extended Abstract*, 167-169.

Leite, J.A.D., Hartmann, L.A., Fernandes, L.A.D., McNaughton, N.J., Soliani Jr., E., Koester, E., Santos, J.O.S. and Vasconcellos, M.A.Z. Zircon U/Pb SHRIMP dating of Gneissic Basement of Dom Feliciano Belt, Southernmost Brazil. *Journal of South American Earth Sciences* (submitted to).

Lema, H. and Cucchi, R. (1981). Hallazco de Metavulcanitas en el Cerro Tandileufu, Provincia de Buenos Aires. *Revista de la Asociación Geológica Argentina*, 36(1), 103-104.

Linares, E. (1977). Catalogo de edades radiometricas determinadas para la Republica Argentina. *Asociación Geológica Argentina*, Buenos Aires, serv.B, 4.

Lopes, R. da C., Wildner, W., Sander, A. and Camozzato, E. (1999). Alogrupo Guaritas: Aspectos Gerais e Considerações sobre o Posicionamento do Vulcanismo Rodeio Velho (Encerramento do Ciclo Brasiliano ou Instalação da Bacia do Paraná?). *I Simpósio sobre Vulcanismo e Ambientes Associados, Abstracts, Gramado, RS*, p. 17.

Maack, R. (1947). Breves notícias sobre a geologia dos estados do Paraná e Santa Catarina. *Arquivo Biológico e Tecnológico, Universidade Federal do Paraná*, 2, 63-154.

Machiavelli, A., Basei, M.A.S. and Siga Jr., O. (1993). *Suíte Granítica Rio Pien: um arco magmático do Proterozóico Superior na Microplaca Curitiba.* Geochimica Brasiliensis, 7(2), 113-129.

Mantovani, S.M., Hawkesworth, C.J. and Basei, M.A.S. (1987). Nd and Pb isotope studies bearing on the crustal evolution of southeastern Brazil. *Revista Brasileira de Geociências*, 17(3), 263-269.

May, S.E. (1990). *Pan-African Magmatism and Regional Tectonics of South Brazil.* Thesis submitted for the degree of Doctor of Philosophy, The Open University, UK, 327 p. (unpublished).

Marques, J.C., Jost, H., Roisenberg, A. and Frantz, J.C. (1998a). Eventos ígneos da Suíte Metamórfica Porongos na área da Antiforme Capané, Cachoeira do Sul, southern Brazil. *Revista Brasileira de Geociências*, 28(4), 419-430.

Marques, J.C., Jost, H., Roisenberg, A. and Frantz, J.C. (1998b). Rochas metassedimentares, geologia estrutural e metamorfismo da suíte metamórfica Porongos na área da Antiforme Capané, Cachoeira do Sul, RS. *Revista Brasileira de Geociências*, 28(4), 467-472.

Masquelin, H. (1990). Análisis estructural de las zonas de cizalla en las migmatitas de Punta del Este – Uruguay. *Rev. Estudos Tecnológicos, Acta Geológica Leopoldensia*, 13(30), 139-158.

Masquelin, H.C. and Sanches Bettucci, L. (1993). Propuesta de evolución Tectono-sedimentária para la cuenca de Piriápolis, Uruguay. *Revista Brasileira de Geociências*, 23 (3), 313-322.

McDermott, P.F. (1986). *Granite petrogenesis and crustal evolution studies in the Damara Pan-African orogenic belt, Namibia.*

Thesis submitted for the degree of Doctor of Philosophy, The Open University, UK, 303 p. (unpublished).
McDermott, P.F. and Hawkesworth, C.J. (1990). Intracrustal recycling and upper-crustal evolution: A case study from the Pan – African Damara mobile belt, central Namibia. *Chemical Geology*, 83, 263-280.
Möller, A., Mezger, K. and Schenk, V. (1998). Crustal Age Domains and the Evolution of the Continental Crust in the Mozambique belt of Tanzania: Combined Sm-Nd, Rb-Sr and Pb-Pb Isotopic Evidence. *Journal of Petrology*, 39(4), 749-783.
Pain P.S.G., Leipnitz, I.I., Rosa, A.N.Z. da and Rosa, A.A.S. da (1997). Preliminary report on the occurrence of Chancelloria sp. In the Itajaí Basin, southern Brazil. *Revista Brasileira de Geociências*, 27(3), 303-308.
Passarelli, C.R. (1996). *Análise estrutural e caracterização do magmatismo da Zona de Cisalhamento Major Gercino, SC.* Master thesis, University of São Paulo, São Paulo (Brazil), 179 p. (unpublished).
Passarelli, C.R., Basei, M.A.S., Ahrendt, H., Wemmer, K. and Siga Jr., O. (1997). Geochronological evolution of Major Gercino shear zone, south Brazil. *I South-American Symposium on Isotope Geology, Ext. Abstracts, Campos do Jordão, SP, Brazil*, 231-233.
Phillip, R.P. (1998). *A Evolução geológica e Tectônica do Batolito Pelotas no Rio Grande do Sul.* Thesis submitted for the degree of Doctor of Philosophy, University of São Paulo, São Paulo (Brazil), vol. 1, 255 p. (unpublished).
Picanço J. Tasssinari, C.C.G., Cordani, U.G. and Nutman, A.P. (1998). Idades U-Pb (SHRIMP), Sm-Nd e Rb-Sr em rochas do Maciço de Itatins (SP): Evidências de Evolução Policíclica. *Anais da Academia Brasileira Ciências*, 70(1), 139-150.
Poiré, D.G. (1993). Estratigrafia del precambrico sedimentario de Olivarria, Sierras Bayas, Provincia de Buenos Aires, Argentina. *XII Congreso Argentino y II Congreso de Exploracion de Hidrocarburos. Actas*, 7 (II), 1-11.
Porcher, C.C., McNaughton, N.J., Leite, J.A.D., Hartmann, L.A. and Fernandes, L.A.D. (1999). Idade Shrimp em zircão: vulcanismo ácido do Complexo Metamórfico Porongos. *I Simpósio sobre Vulcanismo e Ambientes Associados, Abstracts, Gramado, RS*, p. 110.
Porcher, C.C., McNaughton, N.J., Leite, J.A.D., Hartmann, L.A. and Fernandes, L.A.D. (In press). Geochronology of Proterozoic basement-cover relationships in the southern Brazilian Porongos Belt. *Precambrian Research* (submitted to).
Preciozzi, F., Spoturno, J., Heinzen, W. and Rossi, P. (1985). *Memoria Explicativa de la Carta Geológica del Uruguay a la escala 1/ 500.000.* DINAMIGE - M.I.E.M., 90 p.
Preciozzi, F. and Bourne, J. (1992). Petrography and geochemistry of the Arroyo de la Virgen and Isla Mala plutons, Southern Uruguay: Early Proterozoic implications. *Journal South American Earth Sciences*, 6, 169-181.
Preciozzi, F., Masquelin, H. and Basei, M.A.S. (1999a). The Namaqua / Grenville Terrane of Eastern Uruguay. *II South American Symposium on Isotope Geology, Actas*, Cordoba, Argentine, 338-340.
Preciozzi, F., Basei, M.A.S. and Masquelin, E. (1999b). New Geochronological data from the Piedra Alta Terrane (Rio de La Plata Craton), *II South American Symposium on Isotope Geology, Actas, Cordoba, Argentine*, 341 - 343.
Porada, H. (1979). The Damara-Ribeira orogen of the PanAfrica-Brasiliano cycle in Namibia (Southwest Africa) and Brazil as interpreted in terms of continental collision. *Tectonophysics, Australia*, 57(2-4), 237-265.
Porada, H. (1989). Pan–African rifting and orogenesis in southern to equatorial Africa and eastern Brazil. *Precambrian Research*, 44, 103–136.
Pothe de Baldis, E.D., Baldis, B. and Cuomo, J. (1983). Los fósiles precambricos de la Formacion Sierras Bayas (Olavarria) y su importancia intercontinental. *Asociacion Geológica Argentina Revista*, XXXVIII(1), 73-83.
Quintas, M.C.L. (1994). *O embasamento da Bacia do Paraná: reconstrução geofísica de seu arcabouço.* Thesis submitted for the degree of Doctor of Philosophy, University of São Paulo, São Paulo, Brazil, 218 p. (unpublished).
Ramos, V.A., Leguizamon, A., Kay, S.M. and Teruggi, M. (1990). Evolution Tectonica de Las Sierras de Tand il (Provincia de Buenos Aires). *XI Congreso Geologico Argentino, San Juan, 1990, Actas*, II, 357-360.
Reid, D.L. (1997). Sm-Nd age and REE geochemistry of Proterozoic arc-related igneous rocks in the Richtersveld Subprovince, Namaqua Mobile Belt, Southern Africa. *Journal of African Earth Sciences*, 24(4), 621-633.
Remus, M.V.D. (1998). *Metalogênese do Cobre e Evolução crustal da região de Caçapava do Sul-RS.* Thesis submitted for the degree of Doctor of Philosophy, Federal University of Rio Grande do Sul, RS, Brazil (unpublished).
Remus, M.V.D., Hartmann, L.A., McNaughton, N.J. and Fletcher, I.R. (1999). SHRIMP U-Pb zircon ages of volcanism from the São Gabriel Block, southern Brazil. *I Simpósio sobre Vulcanismo e Ambientes Associados, Abstracts, Gramado, RS*, p. 110.
Remus, M.V.D., Hartmann, L.A., McNaughton, N.J., Groves, D.I. and Fletcher, I.R. (in press). The link between hydrothermal epigenetic copper mineralization and the Caçapava Granite of the Brasiliano Cycle in southern Brazil. *JSAES* (Submitted to).
Rostirolla, S. (1991). *Tectônica e sedimentação da Bacia do Itajaí-SC.* Master thesis presented to the Escola de Minas, Universidade Federal de Ouro Preto, Minas Gerais (Brazil), 132 p. (unpublished).
Rozendaal, A., Gresse, P.G., Scheepers, R. and Le Roux, J.P. (1999). Neoproterozoic to Early Cambrian Crustal Evolution of the Pan-African Saldania Belt, South Africa. *Precambrian Research*, 97(3-4), 303-323.
Salamuni, R., Bigarella, J.J. and Takeda, F.K. (1961). Considerações sobre a estratigrafia e tectônica da Série Itajaí. Curitiba, PR. *Boletim Paranaense Geografia* (415), 188-201.
Sander, A. (1992). *Petrologia e litogeoquímica de uma parcela da sequência vulcano-sedimentar do complexo metamórfico Brusque na região do Ribeirão do Ouro, SC.* Thesis submitted for the degree of Doctor of Philosophy, Federal University of Rio Grande do Sul, RS, Brazil, 167 p. (unpublished).
Sanches Bettucci, L. (1998). *Evolución Tectónica del cinturón Dom Feliciano en la región Minas-Piriápolis, Republica Oriental del Uruguay.* Thesis submitted for the degree of Doctor of Philosophy, Facultad Ciencias Exactas y Naturales, University of Buenos Aires (Argentina), 234 p. (unpublished).
Sanches Bettucci, L. and Mezzano Burgueño, A. (1993). Análisis sedimentológico y faciológico de la Formación Rocha (ex-Grupo Rocha). *Revista Brasileira de Geociências*, 23(3), 323-329.
Schukowsky, W., Vasconcellos, A.C.B.C. and Mantovani, M.S.M. (1991). Estruturação dos terrenos Pré-Cambrianos da região sul do Brasil e oeste do Uruguai: um estudo por modelamento gravimétrico. *Revista Brasileira de Geofísica*, 9(2), 275-287.
Siga Jr., O (1995). *Domínios tectônicos da região sudeste do Parand e nordeste de Santa Catarina: Geocronologia e Evolução Crustal.* PhD thesis presented to the Institute of Geosciences - USP, São Paulo, Brazil, 290 p. (unpublished).
Siga Jr., O., Basei, M.A.S., Reis Neto, J.M. dos, Machiavelli, A. and Harara, O.M. (1995). O Complexo Atuba: um cinturão Paleoproterozóico intensamente retrabalhado no Neoproterozóico. *Boletim Instituto de Geociências, Universidade de São Paulo*, 26, 69-98.
Silva, L.C. da and Dias, A. A. (1981). Os segmentos mediano e setentrional do Escudo Catarinense: I – Geologia. *Acta Geológica Leopoldênsia, São Lepoldo, RS*, 5(10), 3-120.
Silva, L.C. da, Oliveira, J.M.P., Aumond, J.J., Lopes, R.M.M., Eipper, J. and Ferro, G. (1985). Caracterização petrográfica da Sequência (Meta) Vulcano-sedimentar Rio do Oliveira (Cinturão do Itajaí Mirim, SC) In: *Simp. Sul-Bras. Geol., 2, Florianópolis, SBG*, 11-23.
Silva, L.C. da, Hartmann, L.A., McNaughtoon, N.J., Fletcher, I.R. (1997).

Shrimp U/-Pb Zircon Dating of Neoproterozoic Granitic Magmatism and Collision in the Pelotas Batholith, Southernmost Brazil. International Geology Review, 41(1), 531-551.

Soliani Jr., E. (1986). *Os dados geocronológicos do Escudo Sul-riograndense e suas implicações de ordem geotectônica*. Thesis submitted for the degree of Doctor of Philosophy, University of São Paulo, São Paulo, Brazil, 425 p. (unpublished).

Teruggi, M., Leguizamon, M.A. and Ramos, V.A. (1988). Metamorfitas de Bajo Grado con afinidades oceanicas en el basamento de Tandil: sus implicaciones geotectonicas, Província de Buenos Aires. *Asociación Geológica Argentina, Revista*, **XLIII**(3), 366-374.

Tessari, R.I. and Picada, R.S. (1966). Geologia da Quadricula de Encruzilhada do Sul, RS. *Boletim da Divisão de Fomento a Produção Mineral, DNPM, Rio de Janeiro*, **124**, 1-147

Trompette, R. (1994). *Geology of Western Gondwanaland 2000 - 500Ma. The Pan-African/Brasiliano amalgamation of South America and adjacent Africa*. Balkema, Rotterdam, 350 p.

Tommasi, A., Fernandes, L.A.D., Machado, R.P., Kraemer, G. and Koester, E. (1994). Incorporação de lascas do manto à crosta continental nas porções externas do cinturão Dom Feliciano – Evidências de uma Zona de Cisalhamento de Escala Litosférica. *Pesquisas*, **21**(2), 79-84.

Umpierre, M. and Halpern, M. (1971). Edades Rb-Sr en rocas cristalinas del sur de la Republica Oriental del Uruguay. *Asociación Geológica Argentina, Buenos Aires*, **XXVI**(2), 133-151.

Varela, R., Cingolani, C. and Dalla Salda, L. (1988). Geocronologia Rb-Sr en granitoides del basamento de Tandil, Provincia de Buenos Aires, Argentina. *Segundas Jornadas Geologicas Bonarenses, Bahia Blanca, Actas*, 291-304.

Zanini, L.F.P., Branco, P.M., Camozzato, E. and Ramgrab, G.E. (1997). *Programa de Levantamentos Geológicos Básicos 1: 100.000; Folhas Florianópolis e Lagoa*, Brasília, CPRM, 252 p.

Orogenic Systems From Southwestern Gondwana

An Approach to Brasiliano-Pan African Cycle and Orogenic Collage in Southeastern Brazil

Mario da Costa Campos Neto

The geological evolution toward the amalgamation of the Gondwana Supercontinent started immediately after the Grenville Orogenic Cycle that built the predecessor of the Rodinia Supercontinent. Since the beginning of the Neoproterozoic break-up and drift controlled by taphrogeny has driven plate kinematics toward orogenic interactions. These processes continued for at least 500 Ma until the Lower Ordovician, following all the steps of the classical Wilson Cycle. An interesting characteristic of these orogenies is the diachronism of events, many of which were coeval with taphrogenic processes elsewhere. Many orogenies were controlled by different kinds of plate interaction occurring diachronically in different places. This plurality of processes that converged to the closure of a wide oceanic space populated by small continental fragments (terranes or microplates) may be described in terms of orogenic systems or branching systems of orogens, rather than the general and geometry-related mobile belt model. The collection of orogenic systems leading to supercontinent amalgamation represents the orogenic collage (Sengor, 1990).

The advance of geochronological knowledge of southeastern Brazil over the last decade, mostly due to advances in U/Pb and Sm/Nd dating techniques, has helped to place more precise limits on events including the recognition of coeval metamorphism and plutonism. Isotope studies have provided a powerful tool for deciphering the sequence of tectonic events including the major lithosphere mantle accretionary processes and contrasting lithosphere signatures between terranes. In southeastern Brazil, two major orogenic systems are recognized. The first, the Tocantins Orogenic System, was related to the closure of the Goianides Ocean, a Tonian ocean situated to the W of the São Francisco Plate. The Tocantins Orogenic System comprises several orogens, which have been elsewhere designated as the Brasília Belt (Fuck *et al.*, 1993) and the Alto Rio Grande Belt (Hasui, 1982). The widespread break-up of Rodinia took place during the Middle Cryogenian. It accounts for the separation of Eastern Gondwana from Laurentia (Park, 1994) and the generation of the Adamastor Ocean (Hartnady *et al.*, 1985), facing the Congo, Kalahari and São Francisco plates.

The main plate convergence controlled several collisions and terrane dockage leading to the southeastward accretion of the Brazilian continental crust. These processes were collectively related to the Mantiqueira Orogenic System. This system comprises the northern part of the Ribeira Belt (Almeida *et al.*, 1973) and the southern part of the Dom Feliciano Belt (Fragoso-Cesar, 1980). The proposed Tocantins and Mantiqueira orogenic systems have their origin in the structural provinces concept of Almeida *et al.* (1981).

The evolution of different types of rifts and related passive continental margins and the importance of their geometry within a plate kinematics scenario may be explained in terms of the model proposed by Lister *et al.* (1986).

Major Continental Plates Framework

The main geotectonic provinces, with special regard to the Precambrian of the South American Platform (the cratonic area for the Andean orogenic episodes) are shown in Figure 1. These provinces are related to the huge continental lithosphere plates (Brito Neves and Cordani, 1991; Brito Neves *et al.*, 1999), that were rifted apart during the break-up of Rodinia at the beginning of the Neoproterozoic (Dalziel, 1997; Weil *et al.*, 1998). Successive collision and plate indentation processes during the global Brasiliano-Pan African Orogenic Cycle record the further amalgamation of these old shields to form the Gondwana Supercontinent.

The Amazonas Plate records roughly NW-SE trending belts, which are successively younger to the SW and added to an Archean northeastern province (Tassinari *et al.*, 1996). They comprise Paleoproterozoic belts from the Transamazonian Cycle (*c.* 2.2 - 1.8 Ga), which are followed by Statherian magmatic arc, and by the southern Mesoproterozoic collision of the Pampia Terrane. A wide settlement of the Grenville Province exceed the Neoproterozoic limits of the Amazonas Plate, if one takes into account the further rifting apart of the Arequipa-Antofalla Terrane and the Eastern Laurentia Plate (Ramos and Vujovich, 1995; Bettencourt *et al.*, 1996; Balburg and Hervé, 1997).

The Rio de la Plata Plate is almost covered by the Paleozoic-Mesozoic Paraná Basin. Its southeastern edge, where the best exposures are to be seen, comprises a cratonic granite-greenstone province, *c.* 2.0 Ga old crosscut by undeformed mafic dyke swarm of *c.* 1.8 Ga. There occurs a reworked gneiss-granulitic border. The Rio Apa granite-gneiss seems to belong to this plate, emerging from the Phanerozoic basin as a small Paleoproterozoic block. The old rocks of the Rio de la Plata Plate first recognized by deep drilling (Cordani *et al.*, 1984) were revealed, as a whole, by geophysical research, including gravity surveying (Mantovani and Shukowsky, 1996), and thermal constraining for P and S-wave velocity perturbations (Van Decar *et al.*, 1995). The paleomagnetic data for the end of the Paleoproterozoic postulate a Laurentia connection of this major segment of continental plate (Agrella Filho *et al.*, 1999).

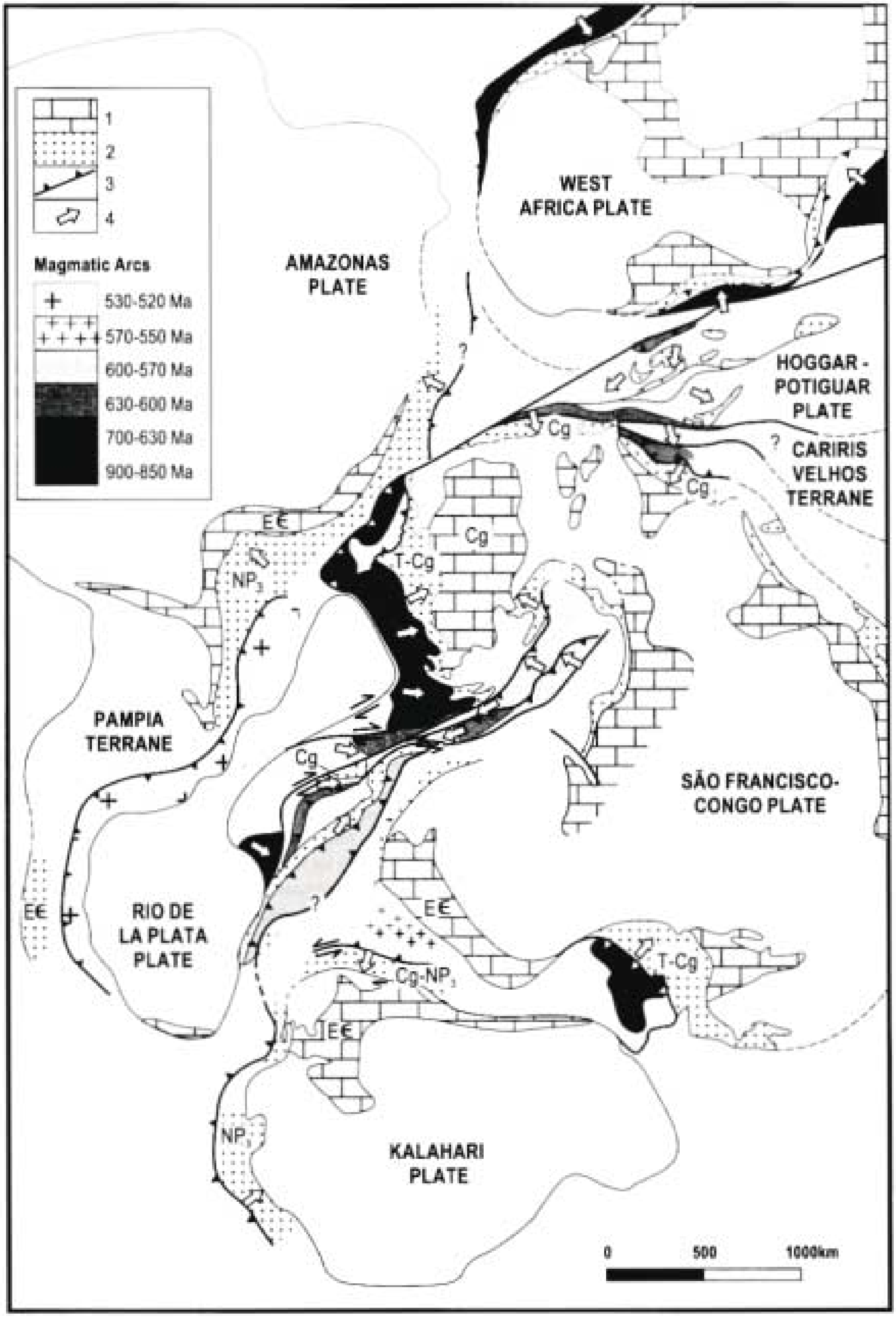


Fig. 1 - Major Western Gondwana continental plates and magmatic arcs. 1. Platform cover, 2. Passive continental margin deposits or thinner continental crust, 3. Subduction zones with sense of dip, 4. Structural vergence.

The horse head-shaped São Francisco Plate, which represents the northwestern extension of the Congo Plate, was defined by Almeida (1977) and Alkmim *et al.* (1993). The remnants of segments of oceanic plate between these blocks can still be recognized.

Statherian taphrogeny: attempts to break the São Francisco Plate up

The major and widespread Upper Rhyacian (*c.* 2.1 Ga) orogenic collage, accompanied by juvenile accretion and the reworking of the oldest Archean crust (Teixeira *et al.*, 1998), consolidated the proto-São Francisco Plate. Following this phase of plate convergence there developed, locally, zones of intense intraplate extension that were collectively integrated in a broad taphrogeny. Failed rifts were responsible for the development of the large Espinhaço Basin, elongated N-S, and the western part of the Araí Basin (Fig. 2). Rhyolitic rocks occur at the base of the sedimentary sequence in both basins, and these are coeval with small tin-bearing granite intrusives. These continental, anorogenic volcanic and plutonic suites are assigned a Statherian age (1.77 - 1.60 Ga, Brito Neves *et al.*, 1979; Turpin *et al.*, 1988; Machado *et al.*, 1989; Pimentel *et al.*, 1991). The sedimentary sequence (about 1200 m thick), consisting of siliciclastic continent-dominated rocks (conglomerate, coarse-grained quartzite and siltstone) form the depositional systems of the pre-rift and rift stages. The upper units of the Espinhaço Supergroup represent a seaward connection (wave-ripple quartzite interbedded with shale, locally with carbonate) unconformably overlie the adjacent non-stretched Archean basement. They were related to a phase of crustal downwarping of the thermal-flexural type that controlled the change in the subsidence regimes without causing the breaking-up of the interior regions of the São Francisco Plate (Dominguez, 1993; Martins-Neto, 1998).

Tabular bodies, up to 110 m thick, of a sub-alkaline to alkaline suite of (Fe-hastingsite and Fe-salite) magnetite-bearing granite, related to an anorogenic, extensional tectonic regime occur at the southern edge of the São Francisco Plate (Taguar Granite), having an Ectasian age (Rb/Sr isochron of 1.4 Ga, $Sr^{87}/Sr^{86}_{(i)}$ = 0.719, Vasconcellos, 1988). Thus, this magmatic event is evidence of another phase of extensional tectonics during the Ectasian, which was followed by the Statherian taphrogeny. However, there is no evidence for further orogenesis in the São Francisco Plate from the end of the Rhyacian System until the Neoproterozoic.

Fragments of the Rodinia history

Central Goiás Terrane

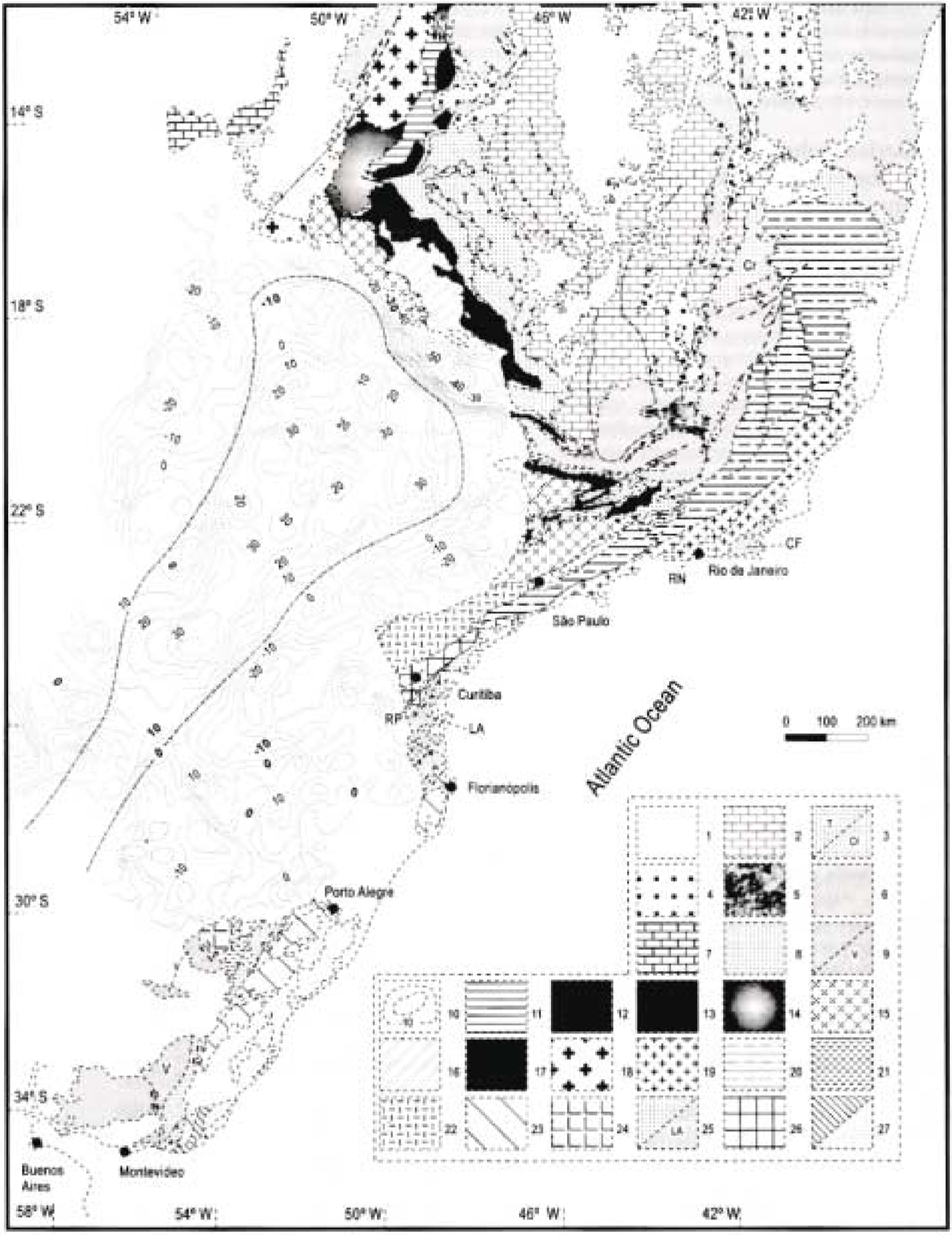
A NNE-SSW trending belt comprising an Archean-Proterozoic continental remnant of an older plate against which docked the Neoproterozoic Mara Rosa Island Arc binds the western edge of the São Francisco Plate. This fragment consists of Paleoproterozoic basic-ultrabasic complexes and Paleoproterozoic and Mesoproterozoic vulcano-sedimentary sequences (Fig. 2).

The Archean fragment comprises a typical granite-greenstone terrane. The metasediments are carbonaceous schist and metarhythmite turbidite beds. The metavolcanic rocks consist of komatiite flows at the base and intercalated mafic and felsic flows at the top. The metaplutonic rocks are mainly tonalite and granodiorite (Rivalenti *et al.*, 1989; Jost *et al.*, 1996). Dome-and-keel is the structural framework, and the interference geometry was produced by successive episodes of plutonism between 2.85 and 2.70 Ga (Queiroz *et al.*, 1999) coeval with the komatiite lava flows (Arndt *et al.*, 1989). The metamorphic recrystallisation quickly followed magmatism (2.7 Ga), and was overprinted by the Rhyacian Orogeny (2.14 Ga, Queiroz *et al.*, 1999).

Stratiform mafic-ultramafic complexes (Barro Alto, Niquelândia and Cana Brava) occur discontinuously for 300 km (Danni *et al.*, 1982) in a narrow belt of mylonitic granite-gneiss. Gabbro, metagabbronorite, peridotite, pyroxenite, anorthosite and late diorite intrusives are the main rock-types found in these complexes. They show petrological gradation from a less-differentiated base with pyroxene and olivine cumulates toward a more-differentiated top with plagioclase cumulates. Interlayers of high-grade metamorphic metasediments can be observed as well as xenoliths in the mafic rocks. Data suggest an anorogenic extensional environment for these intrusions (Girardi *et al.*, 1986; Ferreira Filho *et al.*, 1992) that was rifted apart from an uncertain continental plate. Well-constrained U/Pb (SHRIMP) dates from Niquelândia (Correia *et al.*, 1996) and a Sm/Nd isochron from Cana-Brava (Fugi, 1989) suggest that these complexes were emplaced at the Rhyacian/Orosirian boundary (2.0 Ga). The quartz-diorite intrusion at the base of the Barro Alto Complex gave a U/Pb zircon age of 1.7 Ga (Suita *et al.*, 1996).

At the southeastern extremity of the Archean fragment, there is exposed a narrow metavolcano-sedimentary belt (Moçâmedes Sequence) and was described by Simões and Fuck (1984). It consists of a few metabasalt flows of low-K tholeiitic affinity, followed by a calc-alkaline suite of basic/intermediate and felsic metavolcanic rocks intercalated with metatuff, metachert and pelitic schist, evidence for an active volcanic arc (Barbosa and Jost, 1990). The Rb/Sr ages for this plate convergence manifest by felsic volcanism suggests that this crosses the Paleoproterozoic Orosirian and fini-Statherian boundary with an Sm/Nd T_{DM} age of *c.* 2.2 Ga (Pimentel *et al.*, 1996a).

Another metavolcano-sedimentary sequence occurs discontinuously as an allochthon over the mafic-ultramafic complexes (Fig. 2). These sequences (Juscelândia, Indaianópolis and Palmeirópolis, Marini *et al.*, 1984) are related to a bimodal, continental-type sub-alkaline tholeitic basalt volcanism, and rhyolitic rocks derived from crustal-mantle mixing of felsic magma. The uppermost volcanic layer represents basic rocks from N-MORB type basalt (Moraes *et al.*, 1998) ascribing a rift-drift evolution, the age of which unfortunately remains unknown. However these rocks were metamorphosed in the upper amphibolite facies that is coeval with the granulite facies metamorphism of the mafic-ultramaphic complexes (Moraes and Fuck, 1994; Ferreira Filho *et al.*, 1998). The age of this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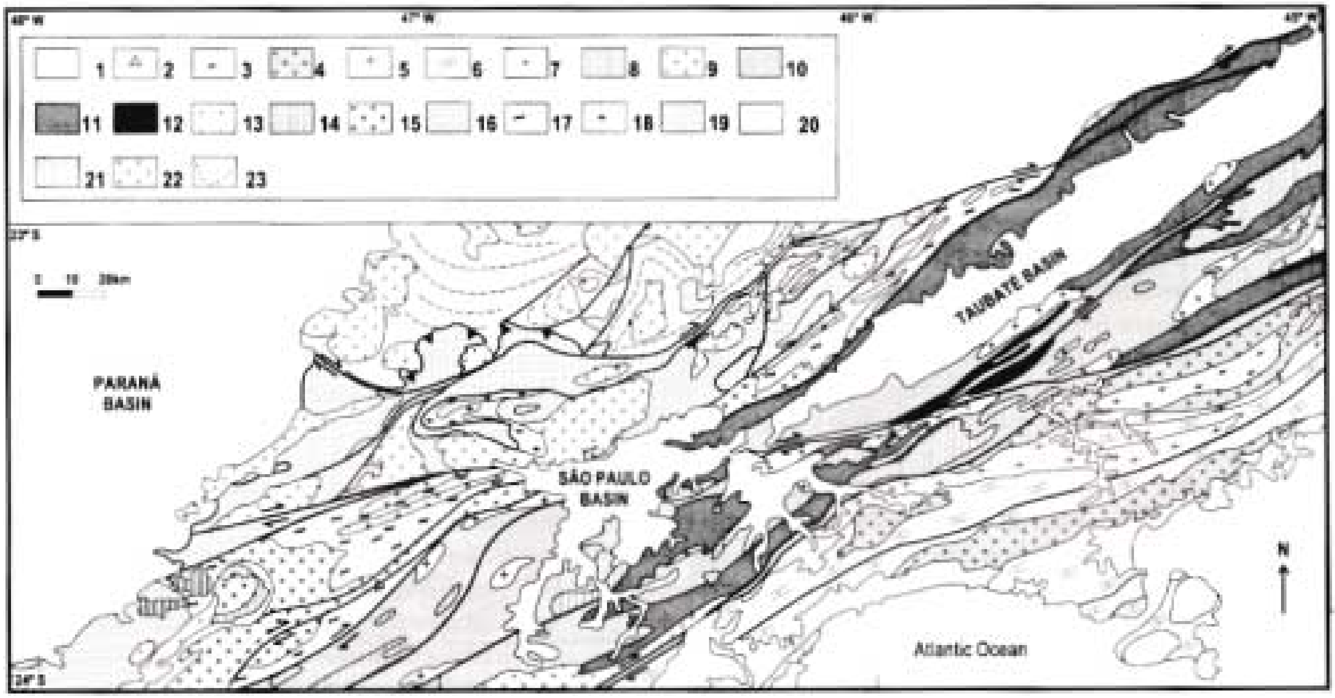


Fig. 2 - Tectonic map of South - Southeastern Brazil. 1. Phanerozoic covers. São Francisco Plate: 2. Cryogenian/Neoproterozoic III platform cover of the cratonic domain (Bambuí Group and correlates), 3. Tonian(T)/Cryogenian(Cr) passive continental margin sequences with remnants of oceanic crust (western Paranoá, Canastra and Carrancas groups, and eastern Macaúbas Group, Salinas and Ribeirão da Fôlha formations), 4. Statherian rift-related sequences (eastern Espinhaço Supergroup, western Araí Group and southern São João del Rei Group). Archaean/Paleoproterozoic: 5. Minas and Rio das Velhas groups, 6. Granite-gneiss basement. Amazonas Plate: Neoproterozoic III/Cambrian: 7. Platform cover of cratonic domain, 8. Varengian rift-drift sequences. Rio de la Plata Plate: 9. Paleoproterozoic granite-gneissic cratonic area and granulite-gneiss reworked Valentines Terrain (V), 10. Gravimetric record of the basement of the Paraná Basin. The gravimetric higher is related to the Rio de la Plata cratonic area. Central Goiás Terrane: 11. Metasedimentary sequence, 12. Ectasian metavolcanosedimentary sequence displaying rift-to-drift assemblages (Juscelândia, Indaianópolis and Palmeirópolis sequences, including the southernmost Statherian? calc-alkaline related Mossâmedes metavolcanosedimentary sequence), 13. Rhyacian/Orosirian stratiform mafic-ultramafic complexes (Barro Alto, Niquelândia and Cana Brava), 14. Archaean/Paleoproterozoic granite-greenstone belt Terrain. Tocantins orogenic system: 15. Cryogenian/Neoproterozoic III magmatic arc (Socorro-Guaxupé Terrane, including Iporá-Jaupaci area), 16. Cryogenian back-arc related metasedimentary sequence with oceanic crust remnants, 17. Anápolis-Andrelândia Terrane (volcano-plutonic arc and related metasedimentary sequence of Araxá Group, Aiuruóca-Andrelândia and Três Pontas-Varginha nappes), 18. Mara Rosa Terrane (juvenile island arc crust). Mantiqueira orogenic system: 19. Neoproterozoic III/Cambrian Serra do Mar terrane (microplate with Rio Negro – RN, c. 630 - 610 Ma, and Rio Doce – RD, c. 580 - 560 Ma magmatic arcs), 20. Juiz de Fora Terrane (Paleoproterozoic crust and Neoproterozoic arc-related metasediments of Embú and Paraíba do Sul complexes) as a microplate setting for (c. 595 - 580 Ma) Galiléia Magmatic Arc, 21. Foreland basin (c. 600 - 560 Ma Itajaí, Camaquã and Arroyo del Soldado groups), 22. Apiaí Terrane (Ectasian? Passive continental margin as a basement of a c. 610 Ma plutonic magmatic arc (Cunhaporanga, Três Córregos and Agudos Grandes batholiths), 23. Pelotas Terrane (c. 620 - 610 Ma plutonic magmatic arc), 24. São Gabriel Terrane (upper Cryogenian island arc), 25. Luis Alves Terrane comprising the Tijucas Belt (rift-to-drift Mesoproterozoic? metavolcanosedimentary sequences: Brusque and Porongos groups) and the Archaean/Paleoproterozoic Santa Catarina granulitic complex, 26. Curitiba Terrane (Paleoproterozoic gneisses and granulites) as a basement of the Rio Pien plutonic magmatic arc (615 Ma), 27. Grenville-Kibaran migmatites with western Neoproterozoic metasediments of Rojas Group, 28. Cabo Frio terrane.

Fig. 3 - Geological Map of the Serra do Itaberaba/São Roque groups and the Embu Complex. 1. Phanerozoic covers. 2. Neoproterozoic III/Cambrian Pico do Itapeva Formation. Serra do Mar Terrane: 3. Peraluminous granites (c. 500 - 540 Ma), 4. Metaluminous granitoid and charnockitic suite, 5. Sil-Grt-bearing migmatite with quartzite resisters, 6. Metaluminous migmatite. Juiz de Fora Terrane: 7. Shear zone-related metaluminous granite, 8. Peraluminous granite, 9. Hbl-Bt-bearing granitoid. Embu Complex: 10. Shelf-related quartzite and schist-quartzite rhythmic succession, 11. Metavolcanosedimentary sequence, 12. Archaean-Paleoproterozoic grey gneisses. 13. A-type granites (c. 575 - 580 Ma). Apiaí Terrane: 14. Peraluminous granites (c. 600 Ma), 15. Metaluminous granitoids (c. 610 - 605 Ma), 16. Terrigenous turbidite deposits of Açungui Group, 17. Schist and gneiss. São Roque Group (back-arc basin?): 18. Upper continental and shallow water quartzite, limestone bioherm and metavolcanic rock association, 19. Metarhythmite, quartzites, metalimestones and metabasaltes with pillow-lava structure (c. 610 Ma), 20. Serra do Itaberaba Group (Ectasian). Socorro-Guaxupé Terrane: 21. Peraluminous granite (c. 625 Ma), 22. Hbl-Bt-bearing granitoids (c. 630 Ma), 23. Sil-Grt-bearing migmatite of Piracaia Complex.

metamorphic event, first related to the Mesoproterozoic Ectasian by Rb/Sr isochrons (Fuck *et al.*, 1989), has been the object of research and interpretation using U/Pb zircon ages. Neoproterozoic Cryogenian age have been obtained (Ferreira Filho *et al.*, 1994; Suita *et al.*, 1996), but the SHRIMP data corroborated the Ectasian age (1.3 Ga; Correia *et al.*, 1999).

None of the data obtained suggest that these rocks do not belong to the São Francisco Plate. Proterozoic mobile belts formed from 1.95 to 1.0 Ga surround the Archean cratons of Laurentia and Baltica. Contrary to the São Francisco Plate, the Paleo-Mesoproterozoic tectonic evolution of the southern part of the Amazonas Plate (Rio Negro-Juruena Province and Rondonia-San Ignácio Province, Teixeira *et al.*, 1989) show successive accretion of juvenile, volcano-plutonic arc crusts. High-grade metamorphism with an average U/Pb SHRIMP age of 1.33 Ga (Tassinari *et al.*, 1999) precedes the Grenville age (1.25 Ga - 900 Ma) Sunsas Orogeny. Slightly after all these orogenic events the rapakivi granite genesis either in Rio Negro-Juruena Province, or in Rondonia-San Ignácio Province took place as extensional, anorogenic episodes (Bettencourt *et al.*, 1999). Therefore, the correlation between the eastern part of the Laurentia Plate and the western part of the Amazonas Plate (Sadowski and Bettencourt, 1996) is well constrained.

Apparently the latest U/Pb results for Barro Alto Granulite and Juscelândia Gneiss, and the earlier Rb/Sr and Sm/Nd data from the Moçâmedes arc-related rocks strongly suggest that these sequences fit better in the Laurentia-Amazonas Plate than the Mesoproterozoic Rodinia Supercontinent.

The Apiaí Terrane

Other Ectasian metavolcano-sedimentary sequences probably belong to the northeastern boundary of the Rio de la Plata Plate. They include the rocks of the Serra do Itaberaba Group and the overlying São Roque Group (Juliani, 1993) as well as volcanic units that show the transition from rifting to the formation oceanic crust such as the Perau Formation (Piekarz, 1981) and the Abapã Formation (Reis Neto, 1994), both basal units of the Açungui Supergroup (Campanha and Sadowski, 1999).

The Serra do Itaberaba Group (Fig. 3) consists of quartzite beds formed in coastal marine margin of a rift system and two major lithostratigraphic units (Juliani, 1993; Juliani and Beljavskis, 1995; Juliani *et al.*, 1996). The basal units consist of metavolcaniclastic rocks, metabasalt and metatuff, calc-silicate rock, graphite-bearing schist, iron formation units and metavolcanic rocks varying in composition from intermediate to acid. There occur hydrothermally altered tholeiitic basalt flows of the N-MORB type associated with a deep-water marine rock assemblage. The transition from tholeiitic to calc-alkaline volcanism is shown by meta-andesite, metarhyodacite and metarhyolite magmatism, which seems to suggest subduction of oceanic crust. This lithostratigraphic unit is overlain unconformably by Mn-rich and Ca-rich metapelite beds, metavolcaniclastic rocks, carbonaceous rocks and an upper Al-rich metapelite zone, locally with intercalations of amphibolite, metarhyolite and black, fine-grained tourmaline-bearing rocks. This upper unit shows the transition to shallow marine conditions and may be related to the beginning of the closure of the ocean basin.

The São Roque Group (Fig. 3) is dominated by three major volcano-sedimentary sequences (Bergmann, 1988). The first sequence consists of metavolcanic basic rocks (including tholeiitic basaltic flows with pillow structures) suggestive of a shallow water environment associated with carbonate deposition and the formation of stromatolitic bioherms (up to 800 m thick). These beds are followed by cyclic, millimetric to metric alternations of meta-arkose and quartzite, locally with metaconglomerate, and metapelite (at least 1000 m thick), with primary structures (wave ripples) representing continental slope to shelf environments. Metacalcareous lenses and a thicker quartzitic rock sequence may occur in the SW. The upper unit (up to 1600 m thick) consists of meta-arkose with cross-stratification and flaser bedding structures. It contains two intercalations of metabasic rocks, the first with pillow lava and the second with vesicular volcanic flows. A well-sorted quartzite suggestive of offshore bars comprises the upper beds. Although still uncertain, the latest U/Pb data point at a Neoproterozoic III (610 Ma) age for the submarine basaltic flows (Hackspacher *et al.*, 1999). On the whole, the volcanic and sedimentary process could represent a back-arc basin.

The Açungui Supergroup (Campanha and Sadowski, 1999) apparently represents preserved paleogeographic zones of a passive continental margin (Fig. 4). The main metavolcano-sedimentary sequences, which occur as windows below the basal stratigraphic level of the Açungui Supergroup, also have an Ectasian age both on U/Pb zircon and Rb/Sr whole-rock isochrons (Daitx, 1996, and E. Dantas, personal communication). The northwestern belt has a metasedimentary assemblage (up to 5000 m thick) suggestive of a "tropical" shallow shelf environment. The depositional sequence began with a marine transgression that deposited pelite-carbonate (Souza, 1990), which was interrupted by a regression represented by restricted deposits of shelf sands. An upper transgression was related to a sub-tidal carbonate platform containing bioherms of cone-shaped stromatolites (*Conophyton cf. C. garganicum*, Fairchild, 1977).

These units, suggest a drift-type episode, and overlap of about 2000 m of thick rift-type deposits that are mainly composed of metarkose, feldspathic metawacke and metafanglomerate, containing within-plate, ultrapotassic, alkaline metavolcanic and metavolcaniclastic layers, and flood basalt units. The Sm/Nd T_{DM} of these rocks falls in the Upper Archean (*c.* 2.5 Ga, Reis Neto, 1994). To the SW the platform gave rise to the development of a deep-water shelf (impure carbonate units, banded calc-silicate rocks and metapelite) with tholeiitic volcanism that grades in composition to that of shoshonitic metabasalt (Frascá *et al.*, 1990). At the top of this paleogeographic zone, and separated by a SE thrust, there occurs evidence of a regressive cycle in the form of some 2500 m of psammitic-carbonate sediments deposited on an epeiric platform. These sediments grade to carbonate facies from a shelf-ramp environment within the influence of storm waves (Campos Neto, 1983; Pires, 1991). Stratigraphicaly bellow the transgressive sequence that consist of about 1150 m of

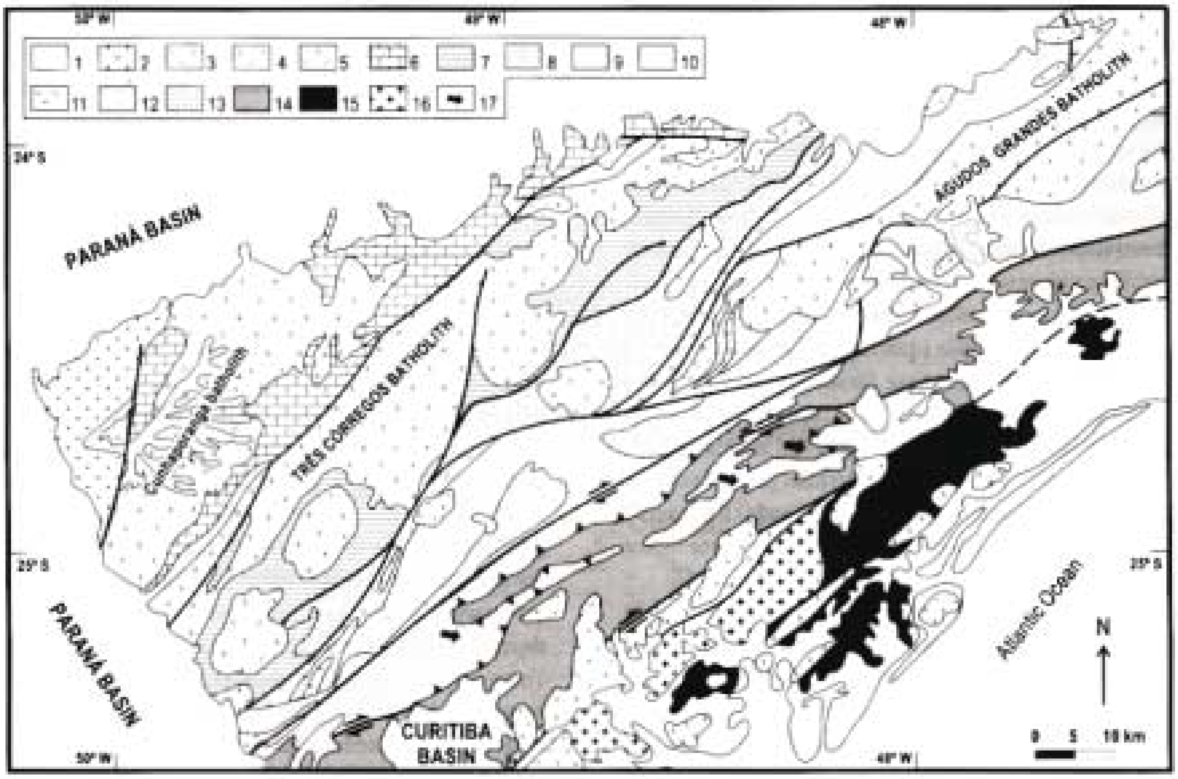


Fig. 4 - Major tectonic units from Açungui Supergroup. 1. Phanerozoic cover, 2. Foreland basin (*c.* 540 Ma), 3. A-type granitoid (*c.* 590 Ma), 4. Hbl-Bt-bearing porphyritic syn-to-late-orogenic granitoid (*c.* 615 - 605 Ma). Apiaí Terrane: 5. Flysch-type deposits (Iporanga Formation), 6. Carbonate platform (Itaiacóca Group), 7. Distal carbonate platform (Água Clara Formation), 8. Shelf-break carbonate ramp (Lageado Subgroup), 9. Deep-water turbidites (Votuverava Formation), 10. Rift-like volcano-sedimentary sequence (Ectasian Perau Formation, and the northern Abapã Formation), 11. Schist and gneiss from unknown environment. Juiz de Fora Terrane: 12. Schist and gneiss from Embú Complex. Curitiba Terrane: 13. Carbonate-psammitic passive continental margin deposits (Cryogenian?, Capirú and Setuva formations), 14. Upper Rhyacian basic-intermediate gneiss. Luis Alves Terrane: 15. Metasediments from Tijucas Belt, 16. Santa Catarina granulite complex (Archaean-Paleoproterozoic). 17. *C.* 600 Ma tectonic transport.

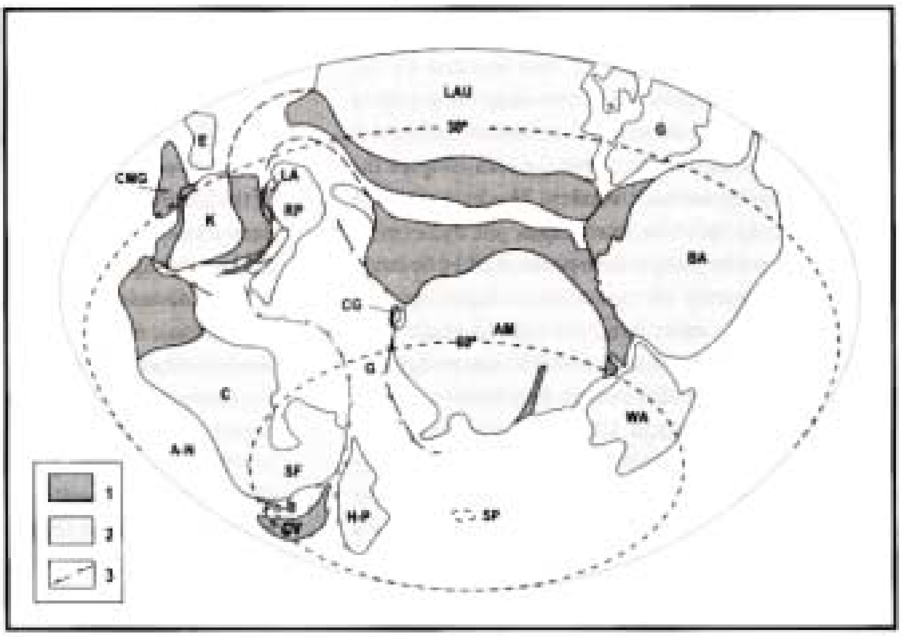


Fig. 5 - A modified view for the Rodinia reconstruction (after Weil et al., 1998).
The shape of Amazonas, São Francisco, Rio de la Plata and Western Africa continental segments of plates is modified. Speculative position of the postulate Hoggar-Potiguar Plate and Cariris Velhos Terrane. 1. Grenville orogens; 2. Continental plate segments; 3. Tonian rift lines. AM - Amazonas, BA - Baltica, C - Congo, CV - Cariris Velhos, CMG - Coates Land-Maudheim-Grune-Hogna Province, E - Ellworth-Withmore Mountain Block, G - Greenland, H-P - Hoggar-Potiguar, K - Kalahari, LAU - Laurentia, RP - Rio de la Plata, WA - Western Africa. AN - Arabian-Nubian Ocean, CG - Central Goiás Terrane, CV - Cariris Velhos Terrane, G - Goianides Ocean, LA - Luis Alves Terrane, PhB - Pharusian-Borborema Ocean, SP - South Pole.

metasandstone beds with lenses of oligomictic metaconglomerate grading upwards to thin beds of metarhythmite there occur turbidite beds to the E that were deposited in a channel-lobe transition zone gradational to the proximal depositional systems to the W. A wide deep-water basin plain developed to the SW where large amounts of pelite (slate and phyllite) with thin, fine-grained, metasandstone beds suggestive of rhythm-enhanced distal turbidity currents were deposited.

The stratigraphic floor of the continental rise represents a shallow water rift basin (quartzite and metalimestone). It grades to a deep-water zone (graphitic metapelite and metachert) with remnants of oceanic crust (metabasalt from tholeiitic, sub-alkaline, T-MORB type), and associated iron-formation units, to high-K quartz-sericitic rocks of volcanic provenance. Isotope stratigraphy on Sr, C and O from calcite and barite minerals, show a Mesoproterozoic marine source (Daitx, 1996).

A narrow belt of metarudite beds occurs at the boundary between the shelf-ramp carbonates and the deep-water basin. They consist of lenses of polymictic metabreccia and metaconglomerate overlying metarhythmite beds (Iporanga Formation). Characteristically these metarudite sediments contain angular fragments and rounded pebbles of surrounding host rocks and already deformed phyllite, quartzite and some igneous rocks; they are devoid of limestone clasts. This means that the Iporanga Formation could represent an orogenically controlled deposit (*e.g.*, flysh). The metabasalt flows that occur within this formation have calc-alkaline affinity in agreement with the model for compressive tectonic regimes.

Although the major structural framework had been built during the Neoproterozoic orogeny, the remaining metamorphic mineral assemblage locally showing kyanite ghosts (Juliani, 1993) was almost destroyed by a lower-pressure main foliation, and it was crosscut by the intrusion of the Neoproterozoic syn-orogenic granitoid batholiths. Among the majority of such rift-to-drift basins some of the volcanic layers show the transition to the plate convergent tectonic setting. The calc-alkaline volcanism marks the whole stratigraphic pile that extends into the uppermost compressive phase marked by basin in-fill. Unfortunately the age of this orogen remains uncertain. On the other hand the Sm/Nd whole-rock isochron for a metadolerite and Pb/Pb isotope data for carbonate rocks from shallow and deep-water platform environments of the Açungui Supergroup strongly suggest the Stenian-Tonian transition (*c.* 1.0 Ga) for the first metamorphic recrystallisation (Reis Neto, 1994; Poidevin *et al.*, 1997). Thus a Grenville Stenian age would be a coherent date for the closure of an Ectasian ocean crust.

The São Francisco Plate Margins: paleotectonic approach

The beginning of the Neoproterozoic is recorded in the São Francisco, Congo and Amazonas plates by a broad extensional tectonic regime related to the Tonian taphrogeny that led to the break-up of Rodinia. The paleomagnetic reconstruction of Rodinia (Weil *et al.*, 1998) left a Panthalassan-size ocean on what remained of the Earth's surface. The large Arabian-Nubian Ocean in the western part of the Congo Plate, and the probable South Pole connection with Pharusian Ocean in the western part of the West African Plate are some pieces of this jigsaw. The Goianides Ocean resulted from this former Tonian break-up, and the rifting apart the Amazonas Plate from the Congo-São Francisco Plate (Fig. 5), as well as the Rio de la Plata Plate from their African counterpart and Laurentia.

Farther to the N of the São Francisco Plate, extensional regimes (1.1 Ga - 970 Ma) immediately followed the end of the Grenville Orogeny in the Amazonas Plate, as well as in the Cariris Velhos Terrane (Brito Neves *et al.*, 1999). To the SE of the Amazonas Plate the records of this taphrogenic episode are the Santa Clara rapakivi granitic suite and the subalkaline and alkaline suites of the younger granite intrusives of Rondonia. They are coeval with the eruption of alkaline basaltic flows and the deposition of intracontinental rift sedimentary sequence. These anorogenic rock assemblages give a zircon age between *c.* 1.1 Ga and 980 Ma (Bettencourt *et al.*, 1999). At this time, the northern part of the Congo Plate was also subjected to a phase of dramatic extension that led to the opening of the Zambezi Ocean (Wilson *et al.*, 1993). The Rio de la Plata Continent may have been rifted apart from the western Kalahari taking part in a northeastward drift. The small fragment of the Central Goiás Terrane containing Mesoproterozoic rocks over an Archean-Paleoproterozoic framework could have been separated from the Amazonas Plate.

In the interior of the São Francisco Plate, taphrogeny-related magmatism occurred as within-plate microgabbro dyke swarms of tholeiitic and subalkaline affinities (Dossin *et al.*, 1993) showing a zircon age of *c.* 905 Ma (Machado *et al.*, 1989). Although geochronological data are not available, the rift-drift psammitic-dominated megasequence was established at the western (Paranoá and Canastra groups) and southwestern (Carrancas Group) extensional margins of the São Francisco Plate. They seem to be related to this Tonian taphrogeny. The Paranoá Group is an about 1600m thick megasequence limited by major unconformities. Its basal rudaceous layers locally overlie the Araí Group (Statherian), whereas its uppermost erosive boundary is overlain either by the Sturtian-age diamictite beds (Jequitaí Formation) or by post-Sturtian (Upper Cryogenian) carbonates of the Bambuí Group.

The depositional megacycles of the Paranoá Group (Faria, 1995), represent a marine succession of an internal NNW-SSE elongated rift valley separated from the western shelf-break by paleogeographic ridges like a marginal plateau. Carbonate rock interlayers, at least 200 m thick, occur in the upper transgressive unit, and grade laterally to thick bioherm build-ups. The bioherms occur on the ridges at the northwestern and southwestern edges of the basin as reef lines containing weakly branching, cylindrical and large columnar stromatolites (up to 2.5 m high) of the *Conophyton*-type (Cloud and Dardenne, 1973) developed in a subtidal environment. These carbonate-dominated sequences, which build a local barrier up to 2000 m thick (S of the Vazante Formation, Dardenne, 1981) may be related to strong subsidence due to rift-drift transition. To the W, the uppermost lithostratigraphic units of the quartzitic nappe of Canastra Group consist of pelitic-psammitic rhythmic and graphyte-bearing pelite beds (Campos Neto, 1984a). They were intensively cut by paraconglomerate channel deposits followed by thin sequences of pelitic-psammitic, meta-rhythmite and carbonate-bearing phyllite (Ibiá Formation, Barbosa *et al.*, 1970), similar to a turbidite depositional system. Moreover this upper terrigenous sequence shows a bimodal sedimentary source, either from the shelf deposits or the internal nappes, as well as a bimodal Sm/Nd signature (Fischel *et al.*, 1999; Pimentel *et al.*, 1999a, b). It may be placed on an inherited upper continental rise representing flysch-type deposits, related to the start of orogenesis (accretionary thrust wedge). Discrete diamictite deposits correlated with the Sturtian glaciation (Middle Cryogenian, Hoffman *et al.*, 1998) overlie the passive continental margin sequence and were covered by an extensive carbonate cap developed on a shelf platform (Bambuí Group) throughout the interior of the São Francisco Plate.

This assemblage represented a large "lower-plate"-type passive continental margin (Fig. 6). The pelitic micaschist and quartzite from the western upper nappes (Fig.2) were intruded by small serpentinite bodies associated with podiform chromite bodies and amphibolite, which were described as tectonically dismembered fragments of an oceanic crust, the Abadiânia-type ophiolite *mélange* (Strider and Nilson 1992). Other small belts of metavolcanic rocks have signatures typical of ocean floor basalt, and display a Tonian Sm/Nd whole-rock isochron age with $\varepsilon_{Nd} = +5.3$ (Fischel *et al.*, 1999). They form a chain of occurrences some 200 km long that represent the remnants of depleted mantle-derived oceanic crust.

On the other hand at the northeastern margin of the São Francisco Plate the Sturtian age glacial deposits attest to the beginning of the passive continental margin. Up to 100 m of massive diamictite beds (locally with tillite on striated pavements; Rocha-Campos and Hasui, 1981) of glacio-marine origin (Jequitaí Formation) grade eastwards toward the basin. In the basin there occur deposits some 10 000 m thick of glacial rocks reworked by subaqueous debris flows and turbidity currents (Macaúbas Group) that reach the deep-sea continental rise (Marshak and Alkmim, 1989; Uhlein *et al.*, 1998). The eastern assemblage is witness to the continental break-up. It corresponds to metamorphosed volcano-exhalative sediments with related amphibolite, derived from ocean floor basalt having a Middle Cryogenian age (Pedrosa-Soares *et al.*, 1992). The shallow-water carbonate and shale beds of the Bambuí Group, are an up to 400m thick platform units that overlie the Jequitaí diamictite beds (Fig.2).

An "upper-plate"-type margin seems to dominate the eastern part of the São Francisco Plate. It consisted of a narrow continental shelf, partially represented by a N-S transpression sliver of metapsammitic and metapelitic rocks (Dom Silvério Group), and by broad occurrences of metabasic and metavolcano-sedimentary sequences. The tectonic provenance of this metamorphosed basic and volcanic assemblage is still unknown, but it seems that it records an extensive Neoproterozoic episode constrained by Sm/Nd data (Pimentel *et al.*, 1997). It was tectonically imbricate with the grey gneiss of the Mantiqueira Complex of Paleoproterozoic age (Figueiredo and Teixeira, 1996).

The occidental and oriental passive continental margins of the edges of the São Francisco Plate were developed diachronically and display an asymmetrical shape (Fig. 7). The basal sedimentary layer of the narrow oriental margin (Sturtian diamictite beds) correlates with the uppermost stratigraphic level of the wide occidental passive continental margin.

The Tocantins Orogenic System

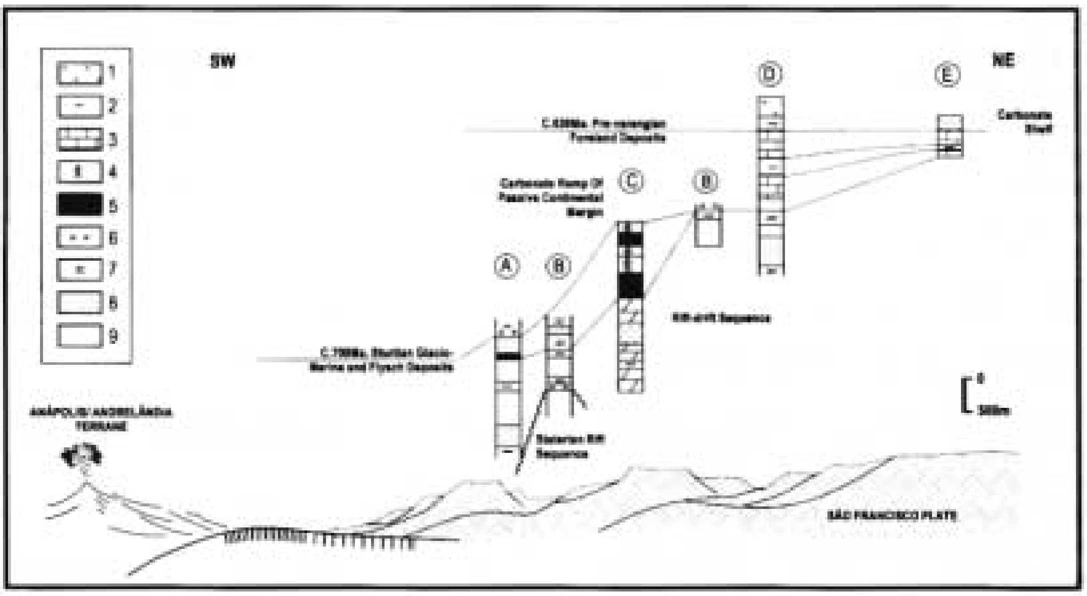
Outline of the orogens

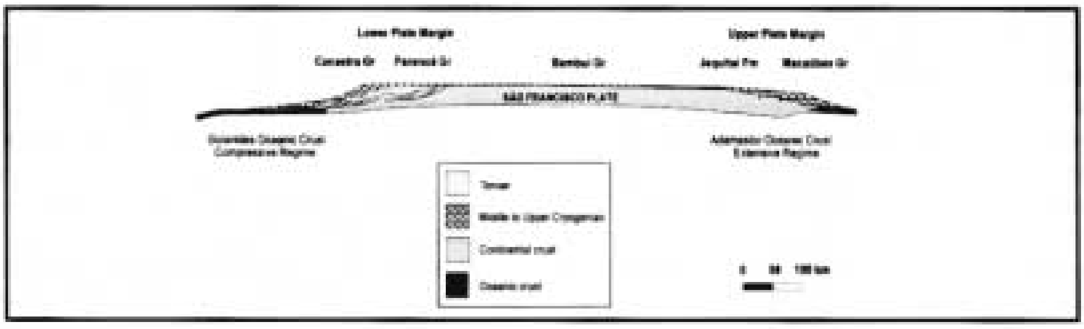
The Tocantins Orogenic System is related to the closure of the Goianides Ocean. It records a long-lived (270 Ma) plate convergence motion presupposing a large oceanic basin. The plate convergence process led to magmatic arc accretion and docking from the Upper Tonian up to the beginning of the Neoproterozoic III, when the main collision of the Central Goiás Terrane and Rio de la Plata Plate against the western São Francisco Plate took place.

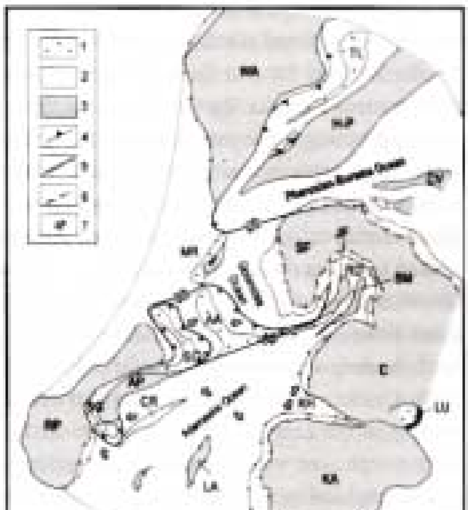
The Goianides Ocean encompasses many terranes. The ancient Central Goiás Terrane is a drift fragment related to the Tonian taphrogeny. The others are mainly orogenic terranes related to a set of subduction zones (Fig.8). The main compression kinematics of oceanic plate is related to W-dipping subduction zones. This kinematic framework is constrained by geological and geophysical records. The first is related either to the anatomy of a "lower plate"-type passive continental margin, or the collision-related E-driven extrusion of the high-pressure nappes. The geophysical records ascribe a W-dipping gravimetric model for the linear negative anomaly related to the suture zone (Marangoni, 1994), and the direct record of seismic P-wave and S wave velocity perturbation models (Van Decar *et al.*, 1995). This one displays large W-dipping low-velocity anomalies (for the P-wave) extending at least to 500 - 600 km depth. It is interpreted as the inherited thermal and chemical remnant of the original mantle-plume formed during the main metamorphic-collisional event, by coupling of the whole upper mantle and the lithosphere plate (Fig. 9). Fossil S-wave mantle anisotropy with the fast-polarization direction trend WNW-ESE could be correlated to the ancient Neoproterozoic plate motion (James and Assumpção, 1996).

The Upper Tonian Orogeny (900 - 850 Ma)

It was the first subduction-related orogen recorded by the accretion of the intra-oceanic Mara Rosa Island Arc. They are juvenile calc-alkaline volcanic belts and mantle derived metatonalite bodies from the Arenópolis and Mara Rosa regions (Pimentel *et al.*, 1992, 1997). Although not supported, Pimentel *et al.* (1999a, b) proposed an E-dipping subduction zone for the generation of this arc. Elsewhere, the Lufilian Arc in the Zambezi Domain, was essentially coeval.







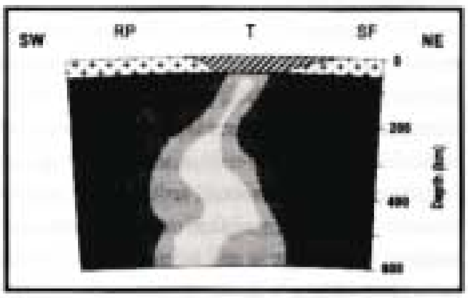


Fig. 6 - Stratigraphic and paleogeographic section through the western passive continental margin of the São Francisco.Plate. 1. Immature psammite and pelite, 2. Pelite, 3. Carbonates, 4. Stromatolite bioherms, 5. Graphite-rich layers, 6. Rudite, 7. Pelite and carbonate, 8. Psammite and pelite, 9. Psammite. A. Canastra Gr (Campos Neto, 1984), B. Paranoá Gr (Faria, 1995), C. Vazante Fm (Dardenne, 1981), D. Paranoá and Bambuí Grs (Dardenne, 1981), E. Bambuí Gr (Dardenne, 1981).

Fig. 7 - Scketch of E-W paleogeographic section through the western to eastern passive continental margins of the São Francisco Plate at c. 700 Ma.

Fig. 8 - Attempt view of the lost oceans in a c. 750 - 700 Ma paleogeographic outline. 1. Tonian/Cryogenian magmatic arcs, 2. Passive continental margin deposits and thinner continental crust domains, 3. Mesoproterozoic belts, 4. Continental plate segments, 5. Subduction zone with sense of dip, 6. Cryogenian sutures, 7. Ancient rift lines. AA - Anápolis-Andrelândia, AP - Apiaí, C - Congo, CG - Central Goiás, CR - Curitiba, CV - Cariris Velhos, H-P - Hoggar-Potiguar, JF - Juiz de Fora, KA - Kalahari, KH - Khomas Gulf, LA - Luiz Alves, LU - Lufilian, MR - Mara Rosa, RD - Rio Doce gulf, RP - Rio de la Plata, SG - Socorro-Guaxupé, SF - São Francisco, SG – São Gabriel, SM - Serra do Mar, TI - Tilemsi, WA - Western Africa.

Fig. 9 - Cross-section through the P-wave velocity perturbation models. Modified after Vandecar et al. (1995) with continental crust interpretation. The low-velocity anomalies (clear grey tonalitie) grade to the high-velocity anomalies (dark grey tonalitie). RP: Rio de la Plata Plate, T: Tocantins Orogenic System, SF: São Francisco Plate.

The Lower to Middle Cryogenian Orogeny

Two kinds of orogenies seem to have occurred at that time. The first is related to the docking of the Mara Rosa Arc, whereas the second is related to a wide arc generation. An asymmetric subduction-controlled non-continental-override-type collisional orogen could explain the docking of the Mara Rosa island-arc to the non-destructive western edge of the Central Goiás fragment. A collection of U/Pb, Sm/Nd and Re/Os isotope data obtained for the Central Goiás and in the Mara Rosa metasediments (Ferreira-Filho *et al.*, 1994; Correia *et al.*, 1996; Suita *et al.*, 1996; Pimentel *et al.*, 1998) support an age of 800 - 770 Ma for the metamorphic event. These values are not recorded in the metamorphic rocks from the passive continental margin. This metamorphism is mainly described as of low to medium pressure, high temperature-type, interpretatively related to the lateral docking of the hot arc crust. In spite of the poorly constrained data on the kinematics that controlled the main metamorphic structure, a dextral lateral motion is proposed, based on near N-S trending mineral and stretching lineations as well as shear-sense indicators (Fonseca, 1996).

Two sets of arc-derived rocks dominantly farther S of the Mara Rosa Arc are separated by the remnants of oceanic crust in a discontinuous NW-SE belt for at least 140 km in the Hidrolina-Morrinhos area (Figs. 2 and 8). The western arc assemblage comprises the calc-alkaline plutonic-volcanic rocks of the Iporá-Jaupaci area (Pimentel *et al.*, 1992), correlated by means of a NW-SE belt (partially covered by the sediments of the Paraná Basin) with the southernmost Socorro-Guaxupé Terrane (Campos Neto and Figueiredo, 1995). The Iporá-Jaupaci area is mainly composed of tonalitic-granitic rocks associated with metarhyolite, both derived from a juvenile source ($\varepsilon_{Nd(T)}$ positive). They are younger (760 Ma) and chemically more mature rocks ($K_2O/Na_2O>1.0$) than those of the Mara Rosa Arc (Pimentel *et al.*, 1992) indicating a partial intra-oceanic setting and the proximity of an older and thin continental crust. The continental arc basement became isotopically recorded farther to the S in the Socorro-Guaxupé Terrane (Janasi, 1999).

The Eastern Arc consists of calc-alkaline volcaniclastic and volcano-sedimentary derived metagreywacke. Metaluminous to slightly peraluminous granitoid plutons (780 to 700 Ma) having juvenile magmatic component have been described in this area (Pimentel *et al.*, 1999a). However a crust-derived magmatism developed to the W into pelitic and pelitic-carbonaceous schist and quartzite. They are peraluminous granite bodies and sub-volcanic felsite intrusives (790 Ma) having an older Sm/Nd signature (Pimentel *et al.*, 1992, 1999a). These contrasting magmatic environments are separated by the tectonic high of the Neoproterozoic III lower crust fragment related to the Anápolis-Itauçu granulite-granite-migmatitic complex. Characteristically, the rocks of this arc domain (Anápolis-Andrelândia Terrane) display bimodal peaks of Sm/Nd model ages, within the 1.95 up to 1.16 Ga edges (Pimentel *et al.*, 1999b; Janasi, 1999; Campos Neto and Caby, 1999b). It suggests a magmatic component from a juvenile source, and low crustal residence for the immature fan deposits, as well as the local presence of an older crust as a magma contaminant or as a sedimentary source.

The metasedimentary assemblage of the Araxá Nappe (Seer *et al.*, 1998) belongs to this tectonic setting. Nevertheless, the metapelite, quartzite and the few carbonate lenses from Pirinópolis area (Goiás), separated from the Anápolis-Andrelândia Terrane by a metamorphic jump and by a line of ocean crust remnants would be related to the passive continental margin deposits, rather than to the Araxá Group. On the other hand the western peraluminous granite and felsite bodies might have been derived from deep crustal fusion related to a strong back-arc stretching (Fig. 10-A), prior to the collisional episode, as previously proposed (Pimentel *et al.*, 1999a).

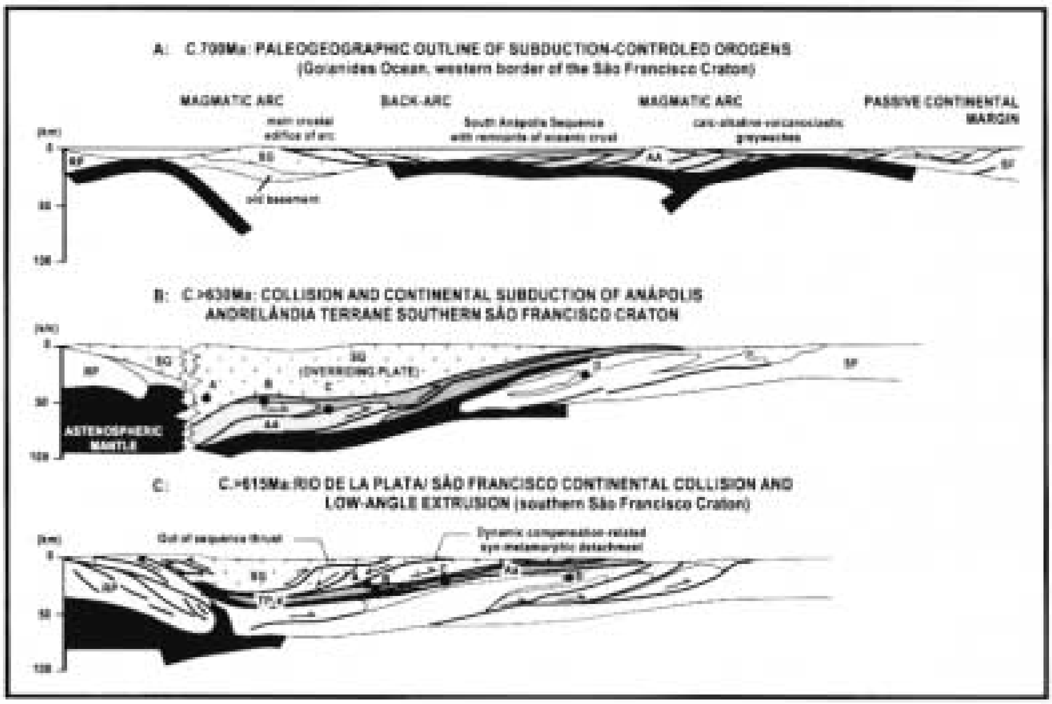
The Neoproterozoic III Orogeny

The roughly northeastwards motion of the Rio de la Plata Plate, as well as the amalgamated Mara Rosa Arc and Central Goiás Terrane, led these continental plate segments to collide against the western passive continental margin of the São Francisco Plate. This continental override-type collisional orogen is characterized by strong near-horizontal displacement of nappes transported over the outer domains of the parauthocthonous passive continental margin, up to the autochthonous Bambuí Platform. The shape of the lower-plate type passive continental margin favours the underthrust or continental subduction of the continental plate subjecting the lower nappes to medium-pressure metamorphic conditions. Except for the upper nappes, the metamorphism is typically reversed in each allochton (Simões, 1995; Campos Neto and Caby; 1999a). The shape of the collisional terranes, convex towards the passive continental margin (Fig. 2), may be explained by an indentor action related to the original shape of the western part of the São Francisco Plate (Fig. 8). This process may have controlled the oblique, sinistral convergence in the generation of the spoon-like flat geometry of ESE-displaced Araxá and Passos nappes (Fig. 2) associated with internal a-type, NE-verging, reverse folds (Seer *et al.*, 1998; Valeriano *et al.*, 1998). A collision-related peripheral foreland basin adjoins the stable edge of the continental subducted margin (Três Marias Formation).

The internal terranes of the Tocantins Orogenic System give robust U/Pb or Sm/Nd ages at *c.* 625 ± 5 Ma related to the main metamorphic recrystallization. This is in turn associated with the emplacement of syn-orogenic subduction-related porhyritic granitoid batholiths (Töpfner, 1996; Pimentel *et al.*, 1999a; Campos Neto and Caby, 1999a; Fischel *et al.*, 1999; Tassinari *et al.*, 1999). The outward propagation of the orogenic wedge and the out-of-sequence thrust sheets continued to 610 Ma (Janasi, 1999) up to their flat and ruptil emplacement over the authocthonous plataform units at *c.* 600 Ma (Valeriano, 1992).

The Southern Tocantins Orogenic System

The southernmost part of the Tocantins Orogenic System (Fig. 11) consists of a flat-lying package of nappes displaced to the ENE. They represent a diachronic thick-



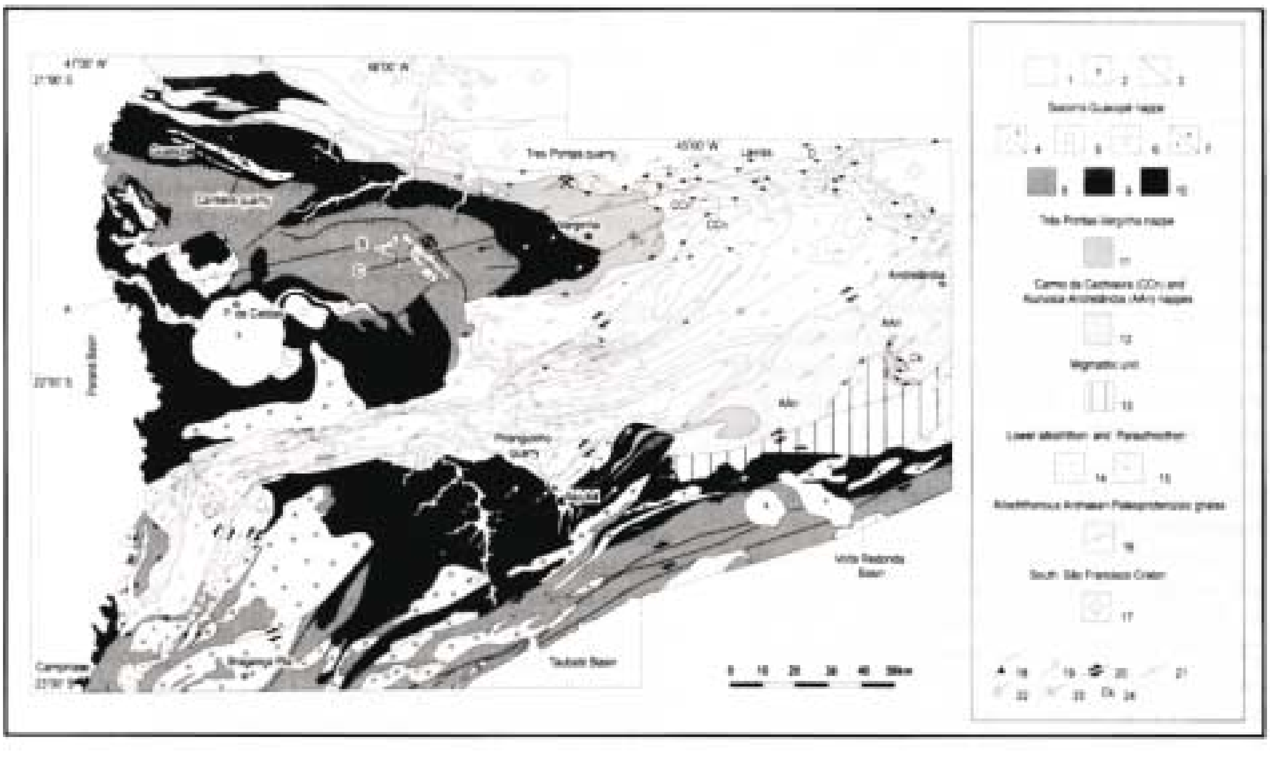


Fig. 10 - Sketches of plate kinematics evolution of the Southern Tocantins Orogenic System

Fig. 11 - Geological map of the nappe system south of the São Francisco Craton.
1. Phanerozoic deposits and basins, 2. Cretaceous alkaline plutons, 3. Pull-apart basins, 4. (small crosses) K-granite, diorites and gabro (c. 580 Ma) and (crosses) shear zone related granites (c. 600 - 580 Ma), 5. Syenitic plutons (c. 612 Ma); 6. Mangeritic-Granitic Suite (c. 625 Ma), 7. (x) K calc-alkaline porphyritic granitoid (c. 630 Ma) and (small x) Charnockitic suites (c. 640 Ma), 8. Upper Migmatitic unit and the southern schist belt, 9. Middle Diatexitic unit, 10. Basal Granulitic unit, 11. Ky-Grt granulite, 12. Metapelite and metagreywacke. 13. Migmatite and tourmaline-bearing granite, 14. Allocthonous quartzite-schist assemblage, 15 Parautochthonous quartzite-schist assemblage, 16. Allochthonous Archaean-Paleoproterozoic gneiss: 17. South margin of São Francisco Craton: 18. Major thrusts; 19. Displacement vectors; 20. Strike-slip shear-zone; 21. Detachment zone, 22. Antiformal axis, 23. Synformal axis, 24. Carvalhos Klippen.

skinned and frontal growing nappe system that underwent a minimum displacement of 300 km, accounting for the deeper crustal levels that are progressively exposed in the western allochthons. The uppermost tectonic unit derived from the magmatic arc (Socorro-Guaxupé Terrane) displays high-pressure and high-temperature basal granulite. The high-pressure terranes are underlain by the Ky-granulite from the Três Pontas-Varginha Nappe, and the metapelite-metagreywacke with few eclogite relicts that form the Carmo da Cachoeira and Aiuruoca-Andrelândia nappes. The lower allochthonous units consist of medium-pressure metamorphic sequences associated with polymetamorphic orthogneiss (Luminárias-Carrancas and Lima Duarte nappes) which grade, in trailing imbricate fan-type thrusts, to the parautochthonous units (Campos Neto and Caby, 1999a) both from the passive continental margin.

The upper high-temperature Socorro-Guaxupé Nappe

The Socorro-Guaxupé Terrane is a giant allochthon (at least 10 km thick) showing a right-way-up crustal section of hot and partially melted layered crust. It comprises (Figs. 11 and 12) basal banded enderbite (Basal Granulite Unit) *partially derived from a Neoproterozoic near-juvenile* immature magmatic arc (Sm/Nd T_{DM} of *c.* 1.3 Ga). These granulite bodies grade upwards into a predominant grey to pink, metaluminous migmatite (Middle Diatexitic Unit) mixing towards their top with pelitic to semi-pelitic migmatite (Upper Migmatite Unit). From these migmatites longitudinal dextral strike-slip shear zones control a major *metamorphic span to greenschist metamorphic facies* towards the S (Campos Neto and Caby, 1999b). Various syn-orogenic deformed plutons show chemical and isotope signatures compatible with the evolved subduction-related magmatism of a continental setting. They are porphyritic rocks from the charnockitic suite that develop upwards to porphyritic granitiod batholiths. Janasi (1999) showed that the Nd T_{DM} record of the Socorro-Guaxupé Nappe defines coherent domains in which an old residual migmatitic granulite can be found (Nd T_{DM} >1.8 Ga), whereas in other domains no sign of basement may be ascribed (Nd T_{DM} of *c.* 1.5 Ga). Similar bimodal distribution (Fig. 13) accounts for Neoproterozoic magmatic arc-derived "primitive" metagreywacke (Nd T_{DM} around 1.2 Ga) and for old basement-derived metapelite and "evolved" metagreywacke (Nd T_{DM} between 2.2 and 1.8 Ga).

High-temperature throughout the crust section is the main characteristic of the Socorro-Guaxupé Nappe (Fig. 14 A). The garnet-ortopyroxene-bearing granulite from the base of the nappe underwent T = 750 - 870 °C and P = 11.5 - 14.0 kbar metamorphic conditions and gave a near isobar heating evolution toward 860 - 920 °C (Del Lama *et al.*, 1994; Campos Neto and Caby, 1999b). These metamorphic data are consistent with dry melting of residual granulite in deeper levels of the crust generating anhydrous mangeritic magmas (Janasi, 1997; Campos Neto *et al.*, 1988). The stratified behavior of the Socorro-Guaxupé Nappe is related to the higher thermal flow and controlled by the crustal level in which widespread anatexis took place (Janasi, 1999). Thus, the pink anatectic biotite-bearing granite that belongs to the Middle Diatexitic Unit was derived from metaluminous or higher Rb/Sr sources. They were generated by biotite dehydration melting at intermediate levels of the crust under *c.* 850 °C. On the other hand the supracrustal rocks from the Upper Migmatitic Unit (T = 800 °C and P = 4.5 kbar, Vasconcellos *et al.*, 1991) are related to the shallowest muscovite dehydration melting as a magma source for the garnet-biotite parauthocthonous granite. Sm/Nd and U/Pb results point that the age of this metamorphism is in the 630 - 625 Ma range. The former age is related to the syn-orogenic granitoid and basal garnet granulite. The latter is well constrained for the uppermost garnet-biotite granite and the intermediate crustal level anactetic pink granite and mangeritic gneiss (Basei, 1985; Janasi, 1999). The syn-orogenic (*c.* 630 Ma) batholiths comprise an extensive suite of hornblende-biotite bearing porphyritic granitoid plutons derived from high-K calc-alkaline melts (Janasi and Ulbrich, 1991). They are intrusive into the Serra do Itaberaba Group (northern part of the Apiaí Terrane). There they comprise porphyritic to equigranular tonalitic-granitic gneiss.

The basal thrust of the Socorro-Guaxupé Nappe comprises a thick sequence (up to 3000 m) with flat-lying syn-metamorphic foliation and a strong plane-linear fabric *bearing top-to-ENE shear sense indicators. A sole thrust* overlaps the high-pressure nappes as well as the lower allochthon related to the passive continental margin units, resulting in a displacement of at least 150 km. A syn-metamorphic detachment, locally accommodated by a NW-SE oblique sinistral strike-slip fault, accounts for the direct contact of the uppermost migmatitic unit above the basal *granulite where at least 5000 m of the Middle Diatexitic* Unit were omitted. Late metamorphic NE-SW displaced out-of-sequence thrusts are younger than the *c.* 625 Ma old mangerite. These structures drove the stretched magmatic arc terrane toward its shallowest depth at 610 Ma, testified by the upper level intrusion of the post-kinematics potassic syenite bodies (Janasi *et al.*, 1993).

The high-pressure nappes

The Três Pontas-Varginha Nappe is a thick sheet that crops out for 170 km parallel to its ENE displacement direction. The nappe package consists mainly of coarse-grained rutile-kyanite-garnet granulite, lesser impure quartzite, and minor calc-silicate rocks, gondite, lenses of metabasic rocks and rare sills of mafic-ultramafic rocks. Upwards the kyanite assemblage gives way to sillimanite-bearing granulite grading to migmatite. The underlying Aiuruoca-Andrelândia and Carmo da Cachoeira nappes consist mainly of a layered sequence of peraluminous rutile-kyanite-garnet-muscovite schist and dark-grey, massive garnet-biotite-plagioclase gneiss and schist. Small lenses of metabasic rocks grade to an eclogitic metamorphic assemblage. In like manner to the metasedimentary sequence of the Socorro-Guaxupé Nappe, the chemical and isotope characteristics of these rocks (Campos Neto *et al.*, 1990; Janasi, 1999), under both granulitic or amphibolitic metamorphic facies conditions, is characteristically bimodal (Fig. 13). The rocks from a pelitic source have Sm/Nd T_{DM} and U/Pb ages on detrital zircon at 1.9 Ga (Söllner and Trouw, 1997). The metagreywacke record Nd T_{DM} between 1.16 and 1.55 Ga, points to their Neoproterozoic juvenile volcanic arc source.

A coherent inverted metamorphic pattern is supported by these nappes (Fig. 14). Lower temperatures (650 °C) were attained under high-pressures (12 - 14 kbar) related to the decompression stage of eclogite conditions (660 °C - 17.5 kbar). Upwards the temperature increase to 700 °C (P = 15 kbar) on Ky-granulite to get 830 - 950°C in a near-isobar heating path. The syn-kinematics decompression related to the outer propagation of the out-of-sequence thrusting varies from 600 - 690 °C and 9 - 11 kbar (Campos Neto and Caby, 1999b). The syn-kinematic cooling of the high-pressure pelitic granulite shown by the closure temperature of monazite in U/Pb system occurred at 612 Ma (Janasi, 1999). It could be related to the outward propagation of the out-of-sequence thrust system.

The medium-pressure lower nappes and foreland orogen propagation

The psammitic rock associations that are related to the passive continental margin deposits occur as the lower allochthon and parautochthonous units (Fig. 11). The flat-lying sheet of the Luminárias-Carrancas Nappe shows minimum transport to the E of about 140 km, and it is composed mainly of white and green mica quartzite grading upward into well bedded to laminated quartzite interlayered with graphitic and aluminous metapelite (Paciullo, 1997). Polymetamorphic orthogneiss and migmatite also belong to the nappe package, and a Barrovian-type mineral succession was described.

In the parautochthonous units the ramp and flat thrust pattern prevail, involving basement orthogneiss and interbedded quartzite and grey phyllite. Locally the metapelite displays a metamorphic assemblage which accounts for medium to high-pressure conditions (P = 7 kbar and T = 500 °C) that may control the appearance of Zn-staurolite and kyanite in the absence of garnet and biotite (Campos Neto and Caby, 1999a).

To the E the main psammitic rock package with sillimanite-bearing metapelite is organized into a large, basement involved allochthon: the Lima Duarte Nappe (Fig. 15). This was also subjected to medium-pressure metamorphism reaching the granulite facies and late anatexis. Garnet-clinopyroxene bearing amphibolite (with IBC-type coronitic textures) and sillimanite-bearing gneiss show the metamorphic peak under Tmax = 700 °C and Pmax = 7 kbar. Nevertheless, the syn-metamorphic structures account for N-displaced unrooted thrust system showing a monazite U/Pb age of 570 Ma (Machado *et al.*, 1996). A thin-skinned duplex displaying the same displacement to the N overprints the Lima Duarte Nappe. Thus, the Lima Duarte Nappe seems to be related to the foreland propagation of transpressive deep structures related to a collisional orogen of the younger Mantiqueira Orogenic System. It represents the Ribeira/Brasilia belts interaction.

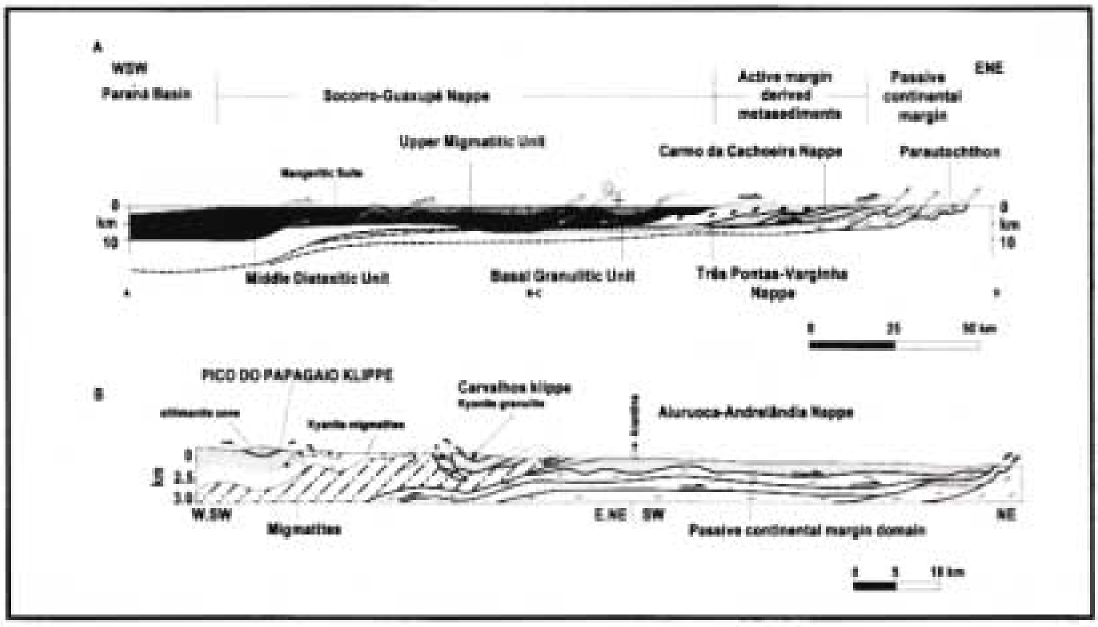
Tectonic evolution of the Southern Tocantins Orogenic System

The Southern Tocantins Orogenic System results from the agglutination of three major geodynamic environments during the Neoproterozoic III Orogeny. The western and uppermost tectonic setting is related to the Socorro-Guaxupé Magmatic Arc terrane, which may represent the deep crustal level of the Iporá-Jaupaci Arc developed during the Middle Cryogenian Orogeny. The high-pressure metasedimentary terranes (Três Pontas-Varginha, Carmo da Cachoeira and Aiuruoca-Andrelândia nappes) are the manifestations of an active margin setting as the source area for wacke deposits, probably upon a forearc-thinned crust belonging to the Anápolis-Andrelândia Terrane. The Tonian-Lower Cryogenian metasedimentary sequence from the southwestern passive continental margin of the São Francisco Plate comprises the metapsammitic-dominated, medium-pressure lower nappes and parauthocthon.

The high-pressure metamorphic conditions recorded by the kyanite-bearing granulite and eclogite from Três Pontas-Varginha Nappe and Aiuruoca-Andrelândia Nappe imply a low thermal gradient (*c.* 11 °C/km), which can only be achieved in an ocean closure (Spear, 1995) through a W-dipping subduction zone. The microplate segment represented by these nappes was subducted to a minimum depth of 60 ± 5 km. It is assumed that large amounts of oceanic crust and continental crust material from southern part of the Anápolis-Andrelândia Terrane were lost by subduction. Only a small part of the subducted slabs is preserved within the high-pressure nappes. Modern analogues can be seen in the Alpine Tertiary subduction of a large part of the paleo-European margin below the Apulian Plate (Schmid *et al.*, 1996).

An inverted metamorphic field gradient was brought about from the perturbed paleogeotherm to an upper steady state thermal gradient (20 °C/km) of the upper sillimanite-bearing granulite towards the base of the high-temperature Socorro-Guaxupé Nappe (Campos Neto and Caby, 1999b). This thermal pattern related to diachronic equilibrium at different temperatures throughout the metamorphic prism, as evidenced in the Indian Greater Himalayan Zone (Vannay and Grasemann, 1998), required the rapid emplacement of the upper hot allochthon onto the Três Pontas-Varginha Nappe. It resulted from the subduction-driven arrival of the Socorro-Guaxupé Terrane displaying, at its base, similar dT/dP gradient (Fig. 10-B).

At the shallow levels of the Socorro-Guaxupé Terrane (T = 820 °C and P = 4.5 kbar) the metamorphic high-thermal flow suggest a compressed paleogeotherm pattern (*c.* 50 °C/km). This steeper thermal gradient could be related to a lithosphere extension resulting from asthenosphere mantle upwelling with considerable amounts of basic magma underplating. These processes explain the origin of the widespread crustal magmatism found in the Socorro-Guaxupé Terrane, which occurred in a short-lived metamorphic peak confined at 625 ± 5 Ma. Though nearly contemporaneous the "cold" subduction scenario suggested by the kyanite granulite beneath the Socorro-Guaxupé Terrane attests to an initial different tectonic environment for both terranes. A roughly E-dipping



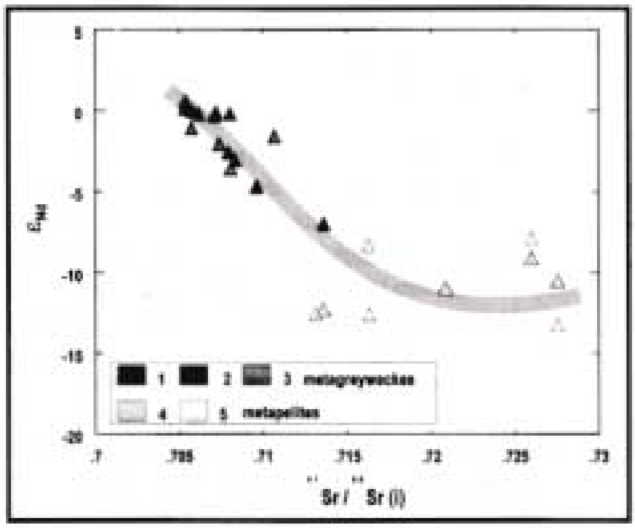


Fig. 12 - A- Cross-section across Socorro-Guaxupé, Três Pontas-Varginha, Carmo da Cachoeira and Luminárias nappes. B- Cross-section across the Aiuruoca-Andrelândia Nappe. Profile A located on figure 11.

Fig. 13 - $\varepsilon_{Nd(625)}$ X initial $^{87}Sr/^{86}Sr$ diagram for metasedimentary rocks (modified after Janasi, 1999). Metagreywackes: 1. Grt-Bt-Pl gneisses from Aiuruoca-Andrelândia Nappe, 2. (Crd)-Sil-Grt-Bt- gneisses from Middle Diatexitic Unit of Socorro-Guaxupé Nappe, 3. Ky-Grt-Pl granulites from Três Pontas-Varginha Nappe. Metapelite: 4. Sil-Ky-Gr-Kf granulite from Três Pontas- Varginha Nappe, 5. Grt-Bt gneiss from Upper Migmatitic Unit of Socorro-Guaxupé Nappe.

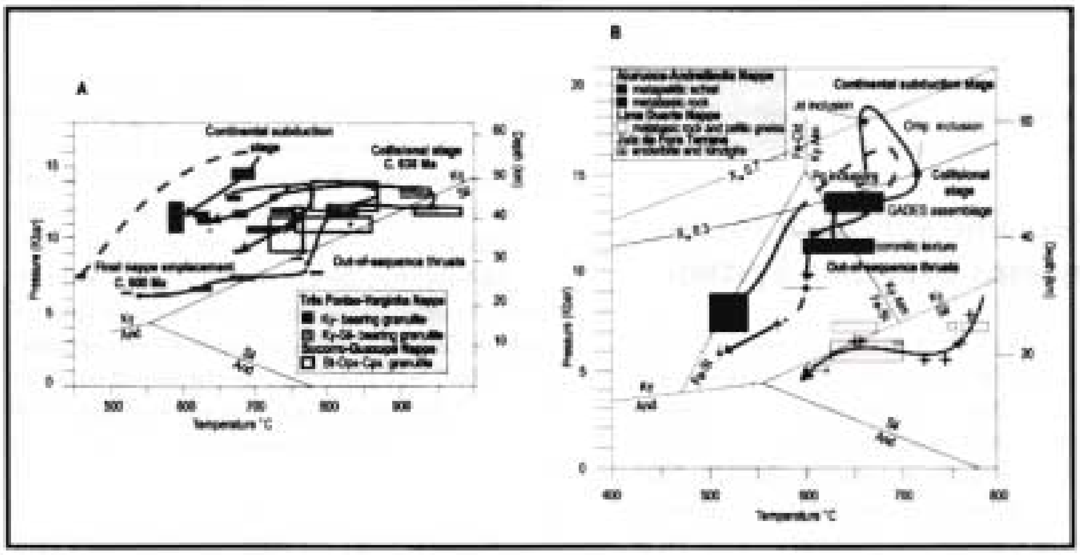


Fig. 14 - Metamorphic paths: A- Três Pontas-Varginha Nappe; B- Aiuruoca-Andrelândia Nappe and Lima Duarte Nappe.

subduction of an oceanic segment of the Rio de la Plata Plate might be suggested (Fig. 10-B). This subduction would be in agreement with the NW-displaced syn-metamorphic structures in the western area of the Socorro-Guaxupé Nappe, as well as in the steeper pattern of the isotherms.

The collisional outward propagation of the nappe pile (Fig. 10-C) submitted significant thinning was accompanied by a temperature decrease in each package of thrusted rock. The preservation of an inverted metamorphic gradient in equilibrium in the kyanite-field requires rapid uplift. This short-lived tectonic scenario began around 630 Ma and ended shortly before 612 Ma. Thus, part of the lower crust was uplifted from subduction prism and driven horizontally and thinned above the passive continental margin resulting in a mean cooling rate of about 15 °C/Ma. These subducted crustal slabs moved back up to the surface at a critical stage of the subduction while continuing underthrusting of the denser lithosphere (Chemenda *et al.*, 1995; Matte *et al.*, 1997). The hinterland-driven syn-metamorphic detachment may result from internal extension by dynamic compensation (Hodges *et al.*, 1996) in response to the possible building of mountain range the size of the Himalayas.

The post-orogenic history started early in the inner terranes at 610 Ma extending throughout 580 Ma up to 550 Ma (Pimentel, *et al.*, 1996a, b; Wernick, 1998). It was accompanied by a dramatic change in the character of the plutonic magmatism. A K-rich granitoid association was recognized conforming roughly to a NNE belt (Itú Belt, Vlach *et al.*, 1990). They are mostly undeformed high-K calc-alkalic hornblende and biotite-bearing porphyritic monzogranite bodies, syenogranite locally with wiborgitic texture, pink inequigranular monzogranite, muscovite and fluorine-bearing granite. Basic and intermediate rocks occur as enclaves and syn-plutonic dykes, and also as some plutons of K-rich diorite and K-syenite intrusives. Besides the large amount of magma contribution from crustal materials, an important contribution came from two fundamentally different mantle-derived magmas. A strongly oxidized magma, poor in basaltic component in the origin of the K-syenite; and the K-dioritic magmas with shoshonitic affinities (Janasi *et al.*, 1993). This post-orogenic episode seems to represent the reactivation and melt of the subcontinental lithosphere mantle involved in extension regime on the inner orogens.

Paleotectonic approach for the Mantiqueira Orogenic System

The above-mentioned phases of convergence are related to the global break-up of Rodinia and dispersion of their descendants, that commenced at *c.* 750 Ma, including the separation of Eastern Gondwana from the western margin of Laurentia (Park, 1994). Accordingly (and differently from the orogenic conditions of the Goianides oceanic realm), in the São Francisco-Congo Plate taphrogenic processes predominated during the Middle Cryogenian. The Sturtian glacial deposits were widespread and they were related to the main rifting stage in the Congo and Kalahari plates, as well as to the São Francisco Plate. Normally they were succeeded by passive continental margin settings and the post-glacial marine cap dolomite beds have covered the São Francisco Plate.

On the African side, the first stage of the Damaran extensional basin was recorded by the rift-related continental sediments of Nosib Group associated with alkaline bimodal volcanic rocks that are older than 750 Ma. On the western of Kalahari Craton eruption of felsic volcanic rocks have been dated at *c.* 745 Ma. The break-up of the supercontinent and opening of the Adamastor Ocean was constrained by the major marine transgression recorded by the deposition of Octavi Group, that contains basal volcanic interlayers giving U/Pb zircon ages between 758 - 745 Ma. Iron-rich diamictite beds (Chuos Formation) have been regarded as Sturtian, since they overlie a shelf ramp facies of the base of the Octavi Group. Farther to the S this oceanic opening process was diachronically younger. In the Gariep Belt the Sturtian diamictite beds (Kaigas Formation, *c.* 720 Ma) precede the passive continental margin succession (Hilda Subgroup) and the oceanic assemblage. Later still, the opening of the intracontinental branch of the Khomas Sea took place following the deposition of the uppermost glacio-marine diamictite beds (Ghaub Formation) considered as Varangian (*c.* 600 Ma) in age (Stanistreet *et al.*, 1991; Frimmel *et al.*, 1998; Hoffman *et al.*, 1998; Kennedy *et al.*, 1998).

As was seen previously, in the eastern part of the São Francisco Plate (South American side) the passive continental margin was starting to develop. This is interpreted from the Sturtian-related glacial deposits (Macaúbas Group), turbidite fans (Salinas Formation) and oceanic remnants (Ribeirão da Folha Formation) placed in the Neoproterozoic at the Middle Cryogenian (Pedrosa-Soares *et al.*, 1998). Nevertheless, an early oceanic basin (Charrua Ocean) that developed during the Early Cryogenian seems to precede the main opening of the Adamastor Ocean, the evidence for which can be seen locally to the E of the Rio de la Plata Plate. Several narrow and elongate (Fig. 8) small plate fragments (descendants of Rodinia) played an important role in the history of this orogenic system (Fig. 2).

The Apiaí Terrane comprising the Mesoproterozoic Ectasian series is overlain by carbonate shelf and terrigenous deep-sea turbidity fan from ramp and rise of a passive continental margin (Açungui Super Group described above). These rocks and the orogenic granitic batholiths show a Paleoproterozoic (*c.* 1.9 Ga) Nd (T_{DM}) signature (Sato, 1998). It is possible that the Apiaí Terrane had been connected with the northeastern part of the Rio de la Plata Plate after the Grenville Orogenic Cycle.

The Luis Alves Terrane conforming to an Archean-Paleoproterozoic lower-crustal fragment (Santa Catarina Granulite Complex) faced to the SE by a terrigenous and narrow shelf-type passive continental margin on which were deposited the Brusque Group and Porongos Group (Basei *et al.*, 1998). Mesoproterozoic Calymmian/ Ectasian Sm/Nd isochron ages are suggested for the basal volcano-sedimentary sequence (Basei, 1985). Farther to the S the metasedimentary belt overrides Paleoproterozoic aged gneiss, and the continuity of the Santa Catarina Granulite Complex was tectonically omitted. In the northern part of the Archean foreland domain, these sequences, and also the later granite intrusives display a Paleoproterozoic (2.0 Ga) Nd (T_{DM}) signature (Basei *et al.*, 1997).

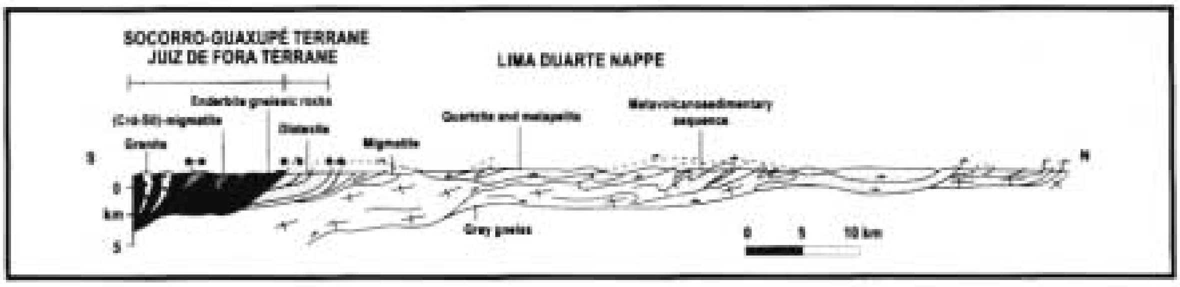


Fig. 15 - Cross-section through Lima Duarte Nappe

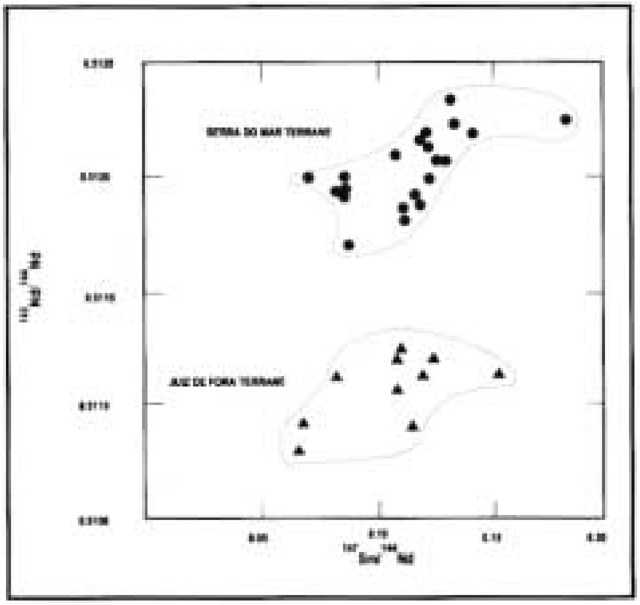


Fig. 16 - $^{143}Nd/^{144}Nd$ X $^{147}Sm/^{144}Nd$ relationships between Serra do Mar and Juiz de Fora terranes.

The Pelotas Terrane comprises a fragment of Paleoproterozoic crust, which was mainly recognized through the evolved arc-related granitoid plutons by the Sm/Nd studies (Basei *et al.*, 1997; Silva *et al.* 1999). In addition to this there is a small unnamed terrane composed of migmatite having a Kibaran age of metamorphism that has preliminarily been identified in southeastern Uruguay, E of the Pelotas Terrane (Preciozzi *et al.*, 1999).

The Curitiba and the Juiz de Fora terranes form a narrow and elongated crustal fragment with a general framework of juvenile Paleoproterozoic calc-alkaline intermediate plutonic rocks partially as a granulite belt. This belt became more evolved farther to the south (Sollner *et al.*, 1991; Sato, 1998; Basei *et al.*, 1997; Heilbron *et al.*, 1998). It seems that these terranes might be derived both from the rifting and break-up of the eastern part of the São Francisco Plate. Somehow the Juiz de Fora Terrane is regarded as belonging to the Paleoproterozoic juvenile within-oceanic to active margin accretion (Duarte *et al.*, 1997; Valladares *et al.*, 1997), which could be related to the Mineiro Mobile Belt, S of the São Francisco Craton (Figueiredo and Teixeira, 1996; Alkmim and Marshak, 1998). Thus the Middle Cryogenian break-up of the eastern part of the São Francisco Plate may have followed the lines of weakness related to an ancient Rhyacian suture zone.

Besides the widespread occurrence of granite, high-temperature metamorphic rocks from sedimentary provenance occur N of the Juiz de Fora Terrane, associated with a few stratiform bodies of mafic-ultramafic rocks. To the S, the Paleoproterozoic Rhyacian partially melted gneiss (Machado *et al.*, 1996), derived from plutonic rocks of calc-alkaline affinities that are tectonically associated with a metasedimentary assemblage (Paraíba do Sul Complex) consisting mainly of immature quartzite, sillimanite-garnet bearing gneiss and migmatite, and marble. An admixture of Paleoproterozoic Rhyacian and Statherian sources for these metasediments could be admitted based on $^{207}Pb/^{206}Pb$ age and Nd (T_{DM}), whereas some interbedded metabasic rocks display Neoproterozoic magma provenance (Valladares *et al.*, 1997). Farther to the S (Fig. 3), metasediments (Embu Complex) overlie a thin and narrow strip of Archean tonalitic-granodioritic gneiss (Tassinari and Campos Neto, 1988). They comprise three major units showing metamorphic grades from green-schist facies to higher amphibolite facies across steep and laterally displaced shear zones. These are (Fig.3) shelf-related quartzite beds, a distal rhythmic succession of pelitic schist and quartzite, and metavolcanic-volcano-sedimentary rocks (amphibolite, calc-silicate gneiss, metagreywacke) with restricted marble and immature quartzite (Fernandes *et al.*, 1990). Although isotope and geochemical data are not available, these sequences are supposed to have derived from a Neoproterozoic tectonically active environment.

The Curitiba Terrane is mainly composed of basic/intermediate banded gneiss and migmatite, and subordinately charnockitic gneiss, related to the Atuba Complex of Paleoproterozoic age (Basei *et al.*, 1992; Siga Jr. *et al.*, 1993). The metasedimentary cover consists of a marine regressive sequence consisting of shallow water carbonate beds (Capiru Formation) associated with a quartzite-schist assemblage (Setuva Formation), grading eastwards to a rhythmic deep-sea type terrigenous sequence. They have characteristics of an extensional basin related to a narrow and asymmetric rift reaching a passive continental margin (Yamato, 1999).

The Serra do Mar Terrane is mainly composed of migmatite and elongate and diffuse bodies of (garnet)-biotite-bearing granite with nebulitic and schilleren structures with variable strain rate. Sometimes they contain enclaves and/or resisters of quartzite, calc-silicate rock, marble and amphibolite. To the N of the State of Rio de Janeiro, preserved from the wide and deep strike-slip shear belt, three distinct crustal segments can be recognized (Sluitner and Weber-Diefenbach, 1989; Campos Neto and Figueiredo, 1995; Wiedemann *et al.*, 1997). They are the westernmost and basal supracrustal unit; the overlying gneiss-migmatite unit; and the eastern and upper granulite-granite-migmatite unit, displaying an inverted structural pile related to a westward nappe displacement. Immature quartzite, garnet-sillimanite-bearing quartzite and quartz-schist, marble and calc-silicate rock, and gneiss from a wacke provenance constitute the majority of the supracrustal units. Slices of metabasic and meta-intermediate rocks display protoliths chemically identified with high-Ti and low-Mg tholeiitic basalt and alkaline basalt associated with an arc-type calc-alkaline andesite (Sad and Dutra, 1988). These assemblages suggest an active continental margin. The gneiss-migmatitic unit consists of (cordierite-sillimanite)-garnet-biotite stromatic migmatite having a wacke-pelitic provenance, whereas the upper granulite-granite-migmatite terrane is mostly composed of syn to late-kinematic peraluminous (cordierite-sillimanite-garnet-bearing kinzigite) and metaluminous diatexite, normally associated with the enderbite series.

The contrasting Sm/Nd isotope signature (Sato, 1998) between the Serra do Mar and Juiz de Fora terranes shows mainly lower $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ ratios for the Serra do Mar rocks than those from the Juiz de Fora Terrane (Fig. 16). On the other hand, the Mesoproterozoic T_{DM} age that dominates all over the Serra do Mar Terrane concurs with the concept of an Africa rifting-apart.

The Cabo Frio Terrane comprises orthogneiss from diorite-tonalite-granite series associated with ortho-amphibolite slices of Paleoproterozoic age (Zimbres *et al.*, 1990; Schmitt *et al.*, 1999). The orthogneiss complex overrides a metasedimentary sequence mainly composed of paragneiss and quartzite, with subordinated lenses of calc-silicate rock and amphibolite (Heilbron *et al.*, 1982). These metasediments have a Mesoproterozoic Nd (T_{DM}) signature suggesting a Neoproterozoic age for their deposition.

The Mantiqueira Orogenic System

The geological scenario for the south-southeastern Atlantic coastal area of Brazil and Uruguay (up to the Serra de Mantiqueira in the continent interior) is a NE-SW trending orogenic system, mostly controlled by steep, strike-slip shear zones. It comprises a series of terranes that have diachronically collided against the just-assembled (630 - 620 Ma) São Francisco-Rio de La Plata plates, thus forming the orogenic system related to the closure history of the Adamastor Ocean.

Regional view on plate convergence in the Southern Mantiqueira Orogenic System

The plate convergence started with the closure of the small (?) Charrua oceanic basin. This event was manifest in the occurrence of metavolcanic and metaplutonic rocks of calc-alkaline affinity (Vila Nova Belt, Silva Filho and Soliani, 1987) showing a U/Pb on zircon age of 705 Ma and a $\varepsilon_{Nd(T)}$ rising to + 7.8 (Babinsky *et al.*, 1996). It was the first well-constrained plate interaction record for a Late Cryogenian intra-oceanic subduction-controlled orogen (São Gabriel Orogeny) in the southern part of the Mantiqueira Orogenic System.

Sedimentary and volcanic belts form the Brusque and Porongos groups, which underwent medium to low-pressure metamorphism associated with WNW verging collision-related structures. Although scarce and imprecise the geochronological data suggests that the metamorphic peak must have been reached around 630 Ma (Silva *et al.*, 1999) slightly before the emplacement of the late-kinematic granite batholiths (Basei, 1985). Many pieces of the plate interactions for this collision-controlled orogen seem to have been disguised by or concealed under the Phanerozoic Paraná Basin (Fig.2).

The Neoproterozoic III subduction-controlled Pelotas magmatic arc orogen (610 to 600 Ma) took place facing the southeastern edge of the Brusque and Porongos orogens. This arc comprises a series of calc-alkaline plutonic rocks and high-K porphyritic granites mingled with coeval migmatite, all displaying flat-lying foliation related to W-NW tectonic transport (Basei, 1985; Tommasi *et al.*, 1994; Fernandes *et al.*, 1995). Based on the extrusion vector of the collision structures and the eastward zonation toward the post-collision pink granite (*c.* 595 Ma), a southeastward-dipping subduction could be suggested.

The main collisional orogenic interaction *c.* 600 Ma was responsible for the juxtaposition of the Pelotas Magmatic Arc, the Luis Alves Terrane, and the southeastern boundary of Rio de La Plata Plate. The major strike-slip shear zones define the main collision boundary: a northern, dextral and a southern, sinistral shear zone that contained and controlled the emplacement of peraluminous and metaluminous granite intrusives. Impactogenic processes were related to the origin of the undeformed volcano-sedimentary basins that were associated with alkaline-peralkaline granitic intrusion. This process took place mostly in the Luís Alves and Curitiba terranes at *c.* 600 - 570 Ma (Siga Jr. *et al.*, 1997).

The development of the collision-related foreland basins at the southern part of the Luis Alves Terrane and the eastern part of the Rio de la Plata Plate are suggested by remnants of continental environments that extend up to deep marine deposits. Unconformity and volcanic episode separated these sedimentary environments. They are the Itajai, Camaquã and Arroyo del Soldado groups, which were intruded by granite and volcanic felsite (560 Ma), and which contain several Vendian species of palynomorphs. Their stratigraphic history was broken at 530 Ma, related to the age of the thin-skinned deformation (Basei *et al.*, 1997; Gaucher *et al.*, 1998).

The Central and Northern Segments of the Mantiqueira Orogenic System

The tectonic evolution of the central and northern segments of the Mantiqueira Orogenic System is related to the diachronous kinematics of oceanic plate convergence generating widespread magmatism. Collision and docking processes leading to the closure of the oceanic spaces between the major terranes controlled the geometry of this tectonic scenario. Thus the knowledge of the petrology and plutonic stratigraphy for the magmatic rocks (Söllner *et al.*, 1987; Söllner *et al.*, 1989; Offman and Weber-Diefenbach, 1989; Soares *et al.*, 1990; Söllner *et al.*, 1991; Janasi and Ulbrich, 1991; Figueiredo and Campos Neto, 1993; Gimenez Filho *et al.*, 1995; Machado *et al.*, 1996; Wiedemann *et al.*, 1997), and the knowledge about the structural framework (Heilbron *et al.*, 1982; Campos Neto and Figueiredo, 1990; Machado and Demange, 1990; Pedrosa-Soares *et al.*, 1992; Campos Neto and Figueiredo, 1995; Heilbron *et al.*, 1998; Ebert and Hasui, 1998) were taken as the principal tools to unravel the tectonic history.

The Rio Negro subduction-controlled orogeny (*c.* **630 Ma**) was the first record of the Central Mantiqueira Orogenic System of an orogen related to a plutonic magmatic arc. It was accreted on the Serra do Mar Terrane (Oriental Terrane of Tupinambá *et al.*, 1998) as a batholithic complex of gabbro-diorite-tonalite from high-Ca and low-K calc-alkaline magma, yielding an $\varepsilon_{Nd(600\ Ma)}$ of - 0.9 (Tupinambá *et al.*, 1998; Tupinambá, 1999). These rocks display a horizontal to northwesterly gently dipping foliation related to a low-pressure, amphibolite facies metamorphism. The blockage of the subduction and docking of the Rio Negro Orogen against the southeastern part of the Juiz de Fora Terrane was suggested herein. The docking process was laterally controlled by the development of variable steep to low-angle dipping mylonitic foliation with NE trending stretching and mineral lineation related to a main dextral displacement, leading to a local override of the Juiz de Fora Terrane upon the magmatic arc (Heilbron *et al.*, 1998; Tupinambá, 1999). This main tectonic edge ("the central tectonic boundary") was developed coeval with the generation of the post-arc metaluminous to peraluminous leucogranite intrusives (*c.* 600 Ma). The oblique juxtaposition of the arc and the Juiz de Fora Terrane released, farther to the N, a remnant of oceanic basin (Fig. 18).

The Paranapiacaba Orogen is related to a complex subduction and extensional tectonic setting that took place farther to the S attaining the domains of Apiaí and Curitiba terranes (Fig. 17). It was coeval or slightly younger than the Pelotas Orogen. This orogenic scenario is pictured by huge and elongated batholiths corresponding to syn-orogenic intermediate to felsic metaluminous, high-K calc-alkaline granitic series. Hornblende-biotite porphyritic quartz monzonite and monzogranite prevails upon the fine- to medium-grained, grey, biotite-bearing monzogranite. Tonalitic rocks are subordinate. They correspond to an evolved magmatism that reworked Paleoproterozoic lithosphere (Reis Neto, 1994; Harara *et al.*, 1997). The Rio Pien Magmatic Arc facing the Curitiba Terrane to the SE shows an age of about 615 Ma (Harara *et al.*, 1997). It slightly proceeded the 610 Ma age of the Agudos Grandes high-K calc-alkaline batholith facing the southeastern part of the Apiaí Terrane. Inequigranular to equigranular muscovite-biotite granite with porphyritic facies normally wrap up the terminations of the calc-alkaline batholiths. They are 605 - 600 Ma and are followed by the younger, post-kinematic (565 Ma), porphyritic biotite syenogranite with an "A-type" chemical signature (Janasi *et al.*, submitted). Farther to the interior of the Apiaí Terrane the westernmost Três Córregos and Cunhaporanga batholiths (Fig. 4) are associated with high-temperature and low-pressure amphibolite facies metamorphism of the country rocks and detachment structures could be recognized. In this domain U/Pb zircon ages are still poorly constrained and the preliminary data display strong variable values suggesting roughly a westward age decreasing from 610 Ma up to 570 Ma for the calc-alkaline granite. The metasedimentary sequences of São Roque Group that occur along the northwestern edge of the Apiaí Terrane (Fig. 3) contain at their base bodies of metabasalt derived from sub-alkaline tholeiitic volcanism, locally displaying pillow structures. The zircon U/Pb age for this submarine volcanic flow has the same value (610 Ma) of the magmatic arc suggesting a narrow back-arc basin (Hackspacher *et al.*, 1999). The collision between the Luis Alves and the Curitiba terranes and between the Curitiba and the Apiaí terranes led to the closure of the São Roque back-arc basin, amalgamating the continental fragments in the southeastern part of the Rio de la Plata Plate. In the Curitiba Terrane the collision led to strong shortening by low-angle, ESE transported ductile shear upon overall crust. It is associated with *c.* 600 Ma (Siga Jr. *et al.*, 1995) metamorphic slices that underwent geenschist to high-grade amphibolite metamorphism, attaining granulite facies metamorphism. Double vergence and contrasted tectonic regimes are suggested for the Apiaí Terrane. A steep dextral strike-slip shear zone controlled the major geometry of the structures normally disguising an early flat-lying foliation. At their southeastern boundary the flat-lying foliation is roughly related to an eastward thrusting transport.

Toward the hinterland the low-grade metamorphic shallow-dipping foliation is related to direct transport to the NW (Campos Neto and Basei, 1983, Juliani, 1993). The central magmatic arc seems to be related to many hinterland and foreland driven detachments, the major ones bordering the batholiths and displaying a strong strike-slip component. An alternative tectonic scenario is also considered: a high-angle, W-dipping subduction zone under the Curitiba Terrane gave way to an active continental margin setting (Rio Pien Magmatic Arc). Toward the hinterland strong lithosphere extension might be related to the generation of the huge batholith zone comprising the wide calc-alkaline plutons nearly coeval with peraluminous magma, both intrusive in older units from a passive continental margin. This lithosphere extension rises, on the plate interior, to rifting and break-up of the São Roque back-arc basin. For both tectonic scenarios the main collision-controlled orogen took place at 600 Ma and was followed by post-orogenic lithosphere extensional regimes around 570 Ma.

The main 600 Ma collisional episode related to the Paranapiacaba and Rio Negro orogens is seen elsewhere in the main magmatic arcs. It is manifest in peraluminous monzogranite and migmatite from the southern Juiz de Fora (Embu Complex) and the southernmost termination of the Serra do Mar Terrane.

To the N, this collisional amalgamation released an oceanic branch between the eastern São Francisco passive continental margin and the Juiz de Fora Terrane (Fig. 18), which was closed by this time. The beginning of the Araçuaí Orogeny took place with the growing of the Galiléia Magmatic Arc in an active margin setting W-facing the Juiz de Fora Terrane through E-dipping oceanic lithosphere subduction. The granitoid plutons of the Galiléia Batholith (Nalini Jr. *et al.*, 1998) show a relatively extensive compositional trend (tonalite-granodiorite-granite), containing amphibole and biotite as major mafic mineral phase, and grossular-rich garnet characterizing a deep crust (up to 10 kbar) magma crystallization. They have a zircon U/Pb age of 595 Ma (Nalini Jr. *et al.*, 1998) and they were followed by semi-circular plutons of the peraluminous Urucum Suite consisting of monazite, tourmaline, garnet and/or two mica granite-types. The Urucum granite intrusives are thought to be a collision-type suite having an age of *c.* 580 Ma (Nalini Jr. *et al.*, 1998). The collisional development of the Araçuaí Orogen (Pedrosa-Soares *et al.*, 1998; Ulhein *et al.*, 1998) is governed by foreland-driven nappes consisting of relatively high-pressure metamorphic rocks at the suture zone (kyanite-garnet-bearing schist). The cratonic edge was reached by the westward thrusting of the Espinhaço Range up to the thin-skinned behavior of the platform cover. To the S, the Abre Campo discontinuity (Fischel *et al.*, 1998) represented the suture boundary between the active margin of Juiz de Fora Terrane and the western Archean-derived thinned continental crust from the passive margin domain (Mantiqueira Gneiss). A ductile and oblique, dextral strike-slip shear zone delineates it.

The Rio Doce Orogeny developed early in the Serra do Mar Terrane, renewed as a microplate after the complete consumption of the oceanic lithosphere plate segment at the eastern São Francisco passive continental margin domain. It resulted from the final convergence leading to the closure of the Adamastor Ocean. Thus its record may be pursued, with variable intensity, throughout the Mantiqueira Orogenic System. As a variance of the Campos Neto and Figueiredo (1995) tectonic scenario, a V-shape oceanic branch must have been released to the NW of the Serra do Mar Terrane and the first record of its lithosphere resorption was the Rio Doce Magmatic Arc (Figueiredo and Campos Neto, 1993). This assumption could be supported by the W-directed low-angle extrusion of the roots of the magmatic arc (Fig. 19A). Elsewhere, on the African side, the structural patterns of the West-Congo Belt (Trompette, 1994), and farther to the S, the oceanic nappe of the Marmora Terrane (Frimmel *et al.*, 1998), both displaying E-displacement toward the cratonic areas, are the converging diagnose for a W-dipping oceanic lithosphere subduction.

This process has been constrained at 575 Ma in the Gariep and Damara belts (Frimmel *et al.*, 1998). Based upon the chemical and compositional zonation of the main plutonic rock-types of the Rio Doce Magmatic Arc, Figueiredo and Campos Neto (1993) also suggested a W-dipping subduction-controlled regime. The steeping of the paleo-isotherms depicted by the close relations between the plutonic calc-alkaline rocks and peraluminous garnet, sillimanite and/or cordierite-bearing diatexite (Wiedemann *et al.*, 1986; Rego, 1989; Campos Neto and Figueiredo, 1990; Fritzer, 1991; Campos Neto and Figueiredo, 1995) might be related to the asthenosphere upwelling below the arc ascribed by subduction at both sides of the northern part of the Serra do Mar Terrane.

The calc-alkaline plutonic rocks of the arc comprise expanded suites of norite-enderbite-charnoenderbite and gabbro-diorite-tonalite-granodiorite forming elongate batholiths. They are mostly a low-K, high-Al calc-alkaline series, for both the western and the easternmost suites. More evolved K-rich tonalite and enderbite also occur within a NE oriented trend (Sluitner and Weber-Diefenbach, 1989; Rego, 1989; Figueiredo and Campos Neto, 1993; Wiedemann *et al.*, 1997). These gneissic batholiths have intrusive contacts into metasedimentary and migmatitic sequences. The deeper granulitic units generally show diffuse boundaries with the country rocks, although containing xenoliths of peraluminous migmatites. The life span of this subduction-related plutono-metamorphic orogen is robustly constrained between 580 - 565 Ma (Söllner *et al.*, 1987, 1989, 1991). To the S, the Rio Negro Magmatic Arc was being renewed by the tonalite-granodiorite-granite suite from Serra dos Órgãos Batholith (Machado and Demange, 1994). It is a thick stratiform-shaped and high-Ca calc-alkaline massif displaying mostly magmatic flow foliation, and having an age of 560 Ma (Tupinambá, 1999).

A widespread series of syn-kinematic peraluminous diatexites, and I-type granite and mangerite suites occur in the Serra do Mar Terrane. They are related to the collision closure of the oceanic remnants. An apparently conflicting low-angle direction of displacement is found in the main rock package of the Serra do Mar Terrane. To the N, thrusting and the building of an inverted crustal pile followed a westward high-temperature ductile shearing. Farther to the S, E-verging structures are in agreement with the E-verging, near-isoclinal recumbent folds (Heilbron *et al.*, 1982) and E-displaced low-angle ductile shear zones and thrusting of the Cabo Frio Terrane (Machado and Demange, 1990). The Cabo Frio Terrane resembles a small fragment of passive margin related to a southwestern promontory of the Congo Plate.

However these structures record diachronic collision episodes. The voluminous ultra-high metamorphic, crustal magma generation and its upper related migmatitic fingers took place coeval with the western overriding of the Serra do Mar Terrane, mostly at its northern extension. They were related to the closure of the V-shaped oceanic remnant between Serra do Mar and Juiz de Fora landmasses. This collisional episode is recorded in the Neoproterozoic III -

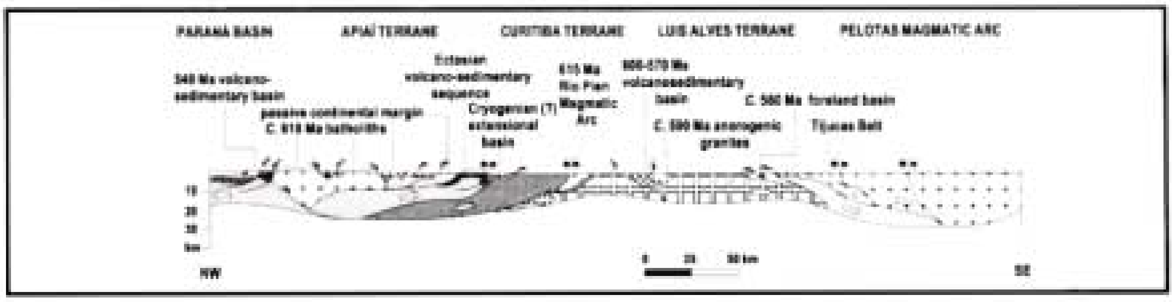
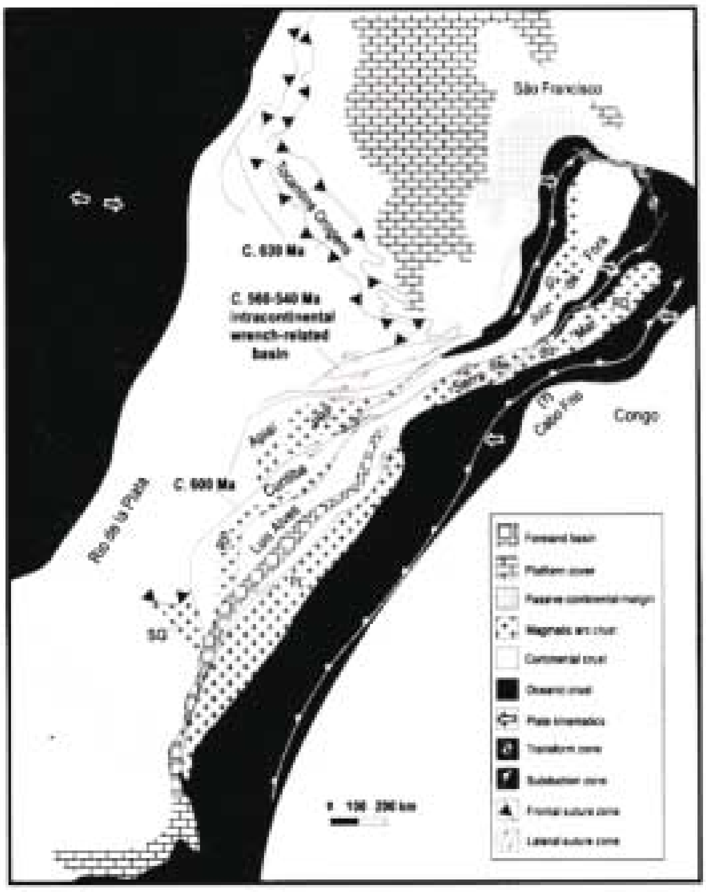
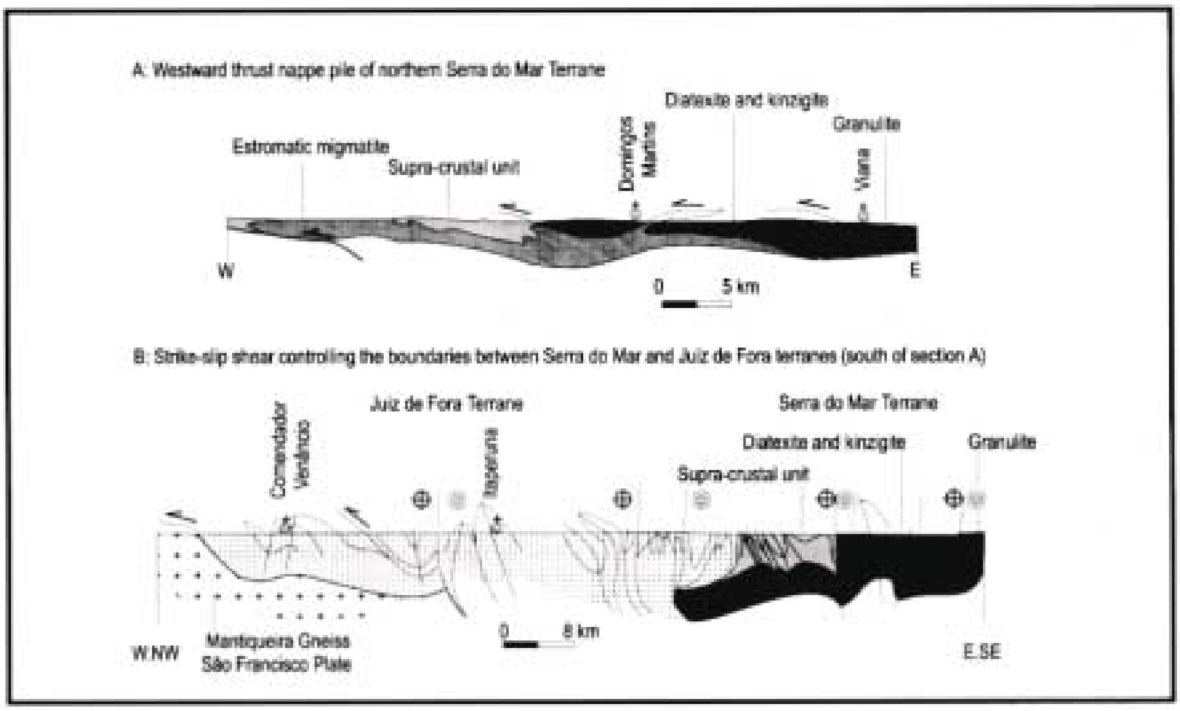


Fig. 17 - Profile through Apiaí and Curitiba terranes.

Fig. 18 - Vestiges of the lost Adamastor Ocean at c. 590 Ma: Paleogeographic sketch of oceanic plate interactions

Fig. 19 - Cross-sections across Serra do Mar and Juiz de Fora terranes.





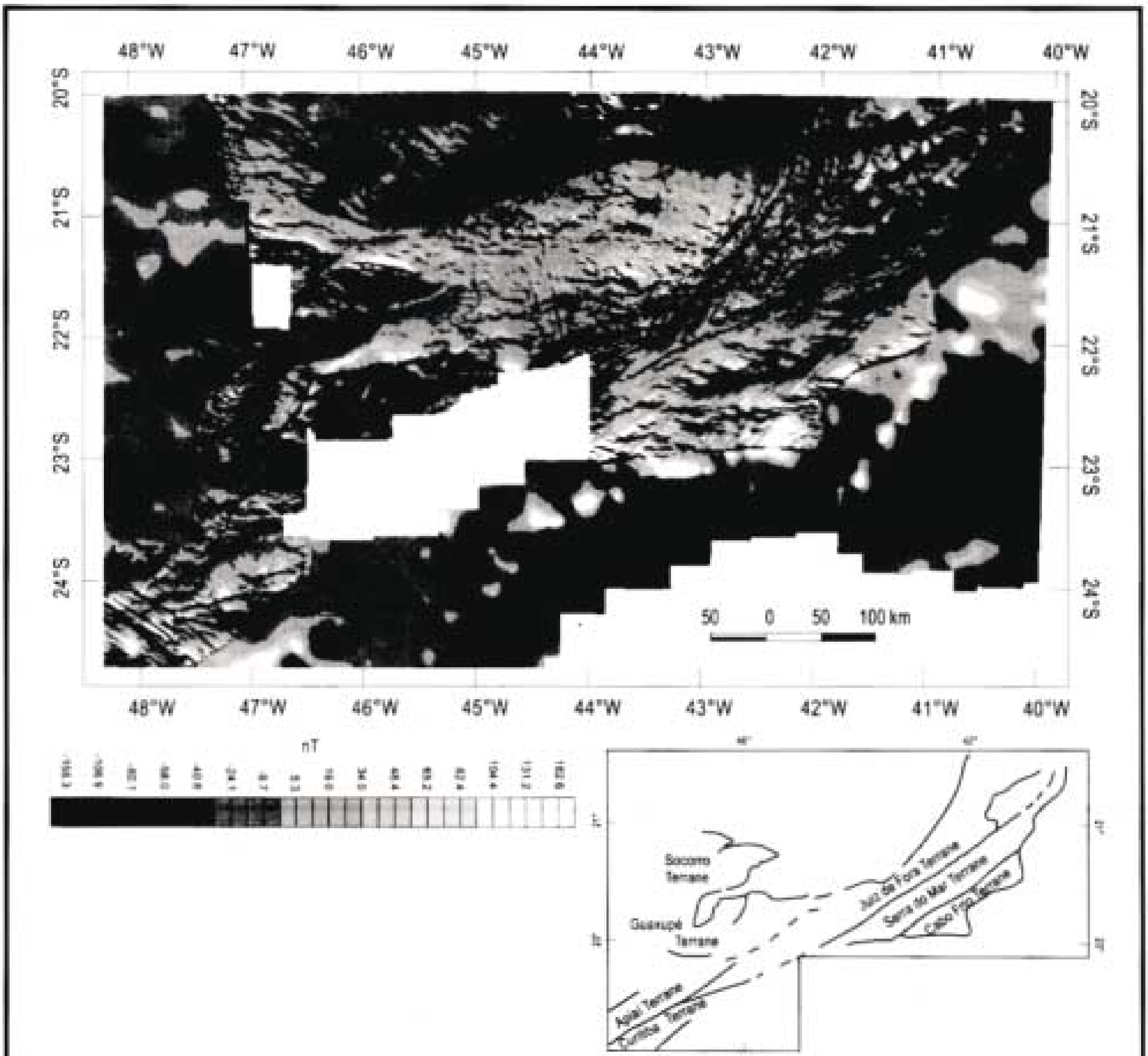


Fig. 20 - Aeromagnetic map from SE-Brazil. The NE-trending anomalies correspond to major shear-zones of the Ribeira belt. (Data processed by DIGEOF - CPRM). Total field with shaded relief, synthetic illumination with 45° of inclination and declination.

Cambrian boundary (*c.* 550 - 540 Ma) by zircon and monazite U/Pb geochronology (Delhal *et al.*, 1969; Söllner *et al.*, 1989, 1991). On the African side, the widespread foreland deposits of the Nama Group and the unconformable uppermost cover of the Fish River Group were the cratonic record of a collision tectonics at the same time (Frimmel *et al.*, 1998).

Nevertheless, the final closure of the Adamastor Ocean was later in the Cambrian times. The high-pressure metamorphic conditions indicated by the kyanite-bearing assemblages in high-temperature metasediments from Cabo Frio Terrane are incompatible with the low-pressure, high-temperature rocks at its neighbouring Rio Negro and Serra dos Órgãos superposed batholiths. These different metamorphic belts are related to an eastern displacement and apparently ascribe a northwestwards continental subduction of the western promontory of the Congo Plate under the Serra do Mar Terrane. This collision-related metamorphism is recorded by monazite and zircon U/Pb geochronology at 520 Ma (Schmitt *et al.*, 1999).

Pb/Pb isotope studies carried out on undeformed carbonate rocks from the Bambuí Group in the cratonic area are evidence for an incorporation of Pb-fluid phase promoting isotopic resetting and severe remagnetization at about 530 - 500 Ma (Babinsky *et al.*, 1999; Agrella Filho *et al.*, 1999). These data ascribe the continent-scale record for the last Mantiqueira collision, although the Atlantic drift must have hid most of the suture lines.

Thus, the estimated life span of the Adamastor Ocean is at least 230 Ma, from the Middle Cryogenian to the Middle Cambrian (750 - 520 Ma). Their complex plate and microplate interactions makes it compatible with the Cordilleran-type evolution.

Late orogenic basins

A series of pull-apart basins occur at the northwestern suture boundary of the Mantiqueira Orogenic System, reaching the Tocantins orogens up to the northern edge of the extrusion-related strike-slip shear zones. The development of these basins shows evidence for several pulses of shear zones displacement. After the high-temperature dextral motion (*c.* 575 Ma, Machado *et al.*, 1996) of the shear zones, and the exhumation of the metamorphic belt, a main sinistral displacement took place developing the pull-apart basins at their releasing bends. The sedimentation was storm-dominated with alluvial fans reaching the asymmetrical basin border and grading to alluvial plain deposits up to distal lacustrine environment though seaward connection (Teixeira, 1995). They were weakly deformed by dextral transpression and metamorphosed under very low grades during the final Mid-Cambrian collision.

Post-orogenic transition from compressive to extensional collapse

Mixing-mingling of mantle and crustal derived magmas strongly characterized the post-orogenic plutonic rocks accreted mainly in the Serra do Mar Terrane. They are coeval to circular zoned plutons of monzogabbro-norite and diorite partially wrapped by granite that characteristically display a wide and discontinuous central co-mingling and mixing zone. More mature magmatic phases comprise diorite-tonalite and granodiorite surrounded by granite (Wiedemann *et al.*, 1997). Allanite granite from the zoned Angelica Pluton displays a zircon age at the Upper Cambrian (*c.* 513 Ma, Söllner *et al.*, 1987). This magmatic event could be connected with the maintenance of the high thermal gradient inducing metamorphism up to Lower Ordovician times (Siga Jr., 1986, Wiedemann *et al.*, 1997; Schmitt *et al.*, 1999; Machado *et al.*, 1996) that followed the final *c.* 520 Ma collision episode.

General framework: the Ribeira Belt

The Ribeira Belt describes the main geometric features produced by the complex and diachronic kinematic plate interaction in the northern part of the Mantiqueira Orogenic System (Fig. 20). The São Paulo Shear Belt, up to 100 km wide, extends for 1000 km, delineating the boundaries between the terranes and normally overprinting low-angle metamorphic slices (Campos Neto and Figueiredo, 1995; Heilbron *et al.*, 1995). The linear belts of strong non-coaxial strain resulted from several pulses of mostly dextral strike-slip motion. They are a crust-scale shear-strain ascribed by an early high-temperature amphibolite to granulite facies metamorphism that decreases up to low greenschist facies toward the SW. Most of the granite bodies (Machado and Demange, 1998) display a shear zone-controlled emplacement. They are equigranular to porphyritic biotite-bearing metaluminous granite intrusives, and deeper garnet bearing coarse-grained to porphyritic granites associated with partial melting of the country rocks. They grade to shallow muscovite-bearing granite farther to the S, into the Embú Complex. Likewise the diachronous southeastward step-up of the collisional episodes, the granite bodies related to the shear zones are older (590 - 565 Ma) in the Juiz de Fora Terrane than those emplaced in the Serra do Mar Terrane (535 - 520 Ma). There the high thermal gradient prevailed up to Lower Ordovician at 503 - 492 Ma (Ebert *et al.*, 1996; Machado *et al.*, 1996; Schmitt *et al.*, 1999).

Toward the S, the strike-slip shear zones display relict sinistral displacement. They are related to amphibolite facies metamorphism in the northwestern part of the Apiaí Terrane (Garcia and Campos Neto, 1997) contrasting with the low-grade mylonite from the Lancinha Shear Zone at the southeastern edge of this terrane. Both shear zones were reprinted by later dextral displacement. It results from dramatic episodes of lateral extrusion controlled by the kinematics of plate interaction related to the closure of the Adamastor Ocean acting against an oblique plate and microplate margins. The main W-oriented compression related to the convergence of the northern part of the Adamastor Ocean Plate explains the dextral vector of the shear zones. On the other hand the NNE- oriented compression related to the precedent southern closure of the oceanic branches results in a sinistral shear zone motion. The *c.* 520 Ma westward collision controlled the locally intense dextral strike-slip overprint.

From Rodinia to Gondwana: SE-Brazil geodynamic evolution

Extensional regimes immediately followed the Grenville orogens in the Amazonas Plate. They were also recorded in the São Francisco Plate, and related to the Tonian-age taphrogeny (1.1 Ga - 900 Ma), leading to the former break-up of Rodinia. An elongate rift valley trending NW dominates the western border of the São Francisco Plate. A wide lower plate-type passive continental margin is the record of the rift-drift evolution. These basin-forming tectonic processes were related to the Tonian taphrogeny and the opening of the Goianides Ocean. The spreading life of this ocean was early and locally replaced by intra-oceanic plate convergence triggering the Tocantins Orogenic System. A set of W-dipping subduction plates occurred diachronously generating island arcs, evolved magmatic arcs, and back-arc basins. The Rio de la Plata Plate and Central Goiás Terrane were driven against the western margin of the São Francisco Plate.

The early orogenies are recorded by the Mara Rosa Island Arc accretion (900 - 850 Ma) further docking over the western boundary of the Central Goiás Terrane (800 - 770 Ma), promoting local metamorphism and relief. Farther to the S, a back-arc spreading basin could have been nearly coeval with the Mara Rosa docking, separating two other magmatic arc built at the Middle Cryogenian: the eastern part of the Anápolis–Andrelândia Terrane, and the western active continental margin related to the Iporá-Jaupací/Socorro-Guaxupé terranes. The supracrustal units of these magmatic arcs were preserved at the opposite termination of both terranes. The southernmost comprises the collisional-related low-angle slices of high-pressure subducted continental slab of the Anápolis-Andrelândia Terrane, suggesting an oblique WSW-driven plate subduction.

At this time, the Upper Cryogenian diamictite beds correlated with the Sturtian-age glacial deposits preceded the carbonate cap toward the São Francisco Plate interior and conformably overlying the western passive continental margin. The ENE displacement of the southernmost high-pressure nappes seems to be controlled by the back up extrusion at a critical state of the W-dipping lithosphere subduction from the western edge of the São Francisco Plate. The magmatic, metamorphic and depositional records of this main collision-controlled orogeny, juxtaposing the Rio de la Plata and the São Francisco plates and closing the Goianides Ocean, can be found everywhere in the Tocantins orogens, constrained at 630 - 620 Ma. To the S of this orogenic system, the outward propagation of the high-pressure nappe system (comprising Ky-granulites and eclogites) reaching a displacement of 200 km, was subjected to strong thinning, and temperature decrease at each thrusted slice.

The overriding of the high-temperature root of the Socorro-Guaxupé Nappe controlled an inverted metamorphism related to a near-isobar heating throughout the upper high-pressure nappe. The external passive continental margin sequences following roughly the kinematics of this nappe displayed a medium-pressure metamorphic pattern. They attained the platform setting as a stretched, flat-lying, spoon-like, metamorphic nappe. The post-orogenic relaxation, at the inner nappe, took place at 612 Ma. Thus, a long life span (Fig. 21) of oceanic plate convergence (*c.* 270 Ma) and the rapid overriding-type collision episode (*c.* 18 Ma) required to preserve the high-pressure metamorphic slices are the main characteristics of the Tocantins Orogenic System, having as modern analogous the India-Asia collision.

At the eastern margin of the São Francisco Plate the Sturtian-age glacial deposits triggered the rift stage that evolved eastwards to a deep marine basin connected with the Adamastor oceanic crust. This extensional tectonic regime was coeval with the African counterpart rifting. Thus, the Mid-Cryogenian global break-up of Rodinia coexisted with the Tocantins orogenies. The diachronism of the succeeding orogenic processes was a principal feature of the Mantiqueira Orogenic System. It was wholly related to complex interactions between several small continental plate fragments mostly turned to microplates closing the branches of the Adamastor Ocean.

The first record of plate convergence occurred early and locally (?) S of the Brazilian counterpart of this oceanic context, as a juvenile island arc accretion at 700 Ma (São Gabriel intra-oceanic subduction-controlled orogeny). Therefore the plate interaction kinematics with an apparent northwestly motion caused the closure of the southern part of the Adamastor Ocean. The southern Mantiqueira orogens, mainly associated with evolved magmatic arcs and high-temperature metamorphic conditions record Neoproterozoic III subduction-controlled orogenies decreasing in age toward the eastern Rio de la Plata boundary, up to the 610 Ma back-arc basin. The main collisional period that closed the southern oceanic branches between terranes and the just-amalgamated Rio de la Plata-São Francisco Plate took place at 600 Ma. Sinistral shearing, oriented NE-SW accommodated the crustal shortening. A wide peripheral foreland basin, containing Vendian palinomorphs, reached deep sea conditions and was contemporaneous with felsite volcanism up to 560 Ma. Small succeeding volcano-sedimentary basins associated with extensive alkaline-peralkaline plutonism occurs essentially within this time span, up to younger (540 Ma) elsewhere farther into the plate interior.

The northern Mantiqueira orogens started with the plate convergence and docking between Juiz de Fora and Serra do Mar terranes (Rio Negro Orogen 630 - 610 Ma) advancing the E-dipping subduction of the eastern margin of the São Francisco Plate. The Araçuaí Orogen that started at 600 Ma was believed to be responsible for the inversion of this passive continental margin and the eastward continental growth related to the subduction-controlled Galiléia Magmatic Arc (NW-facing the Juiz de Fora Terrane) and their collision against the eastern São Francisco passive continental margin (580 Ma). The Serra do Mar Terrane was overprinted by a younger magmatic arc (580 - 565 Ma) further overriding the northeastern part of the Araçuaí Orogen (*c.* 540 Ma) defining, as a whole, the Rio Doce Orogen in a continued continental growing. NE-oriented dextral shearing mostly accommodated the strong E-W shortening and crustal thickening.

This orogeny seems to have been widespread in the Mantiqueira system on account of shortening that controlled the inversion of the NE-trending foreland basins. On the African side the major episode of

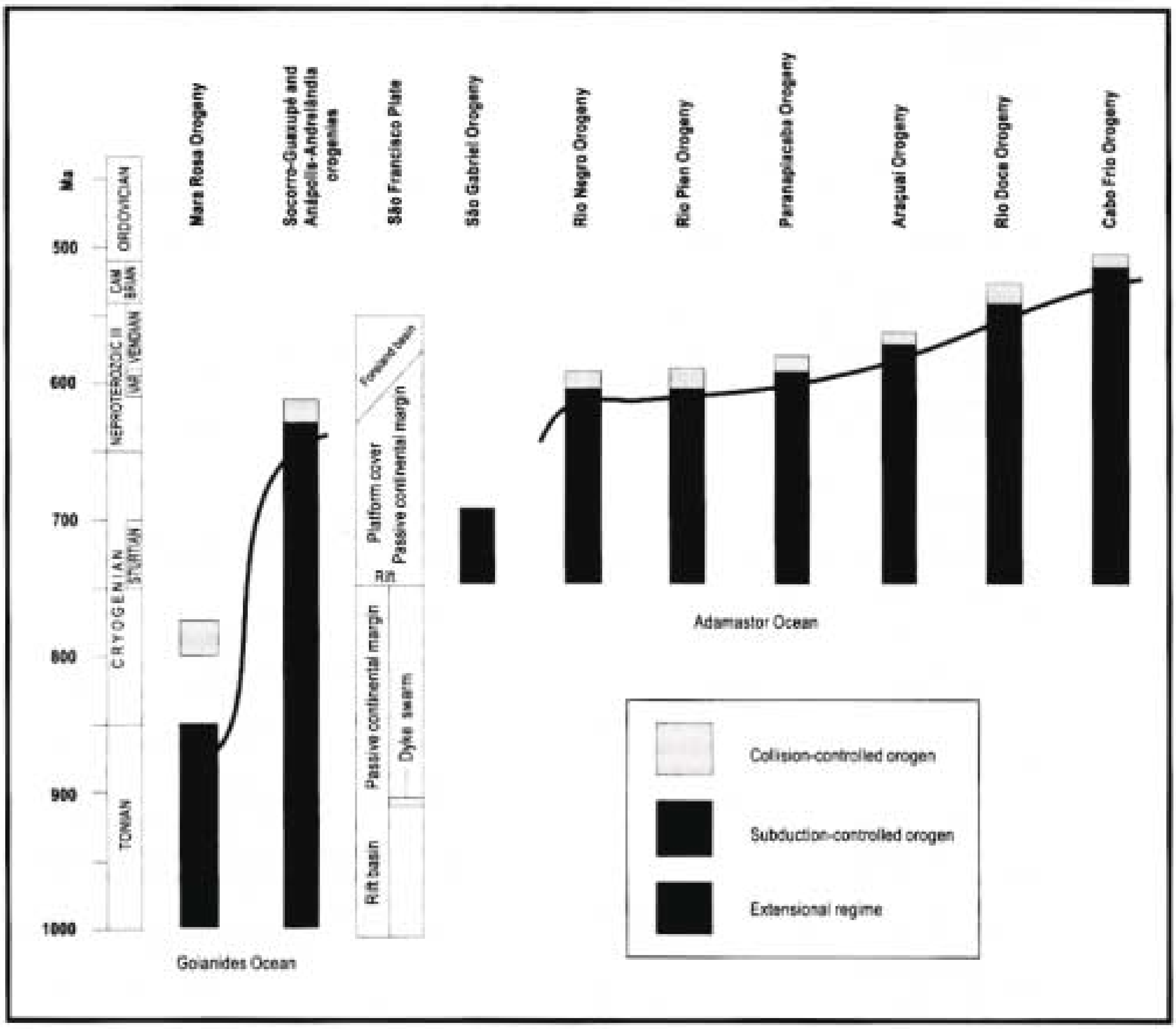


Fig. 21 - Tectonic correlation table.

continent-continent collision and oceanic crust obduction was essentially contemporaneous. Nevertheless, the final consumption of the oceanic lithosphere was related to the collision of the western promontory of the Congo Plate (Cabo Frio Terrane) against the just-amalgamated eastern Brazil, during the Middle Cambrian (520 Ma). The post-kinematic magmatism proceeded up to the Ordovician boundary. Excluding the early São Gabriel juvenile orogenic accretion the main convergence life span of the Adamastor Ocean lasted for 100 Ma (Fig. 21). The continental growth by along-lived succession of orogens, related to arc-derived crustal accretion and docking against an older plate, enhances the analogies between Mantiqueira orogens and the Cordilleran-style tectonics (Clowes *et al.*, 1999).

REFERENCES

Agrella Filho, M.S., Trindade, R.I.F., Siqueira, R., Ponte Neto, C.F. and Pacca, I.I.G. (1998). Paleomagnetic constraints on the Rodinia Supercontinent: Implications for its Late Proterozoic break-up and the fomation of Gondwana. *International Geology Review*, **40**, 171-188.

Agrella Filho, M.S., Babinski, M., Trindade, R.I.F., VanSchmus, W.R. and Ernesto, M. (1999). Simultaneous remagnetization and U-Pb isotope resetting in Late Proterozoic carbonates of the São Francisco craton, Brazil. *Precambrian Research* (in press).

Alkmim, F.F., Brito Neves, B.B. and Alves, J.A.C. (1993). Arcabouço tectônico do Cráton do São Francisco – uma revisão. In: *O Cráton do São Francisco*, eds. J.M.L. Dominguez and A. Misi, pp. 45-62. SBG-Bahia/Sergipe, SGM, CNPq.

Alkmim, F.F. and Marshak, S. (1998). Transamazonian Orogeny in the Southern São Francisco Craton Region, Minas Gerais, Brazil: evidence for Paleoproterozoic collision and collapse in the Quadrilátero Ferrífero. *Precambrian Research*, **90**, 29-58.

Almeida, F.F.M., Amaral, G., Cordani, U.G. and Kawashita, K. (1973). The Precambrian evolution of South American cratonic margin South of Amazonas River. In: *The Oceans Basins and Margins*, I, eds. Nairn and Stelli, pp. 411-446. Plenum.

Almeida, F.F.M. (1977). O cráton do São Francisco. *Revista Brasileira de Geociências*, **7**, 349-364.

Almeida, F.F.M., Hasui, Y., Brito Neves, B.B. and Fuck, R.A. (1981). Brazilian structural provinces: An introduction. *Earth Science Reviews*, 17, 1-29.

Arndt, N.T., Teixeira, N.A. and White, W.M. (1989). Bizarre geochemistry of komatiites from the Crixás greenstone belt, Brazil. *Contribution to Mineralogy and Petrology*, **101**, 187-197.

Babinski, M., Chemale Jr, F., Hartmann, L.A., VanSchmus, W.R. and Silva, L.C. (1996). Juvenile accretion at 750-700 Ma in southern Brazil. *Geology*, **24**, 439-442.

Babinski, M., VanSchmus, W.R. and Chemale Jr., F. (1999). Pb-Pb dating and Pb isotope geochemistry of Late Proterozoic carbonate rocks from the São Francisco basin, Brazil: implications for the mobility of Pb isotopes during tectonism and metamorphism. *Chemical Geology*, **160**, 175-199.

Bahlburg, H. and Hervé, F. (1997). Geodynamic evolution and tectonostratigraphic terranes of the northwestern Argentina and northern Chile. *Geological Society of America Bulletin*, **109**, 869-884.

Barbosa, O., Braun, O.P.G., Dyer, R.C. and Cunha, C.A.B. (1970). Geologia da região do Triângulo Mineiro. *DNPM-DFPM, Bol.*, **136**, 1-140.

Barbosa, P.A. and Jost, H. (1990). Caracterização das seqüências supracrustais precambrianas a leste de Mossâmedes - GO. In: *36th Congresso Brasileiro de Geologia*, **1**, 1-16.

Basei, M.A.S. (1985). *O cinturão Dom Feliciano em Santa Catarina.* Doctor thesis, Universidade de São Paulo, São Paulo, Brazil, 190 p. (unpublished).

Basei, M.A.S., Siga Jr., O., Machiavelli, A., Mancini, F., Avelar, A.S. and Coelho Neto, A.L. (1992). Evolução tectônica dos terrenos entre os cinturões Ribeira e Dom Feliciano. *Revista Brasileira de Geociências*, **22**, 216-221.

Basei, M.A.S., Siga Jr., O., Sato, K. and Sproesser, W.M. (1995). A metodologia Urânio-Chumbo na USP, princípios metodológicos, aplicações e resultados obtidos. *Anais da Academia brasileira de Ciências*, **67**, 221-237.

Basei, M.A.S., Citroni, S.B.V. and Siga Jr., O. (1997). Stratigraphy and age of Fini-Proterozoic Basins of Paraná and Santa Catarina states, southern Brazil. In: Asmus, A.J., Rocha Campos, A.R. and Gamundi, O.L. (eds.). *Sedimentary Basins of South America* (in press).

Basei, M.A.S., Siga Jr., O., Reis Neto, J.M., Harara, O.M., Passarelli, C.R. and Machiavelli, A. (1997). Geochronological map of the Precambrian terrains of Paraná and Santa Catarina States, southern Brazil: Tectonic implications. *South-American Symposium on Isotope Geology, Extended Abstracts*, 44-46.

Basei, M.A.S., McReath, I. and Siga Jr., O. (1998). The Santa Catarina granulite complex of southern Brazil: a review. *Gondwana Research*, **1**, 383-392.

Bettencourt, J.S., Onstott, T.C., Jesus, T. and Teixeira, W. (1996). Tectonic interpretation of 40Ar/39Ar ages on country rocks from the central sector of the Rio Negro-Juruena Province, Southwest Amazonian craton. *International Geology Review*, **38**, 42-56.

Bettencourt, J.S., Tosdal, R.M., Leite, W.B. and Payolla Jr., B.L. (1999). Mesoproterozoic rapakivi granites of the Rondônia Tin Province, southwestern border of the Amazonian craton, Brazil -I. Reconnaissance U-Pb geochronology and regional implications. *Precambrian Research*, **95**, 41-67.

Bergmann, M. (1988). *Caracterização estratigráfica e estrutural da sequência vulcano-sedimentar do Grupo São Roque na região de Pirapora do Bom Jesus - Estado de São Paulo.* Master dissertation, IG-USP, São Paulo, Brazil, 164 p. (unpublished).

Brito Neves, B.B., Kawashita, K., Cordani, U.G. and Delhal, J. (1979). A evolução geocronológica da cordilheira do Espinhaço, dados novos e integração. *Revista Brasileira de Geociências*, **9**, 71-85.

Brito Neves, B.B. and Cordani, U.G. (1991). Tectonic evolution of South America during the Late Proterozoic. *Precambrian Research*, **53**, 23-40.

Brito Neves, B.B., Campos Neto, M.C. and Fuck, R.A. (1999). From Rodinia to Western Gondwana: An approach to the Brasiliano-Pan African Cycle and orogenic collage. *Episodes*, **22**, 155-166.

Campanha, G.A.C. and Sadowski, G.R. (1999). Tectonics of the southern portion of the Ribeira Belt (Apiaí Domain). *Precambrian Research* (in press).

Campos Neto, M.C. (1983). Contribuição à litoestratigrafia e estrutura do Grupo Açungui no sudeste do Estado de São Paulo. In: *4th Simpósio Regional de Geologia, São Paulo, Atas*, 103-112.

Campos Neto, M.C. and Basei, M.A.S. (1983). Evolução estrutural brasiliana do nordeste de São Paulo: dobramentos superpostos e esboço estratigráfico e tectônico. In: *4th Simpósio Regional de Geologia, São Paulo, Atas*, 61-78.

Campos Neto, M.C. (1984a). Litoestratigrafia, relações estratigráficas e evolução paleogeográfica dos Grupos Canastra e Paranoá (região de Vazante-Lagamar, MG). *Revista Brasileira de Geociências*, 14, 81-91.

Campos Neto, M. C. (1984b). Geometria e fases de dobramentos superpostos no oeste de Minas Gerais. *Revista Brasileira de Geociências*, **14**, 60-68.

Campos Neto, M.C., Figueiredo, M.C.H., Janasi, V.A., Basei, M.A.S. and Fryer, B.J. (1988). The São José do Rio Pardo mangeritic-Suite, southeastern Brazil. *Geochimica Brasiliensis*, **2**, 185-199.

Campos Neto, M.C., Perrotta, M.M., Peloggia, A.U. and Figueiredo, M.C.H. (1990). A porção ocidental da Faixa Alto Rio Grande (SP-MG). In: *36th Congresso Brasileiro de Geologia*, **6**, 2615-2630.

Campos Neto, M.C. and Figueiredo, M.C.H. (1990). Evolução geológica dos terrenos Costeiro, Paraíba do Sul e Juiz de Fora (RJ-MG-ES). In: *36th Congresso Brasileiro de Geologia*, **6**, 2631-2648.

Campos Neto, M.C. and Figueiredo, M.C.H. (1995). The Rio Doce Orogeny, southeastern Brazil. *Journal of South American Earth Sciences*, **8**, 143-162.

Campos Neto, M.C. and Caby, R. (1999a). Neoproteozoic high-pressure metamorphism and tectonic constraint from the nappe system south of the São Francisco Craton, southeast Brazil. *Precambrain Research*, **97**, 3-26.

Campos Neto, M.C. and Caby, R. (1999b). Terrane accretion and upward extrusion of high-pressure granulites in the Late Proterozoic nappes of Southeast Brazil: Petrological and structural constraints. *Tectonics* (in press).

Chemenda, A.I., Mattauer, M., Malavieille, J. and Bokun, A.N. (1995). A mechanism for syn-collisonal rock exhumation and associated normal faulting, results from physical modelling. *Earth and Planetary Science Letters*, 132, 225-232.

Cloud, P. and Dardenne, M.A. (1973). Proterozoic age of the Bambuí Group in Brazil. *Geological Society of America Bulletin*, **84**, 1673-1676.

Clowes, R., Cook, F., Hajnal, Z., Hall, J., Lewry, J., Lucas, S. and Wardle, R. (1999). Canada's LITHOPROBE Project (Collaborative, multidisciplinary geoscience research leads to new understanding of continental evolution). *Episodes*, **22**, 3-20.

Cordani, U.G., Brito Neves, B.B., Fuck, R.A., Porto, R., Thomaz Filho, A. and Cunha, F.M.B. (1984). Estudo preliminar de integração do Pré-Cambriano com os eventos tectônicos das bacias sedimentares brasileiras. *Ciência Técnica Petróleo*, **15**, 10-70.

Correia, C.T., Girardi, V.A.V., Lambert, D.D., Kinny, P.D. and Reeves, S.J. (1996). 2 Ga U-Pb (SHRIMP II) and Re-Os ages for the Niquelândia basic-ultrabasic layered intrusion, central Goiás, Brazil. In: *39th Congresso Brasileiro de Geologia, SBG, Extend Abstracts*, 187-189.

Correia, C.T., Jost, H., Tassinari, C.C.G., Girardi, V.A.V. and Kinny, P.D. (1999). Ectasian Mesoproteozoic U-Pb ages (SHRIMP II) for the metavolcano-sedimentary sequences of Juscelândia and Indaianópolis and for high grade metamorphosed rocks of Barro Alto stratiform igneous complex, Goiás State, Central Brazil. In:

2nd South American Symposium on Isotope Geology, Extend Abstracts, 31-33.

Daitx, E.C. (1996). *Origem e evolução dos depósitos sulfetados tipo-Perau (Pb-Zn-Ag), com base nas jazidas Canoas e Perau (Vale do Ribeira, Pr)*. Doctor thesis, UNESP, São Paulo, Brazil, 453 p. (unpublished).

Dalziel, I.W.D. (1997). Late Proterozoic-Paleozoic geography and tectonics: Review, hypothesis, environmental speculation. *Geological Society of America Bulletin*, **109**, 16-42.

Danni, J.C.M., Fuck, R.A. and Leonardos, O.H. (1982). Archaean and lower Proterozoic units in Central Brazil. *Geologische Rundschau*, **71**, 291-317.

Dardenne, M.A. (1981). Os Grupos Paranoá e Bambuí na Faixa Brasília. In: *Simpósio sobre o Cráton do São Francisco, Anais*, 140-157.

Delhal, J., Ledent, D. and Cordani, U.G. (1969). Ages Pb/U, Sr/Rb et Ar/K des formations métamorphiques et granitique du sudeste du Brésil (États de Rio de Janeiro et de Minas Gerais). *Annales de la Société Gélogique Belgique*, **92**, 271-283.

Del Lama, E.A., Oliveira, M.A.F. and Zanardo, A. (1994). Geotermobarometria em rochas do Complexo Campos Gerais ao norte da zona de cisalhamento Varginha. *Revista Brasileira de Geociências*, **24**, 233-239.

Dominguez, J.M.L. (1993). As coberturas do cráton do São Francisco: uma abordagem do ponto de vista da análise de bacias. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and A. Misi, pp. 137-159. SBG, SGM, CNPq,.

Dossin, T.M., Dossin, I.A., Charvet, J., Pouclet, A. and Lapierre, H. (1993). Late Proterozoic mafic dykes swarm from the Espinhaço Range (Minas Gerais, Brazil): geochemistry and tectonic setting. In: *2nd Simpósio sobre o cráton do São Francisco, Anais*, 128-130.

Duarte, B.P., Figueiredo, M.C.H., Campos Neto, M.C. and Heilbron, M. (1997). Geochemistry of granulite facies orthogneisses of the Juiz de Fora Complex, central segment of Ribeira Belt, SE Brazil. *Revista Brasileira de Geociências*, **27**, 67-82.

Ebert, H., Chemale, F., Babinski, M., Artur, A.C. and Van Schmus, W.R. (1996). Tectonic setting and U-Pb zircon dating of the plutonic Socorro complex in the transpressive Rio Paraiba do Sul Shear Belt, SE Brazil. *Tectonics*, **15**, 688-699.

Ebert, H.D. and Hasui, Y. (1998). Transpressional tectonics and strain partitioning during oblique collision between three plates in the Precambrian of southeast Brazil. In: Continental Trasnpressional and Transtensional Tectonics, eds. Holdsworth, R.E., Strachan, R.E. and Dewey, J.F. *Geological Society, London, Special publications*, **135**, 231-252.

Fairchild, T.R. (1977). *Conophyton* and other columnar stromatolites from the Upper Precambrian Açungui Group near Itapeva, SP, Brazil. In: *1st Simpósio de Geologia Regional-SP, Atas*, 179-198.

Faria, A. (1995). *Estratigrafia e sistemas deposicionais do Grupo Paranoá nas áreas de Cristalina, Distrito Federal e São João da Aliança-Alto Paraíso de Goiás*. Doctor thesis, UnB, Brasília, Brazil, 199 p. (unpublished).

Fernandes, A.J., Campos Neto, M.C. and Figueiredo, M.C.H. (1990). O Complexo Embú no leste do Estado de São Paulo: limites e evolução geológica. In: *36th Congresso Brasileiro de Geologia*, **6**, 2755-2763.

Fernandes, L.A.D., Menegat, R., Costa, A.F.U., Koester, E., Porcher, C.C., Tommasi, A., Kraemer, G., Ramgrab, G.E. and Camozzato, E. (1995). Evolução tectônica do cinturão Dom Feliciano no escudo Sul-Riograndense: Parte I – uma contribuição a partir do registro geológico. *Revista Brasileira de Geociências*, **25**, 351-374.

Ferreira-Filho, C.F., Nilson, A. and Naldrett, A. (1992). The Niquelândia mafic-ultramafic complex, Goiás, Brazil: a contribution to the ophiolite X stratiform controversy based on new geological and structural data. *Precambrian Research*, **59**, 125-142.

Ferreira-Filho, C.F., Kamo, S.L., Fuck, R.A., Krogh, T.E. and Naldrett, A.J. (1994). Zircon and rutile U-PB geochronology of the Niquelândia layered mafic and ultramafic intrusion, Brazil: constraints for the timing of magmatism and high grade metamorphism. *Precambrian Research*, **68**, 241-255.

Ferreira-Filho, C.F., Moraes, R., Fawcett, J.J. and Naldrett, A.J. (1998). Amphibolite to granulite progressive metamorphism in the Niquelândia Complex, central Brazil: regional tectonic implications. *Journal of South American Earth Science*, **11**, 35-50.

Figueiredo, M.C.H. and Campos Neto, M.C. (1993). Geochemistry of the Rio Doce Magmatic Arc, Southeastern Brazil. *Anais da Academia Brasileira de Ciências*, **63**, 63-82.

Figueiredo, M.C.H. and Teixeira, W. (1996). The Mantiqueira metamophic complex, eastern Minas Gerais State: preliminary geochronological and geochemical results. *Anais da Academia Brasileira de Ciências*, **68**, 223-246.

Fischel, D.P., Pimentel, M.M., Fuck, R.A., Costa, A.G. and Rosière, C.A. (1998). Geology and Sm-Nd isotopic data for the Mantiqueira and Juiz de Fora Complexes (Ribeira Belt) in the Abre Campo-Manhuaçu Region, Minas Gerais, Brazil. In: *14th International Conference on Basement Tectonics, Extend Abstracts*, 21-23.

Fischel, D.P., Pimentel, M.M. and Fuck, R.A. (1999). Preliminary Sm-Nd isotopic study of the Anápolis-Itaçú Complex, Araxá Group and associated granite intrusions, Central Brazil: implications for the evolution of the Brasília Belt. In: *2nd South American Symposium on Isotope Geology, Extend Abstracts*, 302-305.

Fonseca, M.A. (1996). *Estilos estruturais e arcabouço tectônico do segmento setentrional da Faixa Brasília*. Thesis, UnB, Brasília, Brazil, 172 p. (unpublished).

Fragoso-César, A.R.S. (1980). O Cráton Rio de la Plata e o Cinturão Dom Feliciano no Escudo Uruguaio Sul-riograndense. In: *31th Congresso Brasileiro de Geologia*, **5**, 2879-2892.

Fragoso-César, A.R.S. (1991). *Tectônica de placas no ciclo Brasiliano: as orogenias dos cinturões Dom Feliciano e Ribeira no Rio Grande do Sul*. Doctor thesis, IG-USP, São Paulo, Brazil, 367 p. (unpublished).

Frascá, M.H.B.O., Figueiredo, M.C.H., Almeida, M.A. and Coutinho, J.M.V. (1990). Petrografia e geoquímica da Formação Água Clara - região de Araçaíba, SP. *Boletim do Instituto de Geociências*, USP, **21**, 73-92.

Frascá, M.H.B.O., Campanha, G.A.C., Figueiredo, M.C.H. and Sadowski, G.R. (1997). Geoquímica e ambiência tectônica de metabasitos do Alto e Médio Vale do Ribeira, São Paulo e Paraná. *Revista Brasileira de Geociências*, **27**, 41-48.

Frimmel, H.E., Almond, J. and Gesse, P.G. (1998). Gariep belt and Nama basin. In: *Excursion Guidebook, Gondwana-10, Department of Geological Sciences, University of Cape Town*, 1-75.

Fritzer, T. (1991). Das Guaçuí-Linament und die orogene Entwicklung des zentralen Ribeira-Belts. *Münchner Geologie Hefte*, **2**, 1-195.

Fuck, R.A., Brito Neves, B.B., Cordani, U.G. and Kawashita, K. (1989). Geocronologia Rb-Sr no Complexo Barro Alto, Goiás: evidência de metamorfismo de alto grau e colisão continental há 1300 Ma no Brasil central. *Geochimica Brasilienses*, **3**, 125-140.

Fuck, R.A., Sá, E.F.J., Pimentel, M.M., Dardenne, M.A. and Soares, A.C.P. (1993). As faixas de dobramentos marginais do cráton do São Francisco: síntese dos conhecimentos. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and Misi, A., pp. 161-185. SBG-Ba/Se, SGM, CNPq.

Fugi, M.Y. (1989). *REE geochemistry and Sm-Nd geochronology of the Cana Brava Complex, Brazil*. Master Thesis, Kobe University, Japan, 55 p.

Garcia, M.G.M. and Campos Neto, M.C. (1997). Superposição destral em ampla zona de cisalhamento sinistral: cinturão de cisalhamento São Paulo, imediações de Piracaia-SP. *Revista Brasileira de Geociências*, **27**, 339-348.

Gaucher, C., Sprechmann, P. and Montaña, J. (1998). New advances in the geology and paleontology of the Vendian to Cambrian Arroyo del Soldado Group of the Nico Pérez terrane of Uruguay. *Neues*

Jahrbuch für Geologie und Paläontologie Monatshefte, 3 (in press).

Gimenez Filho, A., Teixeira, W., Figueiredo, M.C.H. and Trevizoli Jr., L. (1995). Geologia, petrografia e litogeoquímica do complexo granítico Três Córregos na região de Barra do Chapéu e Ribeirão Branco, SP. *Revista Brasileira de Geociências*, 25, 92-106.

Girardi, V.A., Rivalente, G. and Sinigoi, S. (1986). The petrogenesis of the Niquelândia layered basic-ultrabasic complex, central Goiás, Brazil. *Journal of Petrology*, 27, 715-744.

Hackspacher, P., Dantas, E.L., Godoy, A.M., Oliveira, M.A.F., Fetter, A. and VanSchmus, W.R. (1999). Considerations about the evolution of the Ribeira Belt in the São Paulo state, Brazil, from U-Pb geochronology in metavolcanic rocks of the São Roque Group. *2nd South American Symposium on Isotope Geology, Extend Abstracts*, 310-313.

Harara, O.M., Basei, M.A.S. and Siga Jr., O. (1997). Geochronoloical and geochemical data on the trasition zone betwee Luis Alves and Atuba Complexes, south Brazil. In: *South-American Symposium on Isotope Geology, Extended Abstracts*, 134-136.

Hartnady, C.J.H., Joubert, P. and Stowe, C.W. (1985). Proterozoic crustal evolution in soutwestern Africa. *Episodes*, 8, 236-244.

Hasui, Y. (1982). The Mantiqueira Province: Archaean strucutre and Proterozoic evolution. *Revista Brasileira de Geociências*, 12: 167-172.

Heilbron, M., Chrispim, S.J., Alves, N.P. and Simões, L.S.A. (1982). Geologia do Cabo dos Búzios (Estado do Rio de Janeiro). *Anais da Academia Brasileira de Ciências*, 54, 553-562.

Heilbron, M., Valeriano, C., Valladares, C.S. and Machado, N. (1995). A orogênese brasiliana no segmento central da Faixa Ribeira, Brasil. *Revista Brasileira de Geociências*, 25, 249-265.

Heilbron, M., Tupinambá, M., Almeida, J.C.H., Valeriano, C.M., Valladares, C.S. and Duarte, B.P. (1998). New constraints on the tectonic organisation and structural styles related to the Brasiliano collage of the Central segment of the Ribeira Belt, SE Brazil. In: *14th International Conference on Basement Tectonics, Extend Abstracts*, 15-17.

Hodges, K.V., Parrish, R.R. and Searle, M.P. (1996). Tectonic evolution of the central Annapurna Range, Nepalese Himalayas. *Tectonics*, 15, 1264-1291.

Hoffman, P.F., Kaufman, A.J., Halverson, G.P. and Schrag, D.P. (1998). A Late Proterozoic Snowball Earth. *Science*, 281, 1342-1346.

James, D.E. and Assumpção, M. (1996). Tectonic implications of S-wave anisotropy beneath SE Brazil. *Geophysical Journal International*, 126, 1-10.

Janasi, V.A. (1997). Late Proterozoic mangerite-granite magmatism in the southeastern Brazil: the São Pedro de Caldas massif. *Anais da Academia Brasileira de Ciências*, 69, 267-394.

Janasi, V.A. (1999). *Petrogênese de granitos crustais na Nappe de Empurrão Socorro-Guaxupé (SP-MG): uma contribuição da geoquímica elemental e isotópica*. "Livre Docência" thesis, IG-USP, São Paulo, Brazil, 304 p. (unpublished).

Janasi, V.A. and Ulbrich. H.G.J. (1991). Late Proterozoic grnitoid magmatism in the state of São Paulo, southeastern Brazil. *Precambrian Research*, 51, 351-374.

Janasi, V.A., Vlach, S.R.F. and Ulbrich, H.H.G.J. (1993). Enriched-mantle contributions to the Itú Granitoid Belt, Southeastern Brazil: evidence from K-rich diorites and syenites. *Anais da Academia Brasileira de Ciências*, 65, 107-118.

Janasi, V.A., Leite, R.J. and VanSchmus, W.R. U-Pb ages and stratigraphy of the granitic magmatism in the Agudos Grandes Batholith (W of São Paulo, Brazil): Implications for the evolution of the Ribeira Belt. *Journal of South American Earth Sciences* (submitted).

Jost, H., Theodoro, S.M.C.H., Figueiredo, A.M.G. and Boaventura, G.R. (1996). Propriedades geoquímicas e proveniência de rochas metassedimentares detríticas arqueanas dos *greenstone belts* de Crixás e Guarinos, Goiás. *Revista Brasileira de Geociências*, 26, 151-166.

Juliani, C. (1993). *Geologia, petrogênese e aspectos metalogenéticos dos grupos Serra do Itaberaba e São Roque na região das serras do Itaberaba e da Pedra Branca, NE da cidade de São Paulo.* Doctor thesis, IG-USP, São Paulo, Brazil, 803 p. (unpublished).

Juliani, C. and Beljavskis, P. (1995). Revisão da litoestratigrafia da faixa São Roque/Serra do Itaberaba (SP). *Revista do Instituto Geológico*, 16, 33-58.

Juliani, C., Pérez-Aguilar, A., Martin, M.A.B. and Beljavskis. P. (1996). Ocorrência e petrografia dos metarriólitos da Formação Nhanguçu - Grupo Serra do Itaberaba (SP). *Revista Brasileira de Geociências*, 26, 113-116.

Kennedy, M.J., Runnegar, B., Prave, A.R., Hoffmann, K.H. and Arthur, M.A. (1998). Two or four Late Proterozoic glaciations. *Geology*, 26, 1059-1063.

Lister, G.S., Etheridge, M.A. and Symonds, P.A. (1986). Detachment faulting and the evolution of passive continental margins. *Geology*, 14, 246-250.

Machado, N., Schrank, A., Abreu, F.R. Knauer, L.G. and Abreu, P.A.A. (1989). Resultados preliminares da geocronologia U-Pb na Serra do Espinhaço Meridional. In: *5th Simpósio de Geologia de Minas Gerais, Anais*, 171-174.

Machado, N., Valladares, C. Heilbron, M. and Valeriano, C. (1996). U-Pb geochronology of the central Ribeira Belt, Brazil and implications for the evolution of the Brazilian Orogeny. *Precambrian Research*, 79, 347-361.

Machado, R. and Demange, M. (1990). Reinterpretação estrutural e tectônica da região a leste da baía da Guanabara e a definição do batolito de Araruama (RJ). In: *36th Congresso Brasileiro de Geologia*, 6, 2744-2754.

Machado, R. and Demange, M. (1994). Classificação estrutural e tectônica dos granitóides neoproterozóicos do Cinturão Paraíba do Sul no Estado do Rio de Janeiro. *Boletim do Instituto de Geociências, Universidade de São Paulo, Série Científica*, 25, 81-96.

Machado, R. and Demange, M. (1998). Caracterização geoquímica e tectônica dos granitóides pré-colisionais Neoproterozóicos do cinturão Paraíba do Sul no Estado do Rio de Janeiro. In: H. Conceição (ed.) Contribuição ao Estudo dos Granitos e Rochas Correlatas. *Sociedade Brasileira de Geologia, Núcleo Bahia-Sergipe, Publicação Especial*, 5, 21-38.

Mantovani, M.S.M. and Shukowsky, W. (1996). Geofísica da litosfera (GEOLIT). Instituto Astronômico e Geofísico, USP, Atividades de Pesquisa, 99-102.

Marangoni, Y.R. (1994). *Modelo crustal para o norte de Goiás a partir de dados gravimétricos*. doctor thesis, IAG-USP, São Paulo, Brazil, 105 p. (unpublished).

Marini, O.J., Fuck, R.A., Danni, J.C.M., Dardenne, M.A. Loguercio, S.O.C. and Ramalho, R. (1984). As faixas de dobramentos Brasília, Uruaçu e Paraguai-Araguaia e o maciço mediano de Goiás. In: *Geologia do Brasil*, Texto explicativo do mapa geológico do Brasil e da área oceânica adjacente incluindo depósitos minerais, eds. Schobbenhaus, C., Campos, D.A., Derze, G.R. and Asmus, H.E., pp. 251-303. Escala 1:2 500 000. MME/DNPM.

Marshak, S. and Alkmim, F.F. (1989). Proterozoic contraction/extension tectonics of the southern São Francisco region, Minas Gerais, Brazil. *Tectonics*, 8, 555-571.

Martins-Neto, M.A. (1998). O Supergrupo Espinhaço em Minas Gerais: registro de uma bacia rifte-sag do Paleo/Mesoproterozóico. *Revista Brasileira de Geociências*, 28, 151-168.

Matte, Ph, Mattauer, M., Olivet, J.M. and Griot, D.A. (1997). Continental subductions beneath Tibet and the Himalayan orogeny: a review. *Terra Nova*, 9, 264-270.

Moraes, R. and Fuck, R.A. (1994). Deformação e metamorfismo das seqüências Juscelândia e Serra da Malacacheta, complexo Barro Alto, Goiás. *Revista Brasileira de Geociências*, 24, 189-197.

Moraes, R., Fuck, R.A. and Figueiredo, A.M. (1998). Litogeoquímica das rochas da Seqüência Juscelândia: evidência de rifte continental

Mesoproterozóico em Goiás. In: *40th Congresso Brasileiro de Geologia*, Anais, 35.

Nalini Jr., H.A., Bilal, E. and Correia Neves, J.M. (1998). Mineralogical, geochemical and isotopic constraints of Late Proterozoic Granitoids (Urucum and Galiléia suites) Eastern Minas Gerais State, Brazil. In: *14th International conference on Basement Tectonics, Extend Abstracts*, 44-46.

Offman, R.A. and Weber-Diefenbach, K. (1989). Two zoned complexes in the Iconha region, Espírito Santo, Brazil: A geochemical characterization of an intrusive series. *Zentralblatt für Geologie und Paläontologie*, 1, 903-916.

Paciullo, F.V.P. (1997). *A seqüência deposicional Andrelândia*. Doctor thesis UFRJ, Rio de Janeiro, Brazil, 265 p. (unpublished).

Park, J.K. (1994). Paleomagnetic constraints on the position of Laurentia from Middle Late Proterozoic to Early Cambrian times. *Precambrian Research*, 69 (1/4), 95-112.

Pedrosa-Soares, A.C., Noce, C.M., Vidal, P., Monteiro, R.L.B.P. and Leonardos, O.H. (1992). Toward a new tectonic model for the late Proterozoic Araçuaí (SE Brazil)-West Congolian (SW Africa) belt. *Journal of South American Earth Sciences*, 6, 33-47.

Pedrosa-Soares, A.C., Vidal, P., Leonardos, O.H. and Brito Neves, B.B. (1998). Late Proterozoic oceanic remnants in eastern Brazil: Further evidence and refutation of an exclusively ensialic evolution for the Araçuaí-West Congo orogen. *Geology*, 26, 519-522.

Piekarz, G.F. (1981). Reconhecimento de unidades correlacionáveis à sequência mineralizada do Perau, Estado do Paraná. In: *3rd Simpósio Regional de Geologia, Curitiba*, 1, 148-154.

Pimentel, M.M., Heaman, L., Fuck, R.A. and Marini, O.J. (1991). U-Pb zircon geochronology of Precambrian tin-bearing continental-type acid magmatism in central Brazil. *Precambrian Research*, 52, 321-335.

Pimentel, M.M. and Fuck, R.A. (1992a). Late Proterozoic crustal accretion in central Brazil. *Geology*, 20, 375-379.

Pimentel, M.M., Heaman, L. and Fuck, R.A. (1992b). Idade do meta-riolito da Seqüência Maratá, Grupo Araxá, Goiás: estudo geocronológico pelos métodos U-Pb em zircão, Rb-Sr e Sm-Nd. *Anais da Academia Brasileira de Ciências*, 64, 19-28.

Pimentel, M.M., Fuck, R.A. and Silva, L.J.H.D.R (1996a). Dados Rb-Sr e Sm-Nd da região de Jussara-Goiás-Mossâmedes (GO) e o limite entre os terrenos antigos do maciço de Goiás e o arco magmático de Goiás. *Revista Brasileira de Geocências*, 26, 61-70.

Pimentel, M.M., Fuck, R. A. and Alvarenga, C.J.S. (1996b). Post-Brasiliano (Pan-African) high-K granitic magmatism in Central Brazil: the role of Late Precambrian-early Paleozoic extension. *Precambrian Research*, 80, 217-238.

Pimentel, M.M., Whitehouse, M.J., Viana, M.G., Fuck, R.A. and Machado, N. (1997). The Mara Rosa Arc in the Tocantins Province: further evidence for Late Proterozoic crustal accretion in Central Brazil. *Precambrian Research*, 81, 299-310.

Pimentel, M.M., Fuck, R.A. and Botelho, N.F. (1999a). Granites and the geodynamic history of the Late Proterozoic Brasília belt, Central Brazil: A review. *Lithos*, 46, 463-483.

Pimentel, M.M., Dardenne, M.A., Viana, M.G., Gioia, B.M.C.L., Junges, S. and Seer, H.J. (1999b). Nd isotopes and the provenance of sediments of the Late Proterozoic Brasília Belt, Central Brazil: geodynamic implications. In: *2nd South American Symposium on Isotope Geology, Extend Abstracts*, 426-429.

Pires, F.A. (1991). Faciologia e análise paleoambiental da seqüência deposicional Furnas-Lageado (Group. Açungui), de idade proterozóica. *Revista Brasileira de Geociências*, 21, 355-362.

Poidevin, J.L., Viallet, Y. and Reis Neto, J.M. (1997). Radiometric ages (Pb-Pb) of the Ribeira belt calcareous rocks. In: *1st South-American Symposium on Isotope Geology, Extend Abstracts*, 242-243.

Preciozzi, F., Masquelin, H. and Basei, M.A.S. (1999). The Namaqua/Greenville terrane of eastern Uruguay. In: *2nd South-American Symposium on Isotope Geology, Extended Abstracts*, 338-340.

Queiroz, C.L., Jost, H. and McNaughton, N.J. (1999). U-Pb - SHRIMP ages of Crixás Granite-Greenstone belt terranes: from Archaean to Late Proterozoic. In: *7nd International Symposium on Tectonics of the SBG, Extended Abstracts*, 35-37.

Ramos, V.A. and Vujovich, G.I. (1995). New appraisal on the southwestern Gondwana terranes and their Laurentian affinities: Laurentia-Gondwanan connections before Pangea. In: *IGCP Field Conference Project 376, Abstracts*, 31-32.

Rego, I.T.S.F. (1989). *Petrologia e geoquímica da unidade charnockítica Bela Joana, região de São Fidélis, RJ*. Doctor thesis, Universidade de São Paulo, SP, Brazil, 348 p. (unpublished).

Reis Neto, J.M. (1994). *Faixa Itaiacoca: registro de uma colisão entre dois blocos continentais no Neoproterozóico*. Doctor thesis IG-USP, São Paulo, Brazil, 253 p. (unpublished).

Rivalenti, G., Girardi, V.A.V., Coltorti, M., Correia, C.T. and Mazzucchelli, M. (1989). Geochemical models for the petrogenesis of komatiites from Hidrolina greenstone belt, central Goiás, Brazil. *Journal of Petrology*, 30, 175-197.

Rocha-Campos, A.C. and Hasui, Y. (1981). Tilites of the Macaúbas Group (Proterozoic) in central Minas Gerais and southern Bahia, Brazil. In: *Earth pre-Pleistoceno glacial record*, eds. Hambrey, M.J. and Harland, W.B., pp. 933-938. Cambridge University Press.

Sad, J.H.G. and Dutra, C.V. (1988). Chemical composition of supracrustal rocks from Paraíba do Sul Group, Rio de Janeiro State, Brazil. *Geochimica Brasiliensis*, 2, 143-165.

Sadowski, G.R. and Bettencourt, J.S. (1996). Mesoproterozoic tectonic correlations between eastern Laurentia and the western border of the Amazon Craton. *Precambrian Research*, 76, 213-227.

Sato, K. (1998). *Evolução crustal da plataforma Sul Americana, com base na geoquímica isotópica Sm-Nd*. Doctor thesis, Universidade de São Paulo, SP, Brazil, 297 p. (unpublished).

Schmid, S.M., Pfiffner, O.A., Froitzheim, N. and Schönborn, G. (1996). Geophysical-geological transect and tectonic evolution of the Swiss-Italian Alps. *Tectonics*, 15, 1036-1064.

Schmitt, R.S., Trouw, R.A.J. and Van Schmus, W.R. (1999). The characterization of a Cambrian (~520 Ma) tectonometamorphic event in the coastal domain of the Ribeira belt (SE Brazil) – using U/Pb in syntectonic veins. In: *2nd South American Symposium on Isotope Geology, Extend Abstracts*, 363-366.

Seer, H.J., Dardenne, M.A. and Fonseca, M.A. (1998). Deformation and tectonic framework of the Meso/Late Proterozoic units of the southern Brasília fold belt. In: *14th International Conference on Basement Tectonics*, 57-61.

Sengör, A.M.C. (1990). Plate Tectonics and orogenic Research after 25 Years: A Tethyan Perspective. *Earth Science Reviews*, 27, 1-201.

Siga Jr., O. (1986). *A evolução geotectônica da porção nordeste de Minas Gerais, com base em interpretações geocronológicas*. Master dissertation, Universidade de São Paulo, SP, Brazil, 140 p. (unpublished).

Siga Jr., O., Basei, M.A.S. and Machiavelli, A. (1993). Evolução geotectônica da porção NE de Santa Catarina e Se do Paraná, com base em interpretações geocronológicas. *Revista Brasileira de Geociências*, 23, 215-223.

Siga Jr., O., Basei, M.A.S., Reis Neto, J.M., Machiavelli, A. and Harara, O.M. (1995). O Complexo Atuba: um cinturão Paleoproterozóico intensamente retrabalhado no Neoproterozóico. *Boletim do Instituto de Geociências, Universidade de São Paulo, Série Científica*, 26, 69-98.

Siga Jr., O., Basei, M.A.S., Reis Neto, J.M., Harara, O.M., Passarelli, C.R., Prazeres Filho, H.J., Weber, W. and Machiavelli, A. (1997). Ages and tectonic setting of alkaline-peralkaline granitoids of Paraná and Santa Catarina, States, southern Brazil. In: *South-American Symposium on Isotope Geology, Extended Abstracts*, 301-303.

Silva Filho, B.C. and Soliani Jr., E. (1987). Origem e evolução dos gnaisses Cambaí: Exemplo de estudo integrado de análise estrutural, petroquímica e geocronologia. In: *3rd Simpósio Sul*

Brasileiro de Geologia, **1**, 127-145.

Silva, L.C., Hartmann, L.A., McNaughton, N.J. and Fletcher, I.R. (1999). SHRIMP U/Pb zircon dating of Late Proterozoic granitic magmatism and collsion in the Pelotas Batholith, southernmost Brazil. *International Geologic Review*, **41**, 531-551.

Simões, L.S.A. (1995). *Evolução tectonometamórfica da nappe de Passos, sudoeste de Minas Gerais*. Doctor thesis, IG-USP, São Paulo, Brazil, 149 p. (unpublished).

Simões, L.S.A. and Fuck, R.A. (1984). Estratigrafia, deformação e metamorfismo do Grupo Araxá na região de Mossâmedes, Goiás. In: *33rd Congresso Brasileiro de Geologia*, 7, 3181-3195.

Slüitner, Z. and Weber-Diefenbach, K. (1989). Geochemistry of charnockitic granulites and associated amphibolitic gneisses in the coastal region of Espírito Santo, Brazil. *Zentralblatt für Geologie und Paläontologie*, **1**, 917-931.

Soares, P.C., Fiori, A.P. and Carvalho, S.G. (1990). Tectônica colisional oblíqua entre o bloco Paraná e a margem sul do cráton do São Francisco, no maciço de Guaxupé(1). In: *36th Congresso Brasileiro de Geologia*, **6**, 2723-2734.

Souza, A.P. (1990). *Mapa geológico na escala 1:50.000 e esboço da evolução tectônica e sedimentar do Grupo Itaiacoca, nas Folhas Barra do Chapéu e Ouro Verde - SP/PR*. Master dissertation, IG-USP, São Paulo, Brazil, 200 p. (unpublished).

Söllner, F., Lammerer, B. and Weber-Diefenbach, K. (1987). The Brasiliano Orogenesis: Age determination (Rb-Sr and U-Pb) in the coastal mountain region of Espírito Santo, Brazil. *Zentralblatt für Geologie und Paläontologie*, **I**, 729-741.

Söllner, F., Lammerer, B. and Weber-Diefenbach, K. (1989). Brasiliano age of a charnoenderbitic rock suite in the Complexo Costeiro (Ribeira Mobile Belt), Espírito Santo/Brazil: Evidence from U-Pb geochronologieon zircons. *Zentralblatt für Geologie und Paläontologie*, **I**, 993-945.

Söllner, F., Lammerer, B. and Weber-Diefenbach, K. (1991). Die krustenenwicklung in der küstenregion nördlich von Rio de Janeiro/Brasilien. *Münchner Geologische Hefte*, **4**, 1-100.

Söllner, F. and Trouw, R.A.J. (1997). The Andrelândia depositional cycle (Minas Gerais/Brazil), a post-Transamazonic sequence south of the São Francisco craton: evidence from U-Pb dating on zircons of a metasediment. *Journal of South American Earth Sciences*, **10**, 21-28.

Spear, F.S. (1995). Metamorphic Phase Equilibria and Pressure-Temperature-Time Paths. *Mineralogical Society of America*, 1-799.

Stanistreet, I.G., Kukla, P.A. and Henry, G. (1991). Sedomentary basinal responses to a Late Precambrian Wilson Cycle: the Damara orogen and Nama foreland, Namibia. *Journal of African Earth Sciences*, **13**, 141-156.

Strider, A.J. and Nilson, A.A. (1992). Melange ofiolítica nos metassedimentos Araxá de Abadiânia (GO) e implicaçoes tectônicas regionais. *Revista Brasileira de Geociências*, **22**, 204-215.

Suita, M.T.F., Kamo, S.L., Krogh, T.E., Fyfe, W.S. and Hartmann, L.A. (1996). U-PB ages from the high-grade Barro Alto mafic-ultramafic complex (Goiás, central Brazil): Middle Proterozoic continental mafic magmatism and Upper Proterozoic continental collision. In: *8th International Conference on Geochronology, Cosmochronology, and Isotope Geology, US Geological Survey Circular 1107, Abstracts*, 309 p.

Tassinari, C.C.G. and Campos Neto, M.C. (1988). Precambrian continental crust evolution of southeastern São Paulo State – Brazil: Based on isotopic evidences. *Geochimica Brasiliensis*, **2**, 175-183.

Tassinari, C.C.G., Nutman, A.P., VanSchmus, W.R., Bettencourt, J.S. and Taylor, P.N. (1996). Ceochronologial sytematics on basement rocks from the Rio Negro-Juruena Province (Amazonian Craton) and tectonic implications. *International Geological Review*, **38**, 161-175.

Tassinari, C.C.G., Cordani, U.G., Correia, C.T., Nutman, A., Kinny, P. and Dians Neto (1999). Dating of granulites by SHRIMP U-Pb systematics in Brazil: constrains for the age of the metamorphism of Proterozoic orogenies. In: *2nd South-American Symposium on Isotope Geology, Extended Abstracts*, 371-375.

Teixeira, A.L. (1995). *Ambientes geradores dos sedimentos da Bacia Eleutério*. Unpublished master dissertation, IG-USP, São Paulo, Brazil, 122 p.

Teixeira, W., Tassinari, C.C.G., Cordani, U.G. and Kawashita, K. (1989). A review of the geochronology of the Amazonian Craton: tectonic implications. *Precambrian Research*, **42**, 213-227.

Teixeira, W., Cordani, U.G., Nutman, A.P. and Sato, K. (1998). Polyphase Archaean evolution in the Campo Belo metamorphic complex, Southern São Francisco craton, Brazil: SHRIMP U-Pb zircon evidence. *Journal of South American Earth Sciences*, 11, 279-289.

Tommazi, A., Vauchez, A., Fernandes, L.A.D. and Porcher, C.C. (1994). Magma-assisted strain localization in an orogen-parallel transcurrent shear zone of southern Brazil. *Tectonics*, 132, 421-437.

Trompette, R. (1994). *Geology of Western Gondwana (2000-500 Ma)*. Balkema, 352 p.

Trouw, R.A.J., Ribeiro, A. and Paciullo, F.V.P (1983). Geologia estrutural dos Grupos São João del Rei, Carrancas e Andrelândia, sul de Minas Gerais. *Anais da Academia brasileira de Ciências*, **55**, 71-85.

Trouw, R.A.J., Ribeiro, A. and Paciullo, F.V.P (1986). Contribuição à geologia da folha Barbacena 1:250.000. In: *34th Congresso Brasileiro de Geologia*, **2**, 974-986.

Tupinambá, M. (1999). Evolução tectônica e magmática da Faixa Ribeira na região serrana do Estado do Rio de Janeiro. Doctor thesis, Universidade de São Paulo, São Paulo, Brazil, 221 p. (unpublished).

Tupinambá, M., Teixeira, W., Heilbron, M. and Basei, M.A.S. (1998). The Pan-African/Brasiliano arc-related magmatism at the Costeiro Domain of the Ribeira Belt, Southeastern Brazil. In: *14th International Conference on Basement Tectonics, Extend Abstracts*, 12-14.

Turpin, L., Maruejol, P. and Cuney, M. (1988). U-Pb, Rb-Sr and Sm-Nd chronology of granitic basement, hydrothermal albitites and uranium mineralization (Lagoa Real, south Bahia, Brazil). *Contribution to Mineralogy and Petrology*, **98**, 139-147.

Uhlein, A., Trompette, R.R., and Egydio-Silva, M. (1998). Proterozoic rifting and closure, SE border of the São Francisco Craton, Brazil. *Journal of South American Earth Sciences*, **11**, 191-203.

Valeriano, C.M. (1992). *Evolução tectônica da extremidade meridional da Faixa Brasília, região da represa de Furnas, sudoeste de Minas Gerais*. Doctor thesis, IG-USP, São Paulo , Brazil, 192 p. (unpublished).

Valeriano, C.M., Simões, L.S.A., Teixeira, W. and Heilbron, M. (1998). Southern Brasília Belt (SE Brazil): thrust-discontinuities and evolution during the Late Proterozoic Brasiliano orogeny. In: *14th International Conference on Basement Tectonics*, 62-65.

Valladares, C.S., Heilbron, M., Figueiredo, M.C.H. and Teixeira, W. (1997). Geochemistry and geochronology of paleoproterozoic gneissic rocks of the Paraíba do Sul Complex (Quirino Unit), Barra Mansa region, Rio de Janeiro, Brazil. *Revista Brasileira de Geociências*, **27**, 11-121.

Vannay, J-C. and Grasemann, B. (1998). Inverted metamorphism in the High Himalaya of Himachal Pradesh (NW India): phase equilibria *versus* thermometry. *Schweizerische Mineralogische und Petrographische Mitteilugen*, **78**, 107-132.

Vandecar, J.C., James, D.E. and Assumpção, M. (1995). Seismic evidence for a fossil mantle plume beneath South America and implications for plate driving forces. *Nature*, **378**, 25-31.

Vasconcellos, A.C.B.C. (1988). *O Grupo Andrelândia na região norte de Ouro Fino, MG*. Master dissertation, IG-USP, São Paulo, Brazil, 199 p. (unpublished).

Vasconcellos, A.C.B.C., Harris, N.B.W. and Tindle, A.C. (1991). The relationships between metamorphism and tectonics: evidence from the Socorro-Guaxupé Thrust nappe, southeastern Brazil. In: Metamorphism, deformationa and structure of the crust, eds. Tuisku, P. and Laajaki, K. *Res Terrae, Publication Department of Geology, University of Oulu, Finland,* **A-5**, 86.

Vlach, S.R.F., Janasi, V.A. and Vasconcellos, A.C.B.C. (1990). The Itú belt: associated calc-alkalic and aluminous A-type late Brasiliano granitoids in the States of São Paulo and Paraná, southern Brazil. In: *36th Congresso Brasileiro de Geologia*, **4**, 1700-1711.

Weil, A.B., Van der Voo, R., Niocaill, C.M. and Meert, J.G. (1998). The Proterozoic supercontinent Rodinia: paleomagnetically derived reconstructions for 1100 to 800 Ma. *Earth and Planetary Science Letters*, **154**, 13-24.

Wernick, E. (1998). The pluriserial Ribeira Magmatic System 590, SE/S Brazil and Uruguay. *Revista Brasileira de Geociências*, **28**, 533-542.

Wiedemann, C., Bayer, P., Horn, H., Lammerer, B, Ludka, I.P., Schmidt-Thomé and Weber-Diefenbach, K. (1986). Maciços intrusivos do sul do Espírito Santo e seu contexto regional. *Revista Brasileira de Geociências*, 16, 24-37.

Wiedemann, C., Mendes, J.C., Moura, J.C., Nascimento, R.C.C. and Ludka, L.P. (1997). Granitoids of the Espírito Santo Magmatic Arc. In: *2nd International Symposium on granites and Associated Mineralizations, Excursion guide*, 57-76.

Wilson, T.J., Hanson, R.E. and Wardlaw, M.S. (1993). Late Proterozoic evolution of the Zambezi belt, Zambia: implications for regional Pan-African tectonics and shear displacements in Gondwana. In: *Gondwana 8-Assembly, Evolution and Dispersal*, eds. Findlay, R.H., Urung, R., Banks, M.R. and Veevers, J., pp. 69-82. Rotterdam, Balkema.

Yamato, A.A. (1999). *Mapeamento Geológico de parte da Folha Bocaiúva do Sul - PR (SG.22-X-D-I-2), escala 1:50.000.* Master dissertation, IG-USP, São Paulo, Brazil, 108 p. (unpublished).

Zimbres, E., Kawashita, K. and VanSchmus, W.R. (1990). Evidências de um núcleo Transamazônico na região de Cabo Frio, RJ e sua correlação com o cráton de Angola. In: *36th Congresso Brasileiro de Geologia*, **6**, 2735-2743.

Phanerozoic Cratonic Cores, Marginal Belts and Sedimentary Basins

"Lower Paleozoic magmatism in South America is most clearly recorded in outcrops in central and northwestern Argentina, extending into southern Bolivia, northeastern Chile and the coast of southwestern Peru. Igneous and metamorphic rocks of this age are well exposed in the geological provinces of central and northwestern Argentina, and its characteristics are considered representatives of the Paleozoic evolution of southwestern Gondwana... This distribution indicates that the basement of the southwestern sector of the South American Continent was mainly developed during the Early Paleozoic, being the Pampean and the Famatinian the best recognized orogenic episodes. These large Paleozoic regions are the basement that structurally controlled the Meso-Cenozoic Andean magmatism associated with the Pacific subduction."

(Rapela, this volume)

"Regarding the distribution of Phanerozoic sedimentary successions and particular plate-tectonic regimes, four major domains are recognized in the South American Plate: the continental interior; the western, convergent margin of the plate; the eastern side, a more than 10 000 km-long divergent margin; and the northern and southern margins of the plate, marked by a regional transcurrent tectonic regime..."

(Milani and Thomaz Filho, this volume)

Patagonia

Victor A. Ramos and María B. Aguirre-Urreta

Patagonia is a unique sector of southern South America that records a complex history of orogeny, continental break-up, seismic ridge collision, and foreland basin development. The geographical/physiographical limits of Patagonia are the Colorado River at about 39 °S in the N and the foothills of the Patagonian Cordillera along the western and southern margins; the crystalline Patagonian domain extends towards the E and constitute the basement for the Argentinean continental margin.

Patagonia encompasses several geological units such as the Somun Cura and Deseado massives, the San Jorge Basin, the Rio Mayo Embayment, and the Austral Basin (Fig. 1). The Patagonides foothills comprise the pre-Andean Patagonian Precordillera, the Bernárdides Mountains, and an extra-Andean region of basaltic plateaux (*mesetas*).

The Somun Cura Massif

This geological province was recognized as a stable cratonic block since the pioneer work of Windhausen (1931), who defined it as the Patagonian Massif. Its western boundary coincides with the Patagonian Precordillera, and is outlined by the Andean Thrust Front, that subdivided a western area with deformed Cenozoic rocks and a more stable area with mild or nihil deformation (Fig. 2).

Towards the SW, the Somun Cura Massif is bounded by the Cañadón Asfalto Basin, and is overlapped by the deposits of the Chubut Group, of Early Cretaceous age. The metamorphic basement consists of gneiss and micaschist metamorphosed in the amphibolite facies, associated with syn-tectonic granitoid plutons. Rb/Sr ages determinations from these rocks gave 850 ± 50 Ma (Sr^{87}/Sr^{86} initial ratio of 0.70734) according to Linares *et al.* (1990) and 620 ± 45 Ma (Sr^{87}/Sr^{86} initial ratio of 0.7038) by Varela *et al.* (1997). These rocks are related to the low-grade El Jagüelito Schist that has been considered by some authors as part of an Early Paleozoic basement. However, equivalent low-grade metamorphic rocks have been dated by Linares *et al.* (1990) at 600 ± 25 Ma (Sr^{87}/Sr^{86} initial ratio of 0.70347). New U/Pb dating on zircon from the southwestern sector of the massif indicated even older Middle Proterozoic ages.

Early Paleozoic marine deposits of Sierra Grande Formation unconformably overlie the eastern sector of the massif (Cortés *et al.*, 1984). Orthoquarzite is interbedded with important zones of oolitic iron beds up to 14 m thick. This cover, of Silurian age, forms a stable platform of clastic beds with minor tuffaceous levels, and the sedimentary rocks have a typical Malvinokaffric fauna. The fauna is characterized by the *Clarkeia antisiensis - Heterorthella* association of Silurian (Wenlockian) age (Manceñido and Damborenea, 1984).

Both the basement, and partially the cover, are intensely intruded by Ordovician, Carboniferous, and Permian granitoid bodies. Some authors recognize two different petrographic suites: an older suite consisting of subduction related tonalite, granodiorite and granite; and a second suite consisting of post-tectonic granite intrusives (Rapela and Caminos, 1987). New Rb/Sr ages from these assemblages gave 483 ± 22 Ma and 467 ± 16 Ma, in granodiorite and granite, and 363 ± 57 Ma and 318 ± 28 Ma in granite bodies emplaced in the Early Paleozoic cover (Varela *et al.*, 1997).

These rocks are covered by widespread acid volcanic and pyroclastic rocks that constitute an extensive rhyolitic plateau. Middle to Late Triassic ages have been recorded in the northwestern sector, associated with *Dicroidium* flora (Stipanicic *et al.*, 1968; Artabe, 1985, 1986). The rhyolitic rocks are younger towards the Atlantic coast. Rb/Sr ages obtained in several different areas of the massif yielded ages from 183 ± 2 to 178 ± 1 Ma (Rapela and Pankhurst, 1993). These ages were confirmed by Ar/Ar dating which range from 186.2 ± 1.5 to 176.9 ± 0.8 Ma in the Marifil Formation (Alric *et al.*, 1996). Both isotope studies pointed to an important short-lived melt episode that involved most of the Somuncura Massif, associated with the break-up of Gondwana (Kay *et al.*, 1993). These rocks are also closely linked with Early Jurassic granitoid with similar composition and tectonic setting (Rapela *et al.*, 1991).

Thermal sag subsidence following the Jurassic extension produced the Chubut Group continental deposits of Early Cretaceous age (Stipanicic *et al.*, 1968; Spalletti *et al.*, 1989). Several shallow marine transgressions in the Maastrichtian (Andreis *et al.*, 1989; Spalletti, 1990), Eocene, and Neogene covered most of the Atlantic side of the massif (Legarreta and Uliana, 1994). A large alkaline flood basalt province covered the Somun Cura Massif at about 27 Ma (Ardolino, 1981). These flows have been interpreted as produced by a transient hot-spot like environment by Kay *et al.* (1993), that in a few million years extruded a large volume of basalt. Late volcanic activity of alkaline basalt and rhyolite is concentrated near Telsen (Corbella, 1984) and in the Sierra de Apas, Los Chacays, Talagapa, and Pire Mahuida (Salani and Page, 1990), in volcanic domes, annular dykes, and minor flows.

The dominant structure of the massif is characterized by half-graben systems, associated with the opening of the South Atlantic Ocean. The orientation of the extension was oblique to the continental margin, and a partition of the stress in strike-slip displacements is evident (Ciciarelli, 1989). The western sector has a mild tectonic inversion produced during the Andean compression.

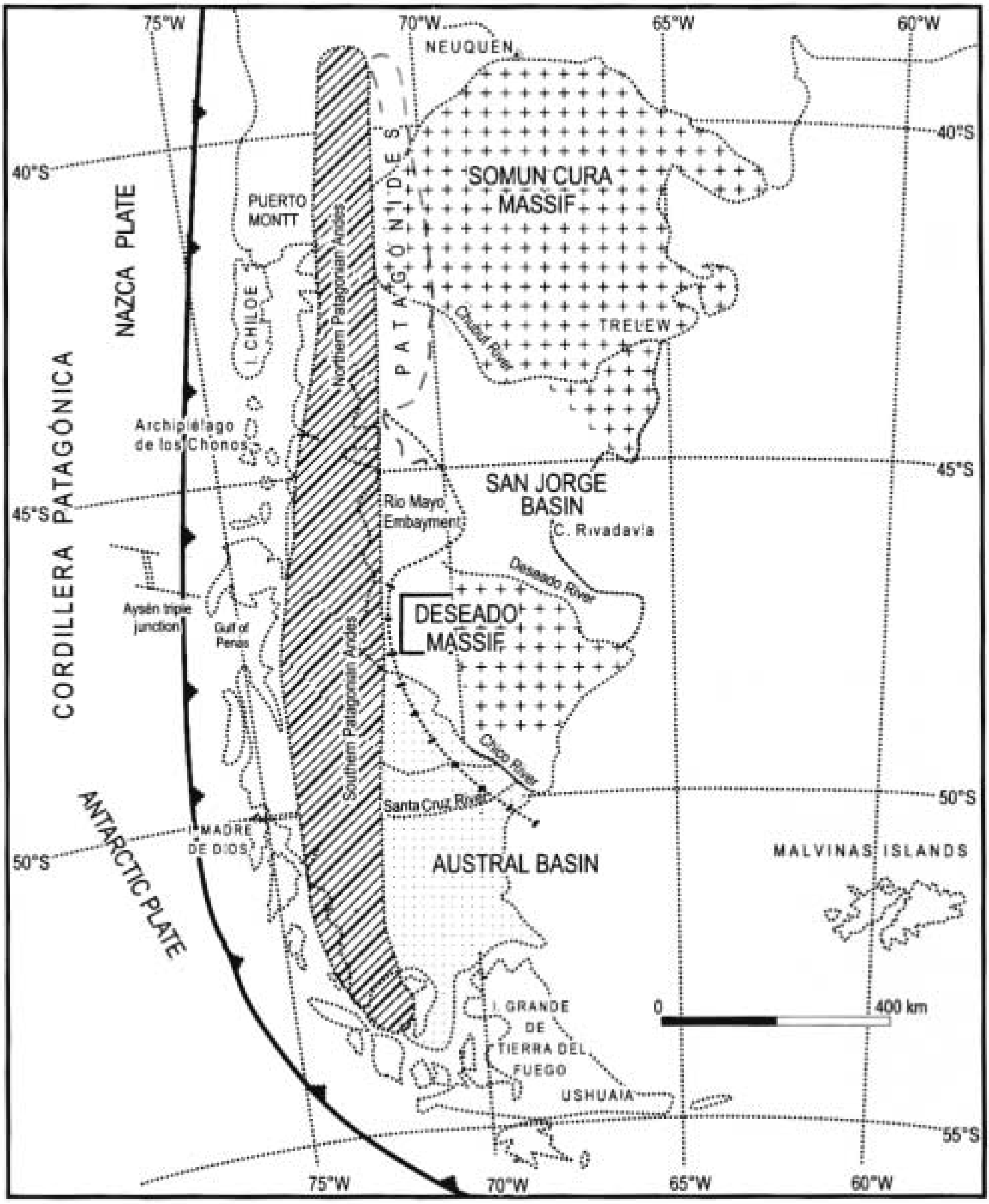
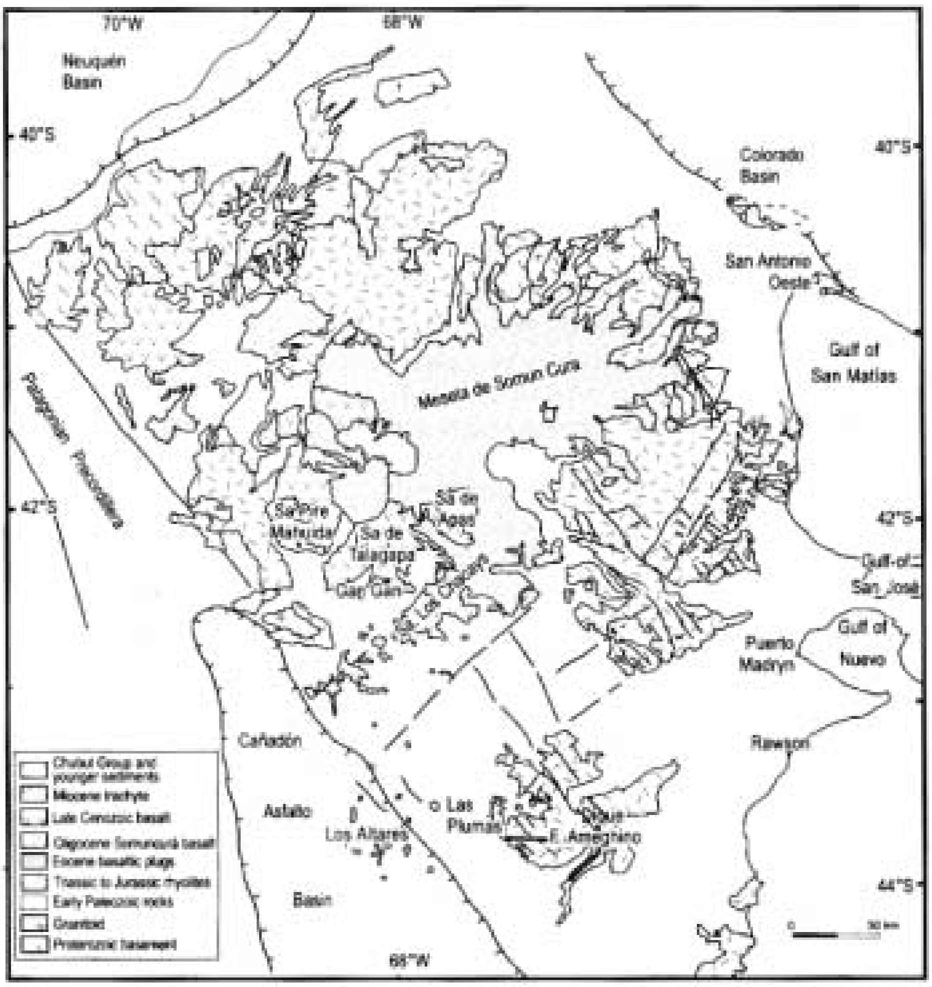


Fig. 1 - Major geological provinces of Patagonia (modified after Windhausen, 1931 ; Groeber, 1938).

Fig. 2 - Main geological features of the Somun Cura Massif in northern Patagonia, and adjacent geological provinces.

Fig. 3 - Main geological features of the Deseado Massif in southern Patagonia (modified after Panza et al., 1994).


70°W
68°W
Neuquén Basin
40°S
Colorado Basin
San Antonio Oeste
Gulf of San Matías
Meseta de Somun Curá
Patagonian Precordillera
Sa Pire Mahuida
Sa de Agua
Sa de Talagapa
Gan Gan
42°S
Gulf of San José
Gulf of Nuevo
Puerto Madryn
Rawson
Cañadón Asfalto Basin
Los Altares
Las Plumas
F. Ameghino
44°S
Chubut Group and younger sediments
Miocene trachyte
Late Cenozoic basalt
Oligocene Somuncurá basalt
Eocene basaltic plugs
Triassic to Jurassic rhyolites
Early Paleozoic rocks
Granitoid
Proterozoic basement


70°W
48°S
Puerto San Julián
50 km
Puerto Santa Cruz
68°W
50°S

The Deseado Massif

This region has been considered as an ancient and independent massif from the Somun Cura area since the work of Feruglio (1946). Leanza (1958) gave the present name to the stable basement area comprised between the Deseado and Chico de Santa Cruz rivers (Fig. 3).

The basement consists of isolated outcrops of phyllite and low-grade schist (Di Persia, 1962) that have been dated at 540 ± 20 Ma by K/Ar method (Pezzuchi, 1978). Silurian granitoid intrusives and sub-volcanic rocks are emplaced in the metamorphic rocks (Palma, 1989; Márquez *et al.*, 1994). Recent U/Pb dating on zircon gave ages of 424, 454 and 472 Ma (Middle-Late Ordovician to Silurian) in granitoid dykes of the Tres Hermanas region. Some tonalitic rocks yield ages between 407 and 402 Ma (Devonian). Detrital zircon fractions from metaquartzite samples gave a $^{206}Pb / ^{238}U$ of 903 Ma (Loske *et al.*, 1999).

On this basement, minor continental basins of La Golondrina and El Tranquilo were developed in Permian and Triassic times (Stipanicic and Reig, 1956; Archangelsky, 1959; Arrondo, 1972). Red arkose, shale, and tuff of these basins have been interpreted as rift sequences by Jalfin (1987) and Jalfin and Herbst (1995). These deposits have a rich Triassic dinosaur fauna in the El Tranquilo Basin (Bonaparte and Vince, 1979) and well-preserved flora and microflora (Herbst, 1988; Zavatieri, 1993).

These deposits are intruded by Late Triassic and Early Jurassic granite of the Central Patagonian Batholith (Stipanicic *et al.*, 1972; Rapela *et al.*, 1991). The extensional regime was related to the break-up phases associated with the South Atlantic opening that developed the large Chon Aike Rhyolitic Province (Kay *et al.*, 1989). The Chon Aike rhyolite and ignimbrite, with scarce basalt, is associated with a clastic continental sequence deposited in half-graben systems (Ramos, 1996). The age of the rhyolite varies from 168 to 170 Ma in the eastern sector of the massif (Pankhurst and Rapela, 1993).

These clastic and volcanic rocks are covered by younger ignimbrite flows, volcanic and clastic sequences of Early Cretaceous age, that indicate a reactivation of the extension in the northwestern sector of the massif (Palma, 1989). Atlantic marine transgressions of Paleocene, Eocene and Oligocene-Miocene age are interbedded with continental clastic and pyroclastic sequences (Legarreta and Uliana, 1994).

The structure of the region displays an intense penetrative deformation in the metamorphic basement (Panza *et al.*, 1994); extensional structures in the Permian and Mesozoic deposits (de Giusto *et al.*, 1980); and sub-horizontal Late Cretaceous-Cenozoic sequences. The Jurassic rocks are associated with important gold low-sulfidation epithermal systems at Cerro Vanguardia and related ore-districts (Genini, 1989; Schalamuk *et al.*, 1995).

The Patagonides

The pioneer work of Keidel (1921) recognized the Patagonides, an old mountain chain situated between the Northern Patagonian Andes and the Somun Cura Massif. The present definition of the Patagonides is mainly based on the work of Groeber (1938) and Frenguelli (1946). This geological province comprises Paleozoic and Mesozoic sedimentary deposits and associated igneous rocks deformed during the latest Cretaceous (Fig. 4). The different ranges of the Patagonides can be subdivided in two distinct morphostructural units, the Patagonian Precordillera and the Bernárdides.

The Patagonian Precordillera

This region comprises a series of pre-Andean mountains developed between the Ñirihuau Basin at the foothills of the Patagonian Andes and the Somun Cura Massif (40°S to 43°30'S). The stratigraphy was established by Franchi and Page (1980), and the main structural features are shown in the Figure 4.

The basement consists of Precambrian metamorphic rocks exposed in the Gastre region that is intruded by Early Paleozoic granitoid plutons (Dalla Salda *et al.*, 1994). These Precambrian rocks are the basement of the Tepuel Late Paleozoic marine basin (Suero, 1962), later defined as a composite basin to include the Jurassic deposits by Ugarte (1966). The Late Paleozoic rocks consist of glacial, glaciomarine, marine and continental deposits several thousand metres thick, developed in a back-arc extensional setting (Ramos, 1983).

Black shale and sandstone beds bearing Late Pliensbachian to Early Toarcian ammonites and bivalves (Riccardi, 1983; Hillebrandt, 1987) are exposed between Esquel and the Sierra de Payaniyeu. These Liassic rocks interfinger with continental and volcaniclastic deposits (Cortiñas, 1984), defining an extensional intra-arc basin. The Liassic rocks are covered by the Middle to Late Jurassic Lonco Trapial volcanics (Nullo, 1983), which correspond to an extra-Andean volcanic arc. This arc is partially coeval with the Lago La Plata Formation along the axis of the Patagonian Cordillera (Ramos, 1983).

Granitoid bodies along the extra-Andean region represent the large Central Patagonian Batholith of Early to Middle Jurassic age, emplaced along extensional faults trending W-NW (Rapela *et al.*, 1991). Lacustrine limestone and black shale of the Cañadón Asfalto Group define a basin trending NW deposited in a half-graben system during Middle to Late Jurassic time (Figari and Courtade, 1993). This renewed episode of extension is obliquely positioned to the E of the Liassic intra-arc basin. An extensive cover of late Early Cretaceous continental deposits of the Chubut Group unconformably overlies the Jurassic rocks. These continental deposits are interfingered with thick tuff layers that are the piroclastic equivalent of the Early Cretaceous volcanic arc of the Divisadero Group, well exposed along the axis of the Patagonian Cordillera (Ramos and Drake, 1987).

The eastern area is also covered by Maastrichtian-Danian marine deposits related to an Atlantic short-lived transgression. The western part is covered by Late Cretaceous tholeiitic basalt flows. Paleogene alkaline basalt related to a ridge subduction are developed S of 43°S (Ramos and Kay, 1992). The structure of the

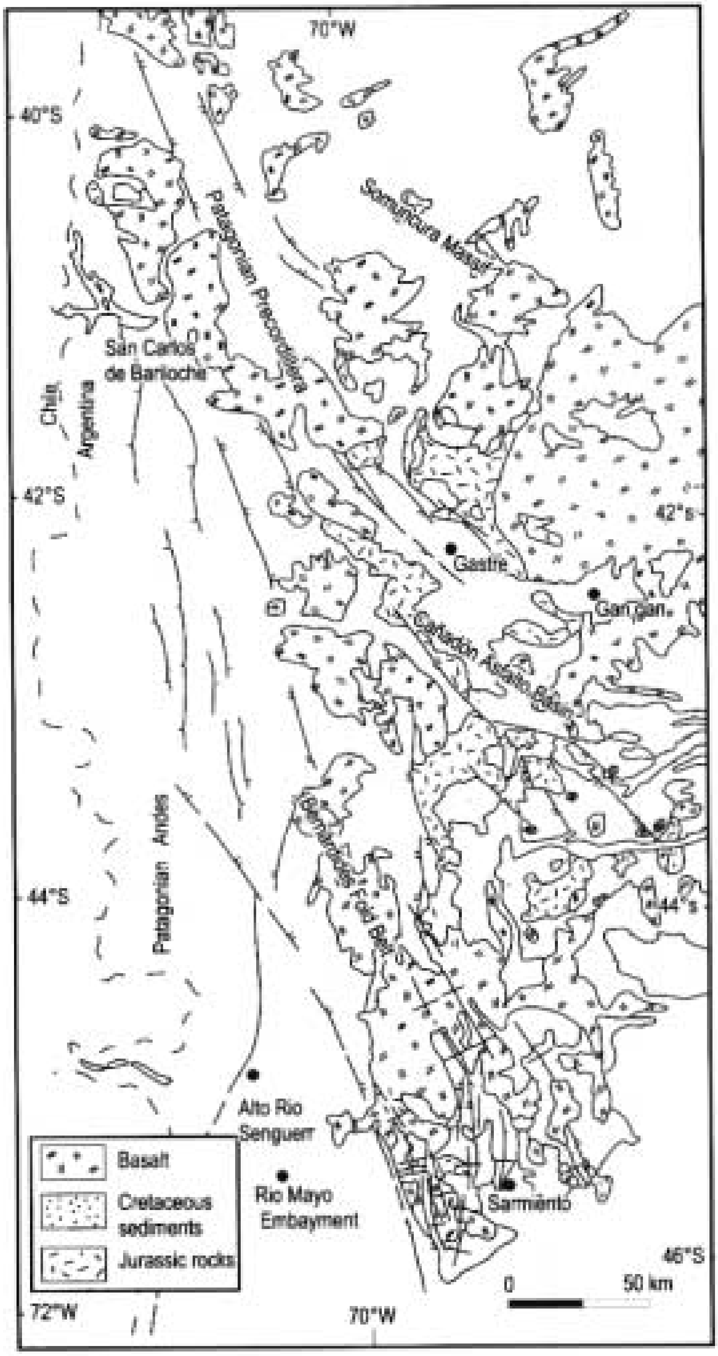
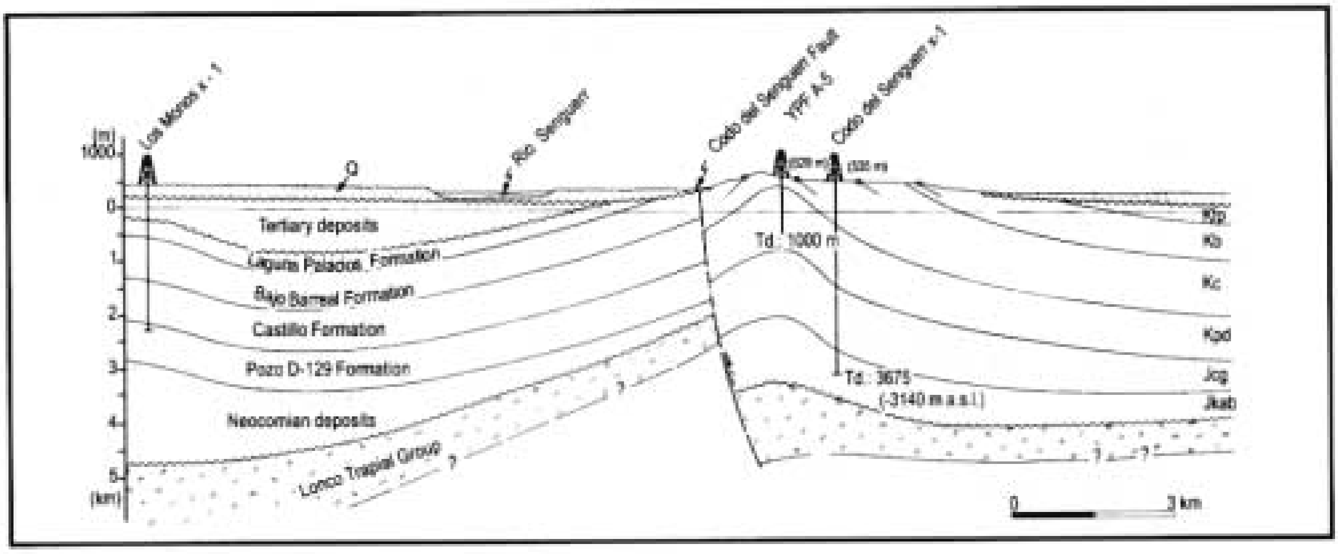


Fig. 4 - Main geological features of Patagonides with the Patagonian Precordillera and the Bernárdides Fold Belt (after Barcat et al., 1984; Figari and Courtade, 1993).

Fig. 5 - Structural section of Codo del Río Senguerr Anticline produced by tectonic inversion of Mesozoic half-grabens (after Ramos, 1999). Note the contrast between the Lonco Trapial Group and Neocomian synrift deposits with the younger sag facies.



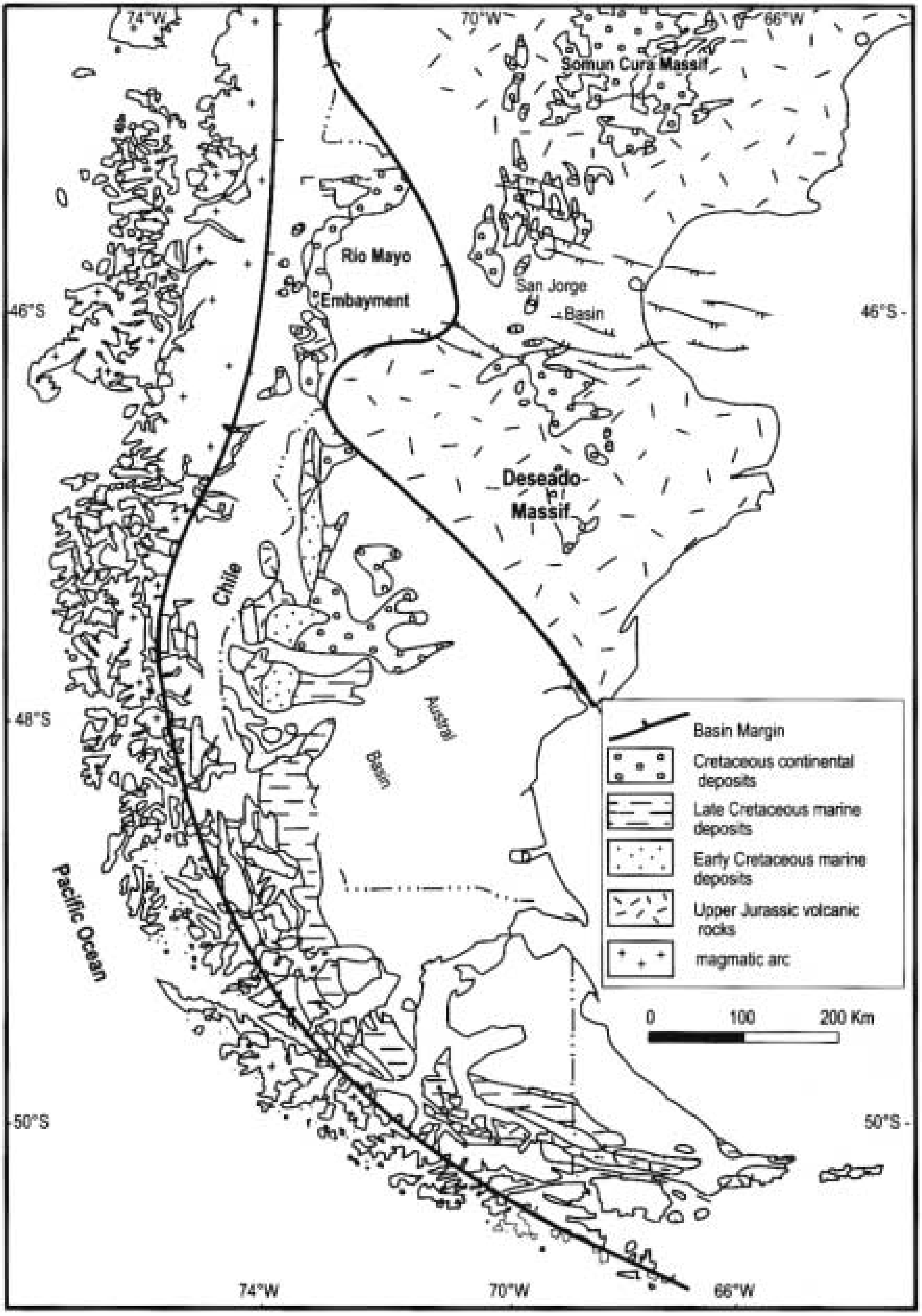


Fig. 6 - Main geological features of southern Patagonia, with main exposures of the Rio Mayo Embayment and the Austral Basin (modified after Riccardi, 1988).

Patagonian Precordillera is characterized by tectonic inversion of previous normal faults with a dominant westerly vergence (Márquez and Giacosa, 1999). Due to the oblique trend of the normal faults, there are important strike-slip displacements (Coira *et al.*, 1975).

The Bernárdides

A series of ranges in the central extra-Andean Patagonia (43°30'S to 46°S) has been grouped as the Bernárdides by Ferello (1969). These ranges are separated from the Patagonian Andes by the depression of the Genoa and Senguerr River valleys. The eastern boundary coincides with Lake Colhue Huapi.

The Bernárdides fold belt is characterized by large anticlinal structures, such as the Codo del Río Senguerr (Fig. 5) and the Sierra de Castillo, bounded by N-NW-trending faults developed by tectonic inversion of early Mesozoic half-grabens (Homovc *et al.*, 1993). The anticlinal structures have vergence to the E or to the W, following the dip of the previous normal faults. The initial fill of these half-grabens is exposed to the N, in the Patagonian Precordillera. The fill consists of Liassic deposits, Middle Jurassic volcanics and lacustrine Pre-Cretaceous shale. The sag phase deposits are represented by the late Early Cretaceous Chubut Group (Sciutto, 1981). These continental and pyroclastic sequences, the oldest units exposed in the Bernárdides, are famous for their abundant dinosaur remains (Bonaparte and Gasparini, 1978). Most of the type localities of the different units of the Chubut Group, as defined by Lesta (1968), are well exposed in the region. These deposits are partially covered by extensive basalt flows, domes, and alkaline intrusives of Paleogene age described by Ferello (1969). The marine deposits are only preserved in the easternmost localities, as the region was a positive area during most of the Cenozoic.

The compressive structure of the region was produced by tectonic inversion and strike-slip displacements in the Late Cretaceous and reactivated in Cenozoic times (Barcat *et al.*, 1984). Although the exposures of these structures end at the Codo del Río Senguerr, it is possible to continue this trend farther S in the subsurface (Figari *et al.*, 1998).

The San Jorge Basin

This E-W-trending, pericratonic continental basin was developed as an aulacogenic basin during the opening of the South Atlantic (de Wit, 1977). The basin was formed between the Somun Cura and Deseado massives. The basement of the basin is exposed in its northern and southern flanks, within the Somun Cura and Deseado massives. It consists of low and high grade metamorphic rocks of Precambrian age intruded by Early Paleozoic granitoid plutons.

An extensional regime started in Early Mesozoic times, with the development of a series of W-NW-trending half-graben systems. The very initial fill of these grabens comprised Early Jurassic deposits and Middle Jurassic volcanics, similar to the previously described syn-rift deposits of the Bernárdides (Fitzgerald *et al.*, 1990; Figari *et al.*, 1996). A late syn-rift sequence is mainly composed of continental fluvial deposits of Neocomian age, that in the western side represent a short-lived marine transgression during the Valanginian (Figari *et al.*, 1999).

Most of the basin fill consists of continental fluvial deposits of the Chubut Group of Early Cretaceous age (Lesta *et al.*, 1980). These rocks correspond to the thermal sag phase of the basin. Conformably overlying the Chubut Group there are Maastrichtian-Danian shallow marine deposits, continental Paleogene tuff beds and clastic deposits of the Sarmiento Group, the late Oligocene-Miocene Patagonian marine transgression and the continental Santa Cruz Formation of early-middle Miocene age (Legarreta and Uliana, 1994). The basin is covered by thick gravel blankets, and isolated patches of basaltic rocks.

The structure consists of a series of gentle anticlines, partially bounded by high angle reverse faults, produced by tectonic inversion of previous normal faults (Fitzgerald *et al.*, 1990; Homovc *et al.*, 1993).

The Río Mayo Embayment

A Neocomian basin developed between the Patagonian Andes and the Bernárdides to the N, and the Somun Cura Massif to the S (Fig. 6). For a short time it was partially linked with the San Jorge Basin. This basin represents an embayment of the Neocomian sea in a retro-arc position, associated with the paleogeographic evolution of the arc system and partially connected with the Pacific Sea (Aguirre-Urreta and Ramos, 1981). This basin is controlled by a series of W-NW half-grabens, such as the Rio Mayo half-graben, only partially inverted during the compressive phase (Peroni *et al.*, 1995). The structural trend is perpendicular to the N-NW trending half-graben system of the Bernárdides (Figari *et al.*, 1999).

The basement consists of the Middle to Late Jurassic Lonco Trapial volcanics, that have been reactivated by the Tithonian-Neocomian rifting (Clavijo, 1986). The fill of the basin consists of conglomerate packages, sandstone and shale of the Tres Lagunas Formation, which is overlain by black shale beds of the Katterfeld Formation. These marine units bear an ammonite fauna indicative of a Valanginian to Hauterivian age (Riccardi, 1970; Aguirre-Urreta and Rawson, 1999). At the base of this marine sequence there is a stratigraphic level that records maximum eustatic conditions, a bed that can also be recognized in the western subsurface part of the San Jorge Basin. The Neocomian deposits have a maximum thickness in the western sector, pinching out to the E in the Bernárdides Fold Belt, and have been interpreted as a late syn-rift by Fitzgerald *et al.* (1990). The fill ends with the fluvial and fan delta sandstone of the Apeleg Formation, of Hauterivian to Barremian age, and the continental deposits of the Chubut Group (Hechem *et al.*, 1993). A thick sequence of Cenozoic sediments covers most of the embayment (Figari *et al.*, 1999).

The Austral Basin

The Austral Basin is a typical retro-arc basin controlled by thermal subsidence until the latest Cretaceous when tectonic flexural subsidence, associated with the initial deformation in the Patagonian Andes gave place to a foreland basin (Biddle *et al.*, 1986).

The Meso-Cenozoic deposits of the Austral Basin (Fig. 6) cover the southern part of Patagonia. The basement of the basin consists of highly deformed sedimentary and metasedimentary deposits of Late Paleozoic age. These rocks also constitute the basement of the Patagonian Andes at these latitudes. A strong angular unconformity, produced during the Gondwanan Orogeny (Du Toit, 1927), separates these rocks from the volcanic, pyroclastic, and volcaniclastic deposits of Middle-Late Jurassic age. The volcanic rocks of El Quemado Formation constitute the easternmost outcrops of the Jurassic arc exposed along the axis of the Patagonian Andes. Thick Meso-Cenozoic sedimentary sequences overlie the previous rocks and constitute the fill of the Austral Basin (Biddle *et al.*, 1986). This basin opens towards the S, where the most complete Late Jurassic and Cretaceous sections were deposited. The Neocomian deposits exposed along the foothills of the Patagonian Andes are mostly marine clastics.

The base of the sedimentary sequence is composed of coarse quartzitic sandstone units ranging from fluvial continental to littoral marine facies of Tithonian to Hauterivian age. These deposits, known as the Springhill Formation, constitute an extensive clastic platform (Robles, 1982). A thick series of black shale beds of basinal facies (Río Mayer Formation) bear fossiliferous calcareous nodules. A prograding sequence of green sandstone and shale beds (Río Belgrano Formation, Barremian), bearing zones with abundant fossils preserved in sandy concretions are covering the underlying shale. Towards the S, the Neocomian deposits are represented by deeper water facies, and thus the fauna is scarce and not well preserved. Ammonites from the Hauterivian - Barremian are widely distributed through the northern part of the basin while the Berriasian and Valanginian faunas are only locally found. They were grouped in six assemblages (Riccardi, 1984, 1988) that were recently updated with new studies (Aguirre-Urreta, 1993).

Towards the S, in the region of lakes Viedma and Argentino, turbidite deposits of Late Cretaceous age are represented by the Cerro Toro Formation (Katz, 1963). Ammonites from Cenomanian to Campanian age have been found in these beds (Riccardi, 1988). These basinal facies are covered by a series of sandstone and shale beds of Maastrichtian to Cenozoic age.

The Patagonian mesetas and adjacent plains

One of the typical geological features characteristic of the extra-Andean Patagonia is the widely exposed basaltic plateaux known as the *mesetas* (Nágera, 1939). The *mesetas* are exposed N and S of the Deseado Massif, from the Atlantic coast up to the foothills of the Patagonian Andes (Ramos *et al.*, 1982). The substratum of these features is composed of Paleogene and Neogene continental deposits bearing a rich and peculiar fauna of giant mammals. These units are interbedded with ash-fall tuffs, that have been extensively used to date the mammal age sequences (Marshall *et al.*, 1983).

The plateau basalts S of 46°S form extensive plateaux such as the Meseta Buenos Aires, Belgrano, Central, La Muerte, Strobel, Las Vizcachas, among others. There are alkaline plateau basalts (Ramos *et al.*, 1982), and more alkaline post-plateau basalts of Neogene age. This retroarc basaltic activity is related to the development of asthenosphere windows associated with the subduction of seismic ridges along the Chile trench, S of the triple junction (Ramos and Kay, 1992).

The *mesetas*, farther S and in the eastern region, are capped by the Patagonian gravels first described by Darwin (1846). These gravels are interpreted as glacial and fluvioglacial deposits of late Neogene and Quaternary age. These deposits are dissected by wide valleys with younger glacial and fluvial deposits. Tertiary molasse sediments overlie some basaltic rocks in the eastern foothills, and are in turn covered by younger basalt flows. Basalt flows are widely distributed in the extra-Andean region, and are covered by glacial and alluvial deposits of Late Tertiary and Pleistocene age. The Pali Aike Volcanic Field exposed at the southern extremity of Patagonia consists of tens of alkaline basaltic monogenic cones of Plio-Pleistocene age (Skewes and Stern, 1979). These are related to neotectonic reactivation of Mesozoic extensional rift faults (Corbella *et al.*, 1996).

Geological history

The Patagonian basement records a Precambrian history that started in the Middle Proterozoic, as detected in detrital zircon fractions from the Deseado Massif (Loske *et al.*, 1999), as well as from the western part of the Somun Cura Massif. The northern sector of this massif has evidence for a Brasiliano deformation, as dated by Linares *et al.* (1990).

Reconstruction of crustal growth of Patagonia began in the Early Paleozoic, when a magmatic arc was developed in the northern margin of the Somun Cura Massif. Age constraints for this magmatic belt are erratic. Middle to Late Ordovician Rb/Sr ages should be confirmed in order to reconstruct a period of reverse polarity of subduction in the southern margin of Gondwana. At this time the Deseado Massif was also the locus of Early Paleozoic magmatism. It has been proposed that both basement blocks were independent terranes amalgamated during the Early Paleozoic to form the Austral continent (Palma, 1989).

Paleomagnetic data are still insufficient, as the Late Paleozoic paleopoles indicate that Patagonia was already part of Gondwana, but there are not good paleopoles in Early Paleozoic rocks (Rapalini and Vilas, 1991). Potential sutures, one N of Somun Cura, and another between the two massives, indicate important Permian rifting. The Permian NW trending rifts are developed in the hanging-wall of the potential suture as described by Melchor (1995) in the Sierra de Carapacha located in the vicinity of the Colorado River (39 °S). The offshore half-graben can be correlated with the onshore La Golondrina rift basin.

The Early Jurassic rift, associated with an intra-arc basin, has a different trend. Most of the first order structures have a N-NW trend as shown for the Patagonides Fold Belt (Barcat *et al.*, 1984). Extension shifted toward the margin where the Cañadón Asfalto Rift, in the middle of the extra-Andean Patagonia, has an unequivocal NW trend (Figari *et al.*, 1996). The Neocomian trend is shifted again to a W-NW trend, almost perpendicular to the Early Jurassic structures (Uliana *et al.*, 1989). After the Neocomian extension, the entire system went into a generalized thermal sag subsidence, as depicted by the large and extensive cover of the Chubut Group.

The first evidence of compression along the Patagonian Andes is registered in the Middle to Late Cretaceous. An angular unconformity between 77 Ma old basalt and Neocomian sediments E of Lake Fontana, marks the approximate orogenic front at that time. The nature of this deformation is still unknown and could reflect either an acceleration of the convergence rate, or a change of the convergence vector close to 80 Ma, or a collision of an oceanic ridge migrating from S to N as detected by Ramos *et al.* (1994). This last proposal is based on the finding of an adakite of 84 - 77 Ma E of Lago San Martín, associated with an important volcanic gap in the Southern Patagonian Andes.

Paleogene deformation is present S of 43°30'S, where the active Paleogene magmatic arc ends. Extensive alkaline basalt flows of Paleocene to Eocene age, together with essexite and other alkaline dyke swarms, widespread in the extra-Andean Patagonia, are associated with Paleogene ridge collision of the Posadas Basalt (Ramos and Kay, 1992). Neogene Andean deformation has two different regimes N and S of the Aysen Triple Junction (46°30'S). The northern sector has a mild deformation that partially inverted the properly oriented extensional structures. As a consequence of that the Bernárdides Fold Belt, almost perpendicular to the main stress, had the largest inversion. The remaining structures had a mild inversion, combined with strike-slip displacements. To the S of 46°30'S a successive sequence of segments of the ocean ridge has collided against the trench and has fostered the uplift of the Southern Patagonian Andes, deformed the molasse sediments in the extra-Andean Patagonia, and controlled the extensive basaltic plateaux (Ramos, 1989). Present neotectonic activity is restricted to S of Aysén Triple Junction, being most of Patagonia stable and undeformed.

REFERENCES

Aguirre-Urreta, M.B. (1993). Neocomian ammonite biostratigraphy of the Andean basins of Argentina and Chile. *Revista Española de Paleontología*, 8(1), 57-74.

Aguirre Urreta, M.B. and V.A. Ramos (1981). Estratigrafía y Paleontología de la Alta Cuenca del río Roble, Cordillera Patagónica. *VIII° Congreso Geológico Argentino (San Luis), Actas*, III, 101-138.

Aguirre-Urreta, M.B. and Rawson, P.F. (1999). The Lower Cretaceous Valanginian Ammonite *Chacantuceras* (gen. nov.), a link between the Neuquén and Austral basins. *Asociación Geológica Argentina, Revista*, 53(3)(1998), 354-364.

Alric, V.I., Haller, M.J. Féraud, G. Bertrand, H. and Zubia, M. (1996). Geocronología 40Ar/39Ar del volcanismo jurásico de la Patagonia Extrandina. *XIII° Congreso Geológico Argentino y III° Congreso Exploración de Hidrocarburos (Buenos Aires), Actas*, V, 243-250.

Andreis, R.R., Bensel, C.A. and Rial, G. (1989). La transgresión marina del Cretácico tardío en el borde SE de la Mesta de Somuncurá, Río Negro, Patagonia septentrional, Argentina. *Simposio sobre el Cretácico de América Latina, Actas*, Buenos Aires, A, 165-194.

Archangelsky, S. (1959). Estudio geológico y paleontológico del Bajo de La Leona (Santa Cruz). *Acta Geológica Lilloana*, Tucumán, 2, 5-133.

Ardolino, A. (1981). El vulcanismo cenozoico del borde suroriental de la meseta de Somun Cura. *VIII° Congreso Geológico Argentino (San Luis), Actas, III, 7-23.*

Arrondo, O.G. (1972). Estudio geológico y paleontológico en la zona de la Estancia La Juanita y alrededores. Museo de La Plata, Revista (N.S.) *Paleontología*, La Plata, 7(43), 1-194.

Artabe, A.E. (1985). Estudio sistemático de la tafoflora triásica de Los Menucos, Provincia de Río Negro, Argentina. Parte I. Sphenophyta, Filicophyta, Pteridospermophyta. *Ameghiniana*, 22(1-2), 3-22.

Artabe, A.E. (1986). Estudio sistemático de la tafoflora triásica de Los Menucos, Provincia de Río Negro, Argentina. Parte II. Cycadophyta, Ginkgophyta y Coniferophyta. *Ameghiniana*, 22(3-4)(1985), 159-180.

Barcat, C., Cortiñas, J.S., Nevistic, V.A., Stach, N.H. and Zucchi, H.E. (1984). Geología de la región comprendida entre los Lagos Musters - Colhué Huapí y la Sierra Cuadrada, Departamento Sarmiento y Paso de Indios, provincia del Chubut. *IX° Congreso Geológico Argentino (S.C. Bariloche), Actas*, II, 263-282.

Biddle, K.T., Uliana, M.A., Mitchum, Jr., R.M., Fitzgerald, M.G. and Wright, R.C. (1986). The stratigraphic and structural evolution of the central eastern Magallanes basin, southern South America. In: Foreland Basins, eds. Allen, P. and Homewood, P. *International Association of Sedimentologists, Special Publication*, 8, 41-61.

Bonaparte, J.F and de Gasparini, Z.B. (1978). Los saurópodos de los Grupos Neuquén y Chubut y sus relaciones cronológicas. *VII° Congreso Geológico Argentino, Actas*, Buenos Aires, II, 393-406.

Bonaparte, J., and Vince, M. (1979). El hallazgo del primer nido de dinosaurios Triásicos (Saurischia, Prosauropoda), Triásico superior de Patagonia, Argentina. *Ameghiniana*, 16(1-2), 173-182.

Ciciarelli, M.I. (1989). *Análisis estructural del sector oriental del Macizo Nordpatagónico y su significado metalogénico.* Universidad Nacional de la Plata, Ph.D. Thesis, La Plata, 179 p. (unpublished).

Clavijo, R. (1986). Estratigrafía del Cretácico inferor en el sector occidental de la Cuenca del Golfo San Jorge. *Boletín de Informaciones Petroleras (Tercera Epoca)*, 9, 15-32.

Coira, B.L., Nullo, F.E., Proserpio, C.A. and Ramos, V.A. (1975). Tectónica de basamento en la región occidental del Macizo Nordpatagónico (prov. de Río Negro y Chubut). *Asociación Geológica Argentina, Revista*, 30(4), 361-383.

Corbella, H. (1984). El vulcanismo de la altiplanicie del Somuncura. In: Geología y Recursos Naturales de la provincia de Río Negro, ed. Ramos, V.A. *IX° Congreso Geológico Argentino*, Buenos Aires, *Relatorio*, I(10), 267-300.

Corbella, H., Chelotti, L. and Pomposiello, C. (1996). Neotectónica del rift jurásico austral en Pali Aike, Patagonia extraandina, Santa Cruz, Argentina. *XIII° Congreso Geológico Argentino y III°Congreso Exploración de Hidrocarburos, Actas*, Buenos Aires, **II**, 383-393.

Cortés, J.M., Caminos, R. and Leanza, H.A. (1984). La cobertura sedimentaria eopaleozoica. In: Geología y Recursos naturales de la provincia de Río Negro, ed. Ramos, V.A. *IX° Congreso Geológico Argentino, Relatorio*, Buenos Aires, **I**, 65-84.

Cortiñas, J.S. (1984). Estratigrafía y facies del Jurásico entre Nueva Lubecka, Ferrarotti y Cerro Colorado. Su relación con los depósitos coetáneos del Chubut Central. *IX° Congreso Geológico Argentino (S.C. Bariloche), Actas*, II, 283-299.

Dalla Salda, L., Dalziel, I.W.D., Cingolani, C.A. and Varela, R. (1992b). Did the Taconic Appalachians continue into southern South America? *Geology*, **20**, 1059-1062.

Dalla Salda, Varela, L., Cingolani, C.A. and Aragón, E. (1994). The Rio Chico Paleozoic crystalline complex and the evolution of Northern Patagonia. *Journal South American Earth Sciences*, 7(3-4), 377-386.

Darwin, Ch. (1846). *Geological observations of South America, being the third part of the Geology of the voyage of the Beagle during 1832-1836*. Smith, Elder, Londres, vii + 279 p.

de Giusto, J.M., di Persia, C.A. and Pezzi, E. (1980). Nesocratón del Deseado. In: Segundo Simposio de Geología Regional Argentina, ed. Turner, J.C.M. *Academia Nacional de Ciencias*, Córdoba, **2**, 1389-1430.

De Wit, M.J. (1977). The evolution of the Scotia Arc as a key to the reconstruction of southwestern Gondwanaland. *Tectonophysics*, 37, 53-81.

Di Persia, C.A. (1962). Acerca del descubrimiento del Precámbrico en la Patagonia Extraandina, prov. de Santa Cruz. *1° Jornadas Geológicas Argentinas (San Juan), Actas*, **2**, 65-68.

Du Toit, A.L. (1927). The fossil flora of the Upper Karoo beds. *Annals of the South African Museum*, 22(2), 289-420.

Ferello, R. (1969). Intento de sistematización geocronológica de las rocas eruptivas básicas en sectores del Chubut y Santa Cruz Norte. *IV°Jornadas Geológicas Argentinas (Mendoza), Actas*, I, 293-310.

Feruglio, E. (1946). Sistemas orográficos de la Argentina. Geografía de la República Argentina. *Sociedad Argentina de Estudios Geográficos, Buenos Aires, GAEA*, **IV**, 220-225.

Figari, E.G. and Courtade, S.F. (1993). Evolución tectosedimentaria de la cuenca de Cañadón Asfalto, Chubut, Argentina. *XII°Congreso Geológico Argentino y II°Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas*, **I**, 66-77.

Figari, E.G., Courtade, S.F. and Constantini, L.A. (1996). Stratigraphy and tectonics of Cañadón Asfalto basin, lows of Gastre and Gan Gan, north of Chubut province, Argentina. In: Advances in Jurassic Research, ed. Riccardi, A.C. *GeoResearch Forum*, Switzerland, **1-2**, 359-368.

Figari, E.G., Courtade, S.F. Calegari, R. Arroyo, H. and Constantini, L.A. (1998). Estructura y estratigrafía del Cerro Ballena, faja plegada meridional de la cuenca del Golfo San Jorge. *X° Congreso Latinoamericano de Geología (Buenos Aires), Actas*, **1**, 18-23.

Figari, E.G., Strelkov, E., Laffitte, G., Cid de la Paz, M.S., Courtade, S.F., Celaya, J., Vottero, A., Lafourcade, P., Martínez, R. and Villar, H.J. (1999). Los sistemas petroleros de la cuenca del Golfo San Jorge: síntesis estructural, estratigráfica y geoquímica. *IV° Congreso geológico de Exploración y Desarrollo de Hidrocarburos (Mar del Plata), Actas*, **I**, 197-237.

Fitzgerald, M.G., Mitchum, R.M., Uliana, M.A. and Biddle, K.T. (1990). Evolution of the San Jorge Basin, Argentina. *American Association of Petroleum Geologists, Bulletin*, **74**(6), 879-920.

Franchi, M. and Page, R.N. (1980). Los basaltos cretácicos y la evolución magmática del Chubut Occidental. *Asociación Geológica Argentina, Revista*, 35(2), 208-229.

Frengüelli, J. (1946). Las grandes unidades físicas del territorio argentino. Geografía de la República Argentina. *Sociedad Argentina de Estudios Geográficos*, Buenos Aires, *GAEA*, III, 1-114.

Genini, A. (1989). Cerro Vanguardia - Provincia de Santa Cruz: nuevo prospecto auroargentífero. *III° Congreso Nacional de Geología Económica (Olavarría), Actas*, III(A), 97-110.

Groeber, P. (1938). *Mineralogía y Geología*. Espasa-Calpe Argentina, 1-492, Buenos Aires.

Hechem, J.J., Figari, E. and Homovc, J. (1993). Secuencias deposicionales en el Neocomiano del Lago Fontana, Chubut, Argentina. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas*, II, 119-123.

Herbst, R. (1988). La flora triásica del Grupo El Tranquilo, provincia de Santa Cruz, Patagonia. Parte II. Filicopsida. *Ameghiniana*, **25(4)**, 365-379.

Hillebrandt, A. von (1987). Liassic ammonite zones of South America and correlations with other provinces. With description of new genera and species of ammonites. In: *Bioestratigrafía de los Sistemas regionales del Jurásico y Cretácico de América del Sur I*, ed. W. Volkheimer, pp. 111-157. Mendoza.

Homovc, J.F., Conforto, G.A. and Lafourcade, P.A. (1993). Faja plegada en el subsuelo de la cuenca del Golfo San Jorge: ejemplo de inversión tectónica. *XII°Congreso Geológico Argentino y II°Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas*, III, 233-240.

Jalfin, G.A. (1987). *Estratigrafía y paleogeografía de las Formaciones la Golondrina y la Juanita, Pérmico de la provincia de Santa Cruz y su relación con rocas de edad similar en las Islas Malvinas*. Universidad Nacional de Tucumán. Ph.D. Thesis, S.M. de Tucumán, 219 p. (unpublished)

Jalfin, G.A. and Herbst, R. (1995). La flora triásica del Grupo El Tranquilo, provincia de Santa Cruz, Patagonia. Estratigrafía. *Ameghiniana*, **32**, 211-229, Buenos Aires.

Katz, H.R. (1963). Revision of Cretaceous Stratigraphy in Patagonian Cordillera of Ultima Esperanza, Magallanes Province, Chile. *American Association of Petroleum Geologists, Bulletin*, 47, 506-524.

Kay, S.M., Ramos, V.A., Mpodozis, C. and Sruoga, P. (1989). Late Paleozoic to Jurassic silicic magmatism at the Gondwanaland margin: analogy to the Middle Proterozoic in North America? *Geology*, **17**(4), 324-328.

Kay, S.M., Ardolino, A., Franchi, M. and Ramos, V. (1993). Origen de la Meseta de Somún Curá: distribución y geoquímica de sus rocas volcánicas máficas. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas*, IV, 236-248.

Keidel, J. (1921). Sobre la distribución de los depósitos glaciares del Pérmico conocidos en la Argentina y su significación para la estratigrafía de la serie del Gondwana y la paleogeografía del Hemisferio Austral. *Academia Nacional de Ciencias, Córdoba, Boletín*, 25, 239-368.

Leanza, A.F. (1958). Geología Regional. In: *La Argentina, Suma de Geografía*, Tomo I, Capítulo III, pp. 217-349, Editorial Peuser, Buenos Aires.

Legarreta, L. and Uliana, M.A. (1994). Asociaciones de fósiles y hiatos en el Supracretácico-Neógeno de Patagonia; una perspectiva estratigráfico-secuencial. *Ameghiniana*, 31(3), 257-281.

Lesta, P.J. (1968). Estratigrafía de la Cuenca del Golfo San Jorge. *III° Jornadas Geológicas Argentinas (Comodoro Rivadavia), Actas*, I, 251-289.

Lesta, P.R., Ferello, R. and Chebli, G. (1980). Chubut Extraandino. In: *Segundo Simposio Geología Regional Argentina, Córdoba, Academia Nacional de Ciencias*, II, 1306-1387.

Linares, E., Ostera, H.A. and Parica, C.A. (1990). Edades radimétricas preliminares del basamento cristalino de las vecindades de Mina Gonzalito y de Valcheta, provincia de Río Negro, República Argentina. *XI° Congreso Geológico Argentino (San Juan), Actas*, II, 251-253.

Loske, W., Márquez, M., Giacosa, R., Pezzuchi, H. and Fernández, M.I. (1999). U/Pb geochronology of pre-Permian basement rocks in the Macizo del Deseado, Santa Cruz province, Argentine Patagonia. *XV° Congreso Geológico Argentino (Salta), Actas* (in press).

Manceñido, M. and Damborenea, S. (1984). Megafaunas de invertebrados paleozoicos y mesozoicos. In: Geología y Recursos Naturales de la provincia de Río Negro, ed. Ramos, V.A. *IX° Congreso Geológico Argentino (S.C. Bariloche), Relatorio*, 413-463.

Márquez, M. and Giacosa, R. (1999). Deformaciones dúctiles paleozoicas y corrimientos terciarios en el Neopaleozoico del Chubut: consecuencias estratigráficas y tectónicas. *Asociación Geológica Argentina, Revista* (in press).

Márquez, M., Giacosa, R., Pezzuchi, H. and Fernández, M. (1994). El basamento pre-Pérmico del noreste del Macizo del Deseado, Argentina. *Zentralblatt für Geologie und Paläontologie, Teil* I, 1993(1-2), 101-114.

Marshall, L.G., Hoffstetter, R. and Pascual, R. (1983). Geochronology of the continental mammal-bearing Tertiary of South America. *Paleovertebrate, Memoir extra*, 1-93.

Melchor, R.N. (1995). *Sedimentología de las unidades paleozoicas aflorantes del centro-oeste de la provincia de La Pampa, Argentina.* Universidad Nacional de La Plata, Facultad de Ciencias Naturales y Museo, Ph.D. Thesis, La Plata, 272 p. (unpublished),

Nágera, J.J. (1939). Geografía Física de la República Argentina. In O. Manito and J.J. Nágera, Geografía Física de las Américas y de la República Argentina. Editorial Kapeluz, Buenos Aires, 232 p.

Nullo, F.E. (1983). Descripción geológica de la Hoja 45c Pampa de Agnia, provincia del Chubut. *Servicio Geológico Nacional, Buenos Aires, Boletín*, **199**, 1-94.

Palma, M.A. (1989). Los eventos geológicos del Macizo del Deseado durante la evolución tectónica del continente Austral. *Reunión sobre Geotransectas de América del Sur, Mar del Plata*, Montevideo, 1-3 Junio 1989, 97-101.

Pankhurst, R.J. and Rapela, C.W. (1993). The Jurassic acidic volcanism of Northeast Patagonia: a short lived event of deep origin. *II° International Symposium on Andean Geodynamics*, Oxford, *Proceedings*, 423-425.

Panza, J.L., Cobos, J. and Ragona, D. (1994). *Mapa Geológico de la provincia de Santa Cruz, República Argentina.* Dirección Nacional del Servicio Geológico, escala 1: 750.000, Buenos Aires.

Peroni, G.O., Hegedus, A.G., Cerdan, J., Legarreta, L., Uliana, M.A. and Lafitte, G. (1995). Hydrocarbon accumulation in an inverted segment of the Andean foreland: San Bernardo belt, Central Patagonia. In: Petroleum Basins of South America, eds. Tankard, A.J. Suárez S., R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 403-419.

Pezzuchi, H.D. (1978). *Estudio geológico de la zona de la Estancia Dos Hermanos, Estancia 25 de Mayo y adyacencias, departamento Deseado, provincia de Santa Cruz.* Universidad Nacional de La Plata, Ph.D. Thesis, La Plata (unpublished).

Ramos, V.A. (1983). Evolución tectónica y metalogénesis de la Cordillera Patagónica. *II° Congreso Nacional Geología Económica (San Juan), Actas*, **1**, 108-124.

Ramos, V.A. (1989). Foothills structure in Northern Magallanes Basin, Argentina, *American Association Petroleum Geologists, Bulletin*, 73(7), 887-903.

Ramos, V.A. (1996). Evolución tectónica de la Plataforma Continental. In: *Geología y Recursos Naturales de la Plataforma Continental Argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 385-404. Asociación Geológica Argentina e Instituto Argentino del Petróleo, Buenos Aires.

Ramos, V.A. (1999). Las provincias geológicas del territorio argentino. In: Geología Argentina, ed. Caminos, R. Instituto de Geología y Recursos Minerales, Buenos Aires, *Anales*, **26**(3) (in press).

Ramos, V.A. and Drake, R. (1987). Edad y significado tectónico de la Formación Río Tarde (Cretácico) Lago Posadas, provincia de San Cruz. *X° Congreso Geológico Argentino (Tucumán), Actas*, I, 143-148.

Ramos, V.A. and Kay, S.M. (1992). The Southern Patagonian plateau basalts: retroarc testimony of a ridge collision, Argentina. Andean geodynamics Symposium, Grenoble, France, eds. Oliver, R.A., Vatin-Perignon, N. and Laubacher, G. *Tectonophysics*, **205**, 261-282.

Ramos, V.A., Niemeyer, H., Skarmeta, J. and Muñoz, J. (1982). The magmatic evolution of the Austral Patagonian Andes. *Earth Science Reviews*, **18**(3-4), 411-443.

Ramos, V.A., Mahlburg Kay, S. and Sacomani, L. (1994). La dacita Puesto Nuevo y otras rocas magmáticas: colisión de una dorsal oceánica cretácica. *VII° Congreso Geológico Chileno (Concepción), Actas*, II, 1172-1176.

Rapalini, A.E. and Vilas, J.F. (1991). Preliminary paleomagnetic data from the Sierra Grande Formation: Tectonic consequences of the first mid-Paleozoic paleopoles from Patagonia. *Journal of South American Earth Sciences*, **4**, 25-41.

Rapela, C. and Caminos, R. (1987). Geochemical characteristics of the Upper Paleozoic magmatism in the eastern sector of Northpatagonian massif. *Revista Brasileira de Geociencias*, São Paulo, **17**(4), 535-543.

Rapela, C.W. and Pankhurst, R. (1993). El volcanismo riolítico del noroeste de la Patagonia: un evento mesojurásico de corta duración y origen profundo. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos (Mendoza), Actas*, IV, 179-188.

Rapela, C.W., Dias, C.F., Francese, J.R., Alonso, G. and Benvenuto, A.R. (1991). El batolito de la Patagonia central: evidencias de un magmatismo triásico-jurásico asociado a fallas transcurrentes. *Revista Geológica de Chile*, **18**(2), 121-138.

Riccardi, A.C. (1970). *Favrella* R. Douville, 1909 (Ammonitina Cretácico Inferior): Edad y distribución. *Ameghiniana*, **7**, 110-138.

Riccardi, A.C. (1983). The Jurassic of Argentina and Chile. In: *The Phanerozoic geology of the world II, The Mesozoic*, eds. Moullade, B.M. and Nairn, A.E.M., pp. 201-263. Elsevier, 8, Amsterdam.

Riccardi, A.C. (1984). Las asociaciones de amonitas del Jurásico y Cretácico de la Agentina. *IX°Congreso Geológico Argentino (S.C. Bariloche), Actas*, IV, 559-595.

Riccardi, A.C. (1988). The Cretaceous system of southern South America. *Geological Society of America, Memoir*, **168**, 1-161.

Robles, D. (1982). El desarrollo de la Formación Springhill en la cuenca de Magallanes. *I° Congreso Nacional de Hidrocarburos (Buenos Aires), Exploración*, 293-312.

Salani, F.M. and Page, R.F.N. (1990). El Complejo Volcánico Pire Mahuida, provincia del Chubut. *Asociación Geológica Argentina, Revista*, **44**(1-4), 364-380.

Schalamuk, I.B., Fernández, R. and Etcheverry, R.O. (1995). Gold-silver epithermal veins in the Macizo del Deseado, Argentina. In: *Mineral deposits; from their origin to their environmental impacts*, eds. Pasava, J., Kribek, B. and Zak, K., pp. 385-388. Balkema, Rotterdam.

Sciutto, J.C. (1981). Geología del codo del río Senguerr, Chubut, Argentina. *VIII° Congreso Geológico Argentino (San Luis), Actas*, III, 203-219.

Skewes, M.A. and Stern, C.R. (1979). Petrology and geochemistry of alkali basalts and ultramafic inclusions from the Pali-Aike volcanic field in southern Chile and the origin of the Patagonian plateau lavas. *Journal of Volcanology and Geothermal Research*, **6**, 3-25.

Spalletti, L.A. (1990). Los ambientes sedimentarios del límite Cretácico-Terciario en el sector occidental del Macizo Norpatagónico. *Asociación Argentina de Mineralogía, Petrología y Sedimentología, Revista*, **19**(1-4), 49-56.

Spalletti, L.A., A. del Valle, A., Manassero, M.J. and Matheos, S.D. (1989). Procedencia y ambiente tectónico de las areniscas cretácico-terciarias del sector norte de la Patagonia Argentina. *Simposios sobre el*

Cretácico de América Latina (Buenos Aires), Actas, A, 149-164.

Stipanicic, P.N. and Reig, O. (1956). Breve noticia sobre el hallazgo de anuros en el denominado "Complejo Porfírico" de la Patagonia Extraandina, con consideraciones acerca de la composición geológica del mismo. *Asociación Geológica Argentina, Revista*, **10**(4), 215-233.

Stipanicic, P.N., Rodrigo, F., Baulíes, O.L. and Martínez, C.G. (1968). Las formaciones presenonianas en el denominado Macizo Nordpatagónico. *Asociación Geológica Argentina, Revista*, **23**(2), 67-98.

Stipanicic, P.N., Toubes, R.O., Spickermann, J.P. and Halpern, M. (1972). Sobre la composición y edad de algunas plutonitas del nordeste de la provincia de Santa Cruz, Patagonia (Rep. Argentina). *Asociación Geológica Argentina, Revista*, **26**(4), 459-467.

Suero, T. (1962). Paleogeografía del Paleozoico superior de la Patagonia (República Argentina). *Asociación Geológica Argentina, Revista*, **16**, 35-42.

Ugarte, F.R. (1966). La cuenca Carbonífera - Jurásica de la Patagonia Meridional. Universidad de la Patagonia, San Juan Bosco, 2. *Ciencias Geológicas* **I-1**, 37-68.

Uliana, M.A., Biddle, K.T. and Cerdan, J. (1989). Mesozoic extension and the formation of Argentine sedimentary basins. In: Extensional tectonics and stratigraphy of the North Atlantic margins, eds. Tankard, A.J. and Balkwill, H.R. *American Association Petroloeum Geologists, Memoir*, **46**, 599-614.

Varela R., Cingolani, C., Sato, A., Dalla Salda, L., Brito Neves, B.B., Basei, M.A.S. Siga Junior, O. and Texeira, W. (1997). Proterozoic and Paleozoic evolution of Atlantic area of North-Patagonian massif, Argentine. *South America Symposium on Isotope Geology (Sao Paulo), Extended Abstracts*, 326-329.

Windhausen, A. (1931). *Geología Argentina*. In: *Geología Histórica y Regional del Territorio Argentino*, ed. Peuser, J., Tomo II, 1-645, Buenos Aires.

Zavattieri, A.M. (1993). Palinología de la Formación El Tranquilo (Triásico), provincia de Santa Cruz, Argentina. *Ameghiniana*, **29**(4), 305-314.

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THE SIERRAS PAMPEANAS OF ARGENTINA:

PALEOZOIC BUILDING OF THE SOUTHERN PROTO-ANDES

Carlos W. Rapela

Lower Paleozoic magmatism in South America is most clearly recorded in outcrops in central and northwestern Argentina, extending into southern Bolivia, northeastern Chile and the coast of southwestern Peru. Igneous and metamorphic rocks of this age are well exposed in the geological provinces of central and northwestern Argentina (Fig. 1), and its characteristics are considered representatives of the Paleozoic evolution of southwestern Gondwana. Outcrops of Lower Paleozoic rocks are much more scattered S of 33°S, where they appear as the basement rocks of the Patagonian massifs and are also recognized in drill cores as far S as the Strait of Magellan. This distribution indicates that the basement of the southwestern sector of the South American Continent was mainly developed during the Early Paleozoic, being the Pampean and the Famatinian the best recognized orogenic episodes (Pankhurst and Rapela, 1998). These large Paleozoic regions are the basement that structurally controlled the Meso-Cenozoic Andean magmatism associated with the Pacific subduction.

Within the classic geological-morphostructural provinces recognized in the Argentinean geology, there are several that include extensive sectors composed mainly of igneous and metamorphic rocks of Early and Middle Paleozoic age. The best exposed and studied are the provinces of the northwestern sector: the Sierras Pampeanas, the Famatina System, the Precordillera, the Eastern Cordillera and Puna; and in Patagonia, the North Patagonian Massif (Fig. 1). The Early Paleozoic tectonic evolution of southern South America described here is based on new geochemical, isotope and petrological data from a 500 km traverse at 31°S - 32°S across the Eastern Sierras Pampeanas, the southern extension of the Famatina System, the Western Sierras Pampeanas, and the Precordillera (Pankhurst *et al.*, 1998; Rapela *et al.*, 1998a, b, 1999). Pre-Silurian metamorphic and magmatic history in this transect (Fig. 1) has been inferred from: (1) dating by conventional U/Pb on abraded zircon, U/Pb SHRIMP analyses, and whole-rock Rb/Sr and K/Ar; (2) thermo-barometry based on microprobe mineral analyses; and (3) Nd and Sr isotopes, and major element and trace element geochemistry of the magmatic suites. Recent accounts on the Early Paleozoic history of the provinces of the northwestern Argentina, and different viewpoints on its geodynamic evolution are presented in the book The proto-Andean Margin of Gondwana, Pankhurst and Rapela (eds), 1998. Pioneering contributions and research on the geology of these provinces, back in time up to 1978, can be consulted in the Segundo Simposio de Geología Regional Argentina, volume I, Turner (ed), 1979.

Gondwana mobile belts and the Precordillera Terrane

Geological and paleontological evidence indicating that the Argentine Precordillera is a Laurentian terrane has revolutionized ideas about the proto-Andean margin of South America (Dalziel, 1997; Benedetto, 1998). Although the exotic origin of the Precordillera has been widely accepted, the timing of accretion and the associated geotectonic models remained controversial (*e.g.* Dalla Salda *et al.*, 1992, 1998; Astini *et al.*, 1995; Dalziel, 1997; Benedetto, 1998; Rapela *et al.*, 1998b). Despite the fact that several problems are still unsolved, unravelling the evolution of the Gondwana margin immediately prior to accretion of the Precordillera terrane has helped to distinguish different mobile belts and to better constrain existing geodynamic models (Rapela *et al.*, 1998b; Sims *et al.*, 1998).

Paleozoic provinces at 22° S to 33° S may be divided into those related to Early to Middle Cambrian accretion (the Pampean Mobile Belt), and those associated with Ordovician subduction (the Famatinian Mobile Belt) and later collision of an exotic terrane (the Precordillera Terrane). More precise geochronological data and detailed petrological studies lead to a more restricted definition of Pampean and Famatinian events and their duration (Rapela *et al.*, 1998b). The Pampean Mobile Belt includes the Eastern Sierras Pampeanas and Eastern Cordillera provinces (Fig. 1). From 22° S to 27° S the Pampean Belt is dominated by the Puncoviscana Formation, which is mainly composed of a thick sequence of low-grade metamorphic rocks including metapelite and turbidite beds with well preserved primary structures (Jeek, 1990), carrying trace fossils and indicating a late Precambrian (Vendian) to early Lower Cambrian (Tommotian) age (Durand, 1996). This tightly folded series is intruded by Cambrian granite and unconformably covered by Middle-to Late Cambrian platform sediments, therefore suggesting a Middle Cambrian deformation (Rossi *et al.*, 1992; Durand, 1996). From 27°S to 33°S only metamorphic equivalents of the Puncoviscana Formation are exposed. The sequences are mostly composed of medium to low pressure and temperature, high-grade metamorphic rocks and anatectic granites of Early to Middle Cambrian age, partially remobilized and intruded by Ordovician to Carboniferous granitoid plutons (Rapela *et al.*, 1998a).

The Famatinian Mobile Belt, encompassing the Puna, the Famatina System, the Sierra de los Llanos de La Rioja and Sierras de San Luis, consists of low- pressure and low-to-high temperature metamorphic rocks, back-arc sediments, and *widespread Early Ordovician magmatic rocks* (Fig. 2).

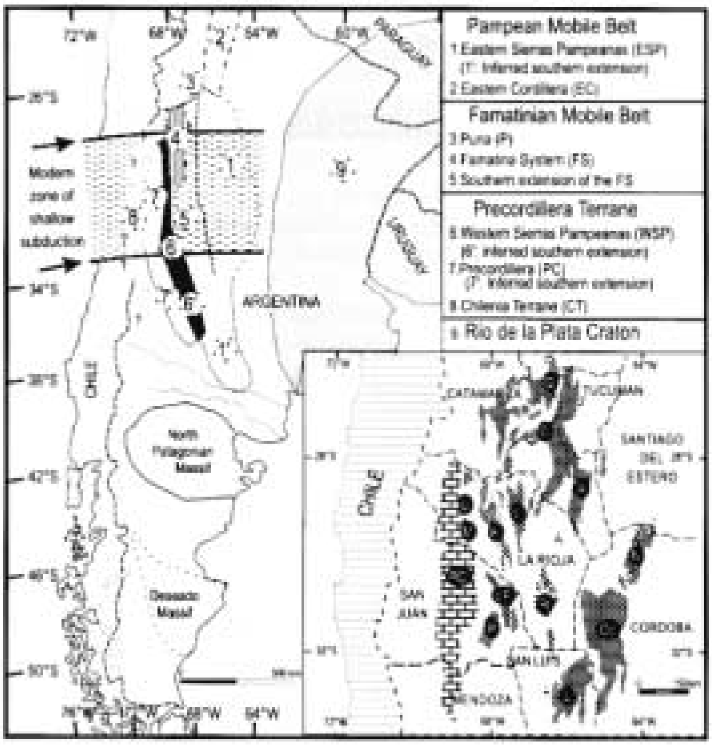
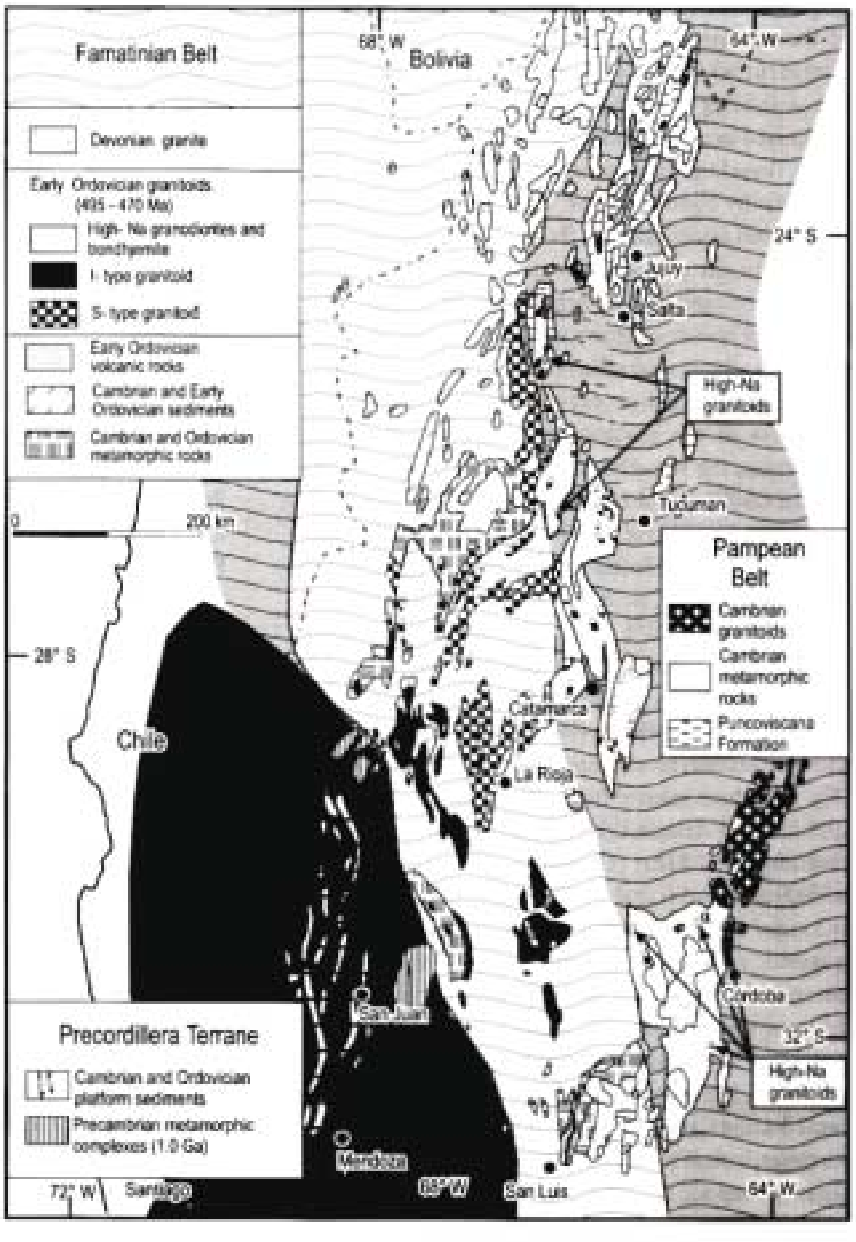


FIG. 1. Early Paleozoic geological provinces in the southern Andes, disclosed by back-thrusting above Miocene "flat-slab" subduction (modified after Rapela et al., 1998b). Principal sierras in this segment are: Q - Quilmes; C - Capillitas; A - Ancasti; F - Famatina; V - Velasco; U - Umango; M - Maz; N - Norte de Córdoba; Co - Córdoba; L - San Luis; R - Llanos de la Rioja and Chepes; VF - Valle Fértil; P - Pie de Palo; PC - Precordillera.

FIG. 2. Early Paleozoic reconstruction of the proto-Andean margin of southern South America after the docking of the Precordillera Terrane.



The Precordillera Terrane (Precordilleran Terrane of Astini *et al.*, 1995) is essentially equivalent to the Cuyania Terrane of Ramos *et al.* (1998) and consists of metamorphic basement partly covered by Cambrian-Ordovician carbonate rocks. In the Precordillera the latter contain typical Laurentian faunas (Benedetto, 1998) and subsequent siliciclastic units that extend into the Upper Paleozoic; a Grenville age for their unexposed basement has been inferred from xenoliths in Miocene volcanic rocks (1.102 ± 0.006 Ga, U/Pb zircon, Kay *et al.*, 1996). Exposed Grenville-age basement occurs both W and E of the Precordillera. Caminos (1973) was the first to recognize the contrasting lithology of the basement rocks occurring to the E of the Precordillera as the Sierra de Pie de Palo, Sierra de Maz and Sierra de Umango, namely the Western Sierras Pampeanas (Fig.1), that are characterized by an absence of major batholiths, abundant ultramafic, mafic and metamorphosed carbonate sequences, with metamorphic assemblages that generally indicate higher-pressure metamorphism than in the Eastern Sierras Pampeanas. The isotope ages indicating a Grenvillian age for the Western Sierras Pampeanas (McDonough *et al.*, 1993; Varela *et al.*, 1996; Pankhurst and Rapela, 1998) is further evidence favouring the allochtony of the Precordillera Terrane, and confirmed the early observations made by Caminos (1973, 1979) indicating sharply contrasting evolutionary lines and petrogenetic processes for the basement rocks of the Eastern and Western Sierras Pampeanas.

The Neoproterozoic-Devonian evolution of the Pampean and Famatinian belts is compared to that of the Precordillera Terrane in Figure 3. The geodynamic model described below follows the main features presented by Rapela *et al.* (1998b), but new U/Pb SHRIMP data have allowed better estimations of the intrusion age of the large peraluminous batholiths (Rapela *et al.*, 1999), and modifications in the interpreted sequence of events.

Supercontinent break-up: opening of the Puncoviscan and Southern Iapetus oceans

Break-up of the the latest Precambrian Pannotia Supercontinent produced continental terranes that were accreted to Laurentia and Gondwana in early Paleozoic collisions. Recently reviewed geological evidence along the conjugate margins indicates a latest Precambrian to Early Cambrian age for the rift-drift transition (Dalziel, 1997). The Puncoviscana Formation of western Argentina is seen as the South American proximal counterpart of deep-marine passive margin sequences overlying the edge of the Río de la Plata Craton (Figs. 1 and 3). Inferred sequential opening of the Puncoviscan and Southern Iapetus oceans (the result of a spreading-center shift?) left fossil oceanic crust between a detached continental block, the Pampean Terrane, and the Gondwana margin (Fig. 3).

Pampean Orogeny: Early Cambrian subduction and terrane collision

The proto-Andean margin of Gondwana changed from passive to active in Early Cambrian time, leading to closure of the Puncoviscan Ocean (Fig. 3). The youngest trace fossils in the Puncoviscana Formation are Tommotian (Durand, 1996), *c.* 534 - 530 Ma. These date the last passive margin deposits, succeeded by a thick accretionary prism. A subduction-related belt of metaluminous calc-alkaline granitoid bodies and dacite-rhyolite was emplaced along the eastern part of the Eastern Sierras Pampeanas (Lira *et al.*, 1996; Rapela *et al.*, 1998a). In the Sierras de Córdoba this is dated at 530 ± 4 Ma by U/Pb on abraded zircon from three plutonic units (Rapela *et al.*, 1998a). Intrusion was followed by crustal thickening and burial to granulite facies conditions (P = 8.6 ± 0.8 kbar, T = 810 ± 50 °C). A clockwise P-T-time path is inferred, producing regional migmatites during peak thermal conditions of P = 5.7 ± 0.4 kbar, T = 820 ± 25 °C; dated at 522 ± 8 Ma by a U/Pb SHRIMP age on monazite. Immediately following this, strongly peraluminous, cordierite- and sillimanite-bearing granite and associated cordieritite were generated by low-P anatexis of metasediments (P = 3.9 ± 0.6 kbar, T = 684 ± 60 °C); an 18 point Rb/Sr whole-rock isochron gave 523 ± 4 Ma, Sr_i = 0.7136, mean square of weighted deviates (MSWD) = 2.0; and U/Pb on abraded zircon gave 523 ± 2 Ma. This orogeny is interpreted as being due to Early to Middle Cambrian collision between the semi-autochthonous Pampean Terrane and Gondwana (Rapela *et al.*, 1998a,b). Middle Cambrian deformation of the Puncoviscana Formation N of 27 °S (Rossi *et al.*, 1992; Bahlburg and Hervé, 1997) is ascribed to the Pampean Orogeny. SHRIMP ages of 1.4 Ga to 600 Ma for inherited zircon from both high-grade metapelite and peraluminous granite suggest provenance from Middle to Neoproterozoic sources.

Ordovician subduction on the Gondwana Margin: the Famatinian Arc

After Pampean terrane accretion, the proto-Pacific margin became passive at this latitude until Early Ordovician time: there was no orogenic activity and little granitoid emplacement in the Sierras Pampeanas between 515 and 490 Ma. The Famatinian Magmatic Arc began to develop in earliest Ordovician time and affected all geological provinces in southwestern Gondwana. An inner arc of sparse high-Al trondhjemite bodies was emplaced at 496 ± 2 Ma in the Pampean foreland of the Sierras de Córdoba; a new cordilleran arc of calcic granitoid plutons began in the W at 492 ± 6 Ma in the Sierras de Los Llanos-Chepes and at 484 ± 5 Ma in the Famatina System (U/Pb SHRIMP and U/Pb conventional on abraded zircon, Rapela *et al.*, 1998a, 1999; Pankhurst *et al.*, 1998). New U/Pb SHRIMP data on the typical peraluminous facies of the large Capillitas and Velasco batholiths that were previously considered to be of Silurian-Devonian age, also revealed an Early Ordovician age (470 ± 3 Ma; 479 ± 3 Ma, Rapela *et al.*, 1999). The new results suggest that supracrustal reworking is a prominent feature of the Early Ordovician climax of the

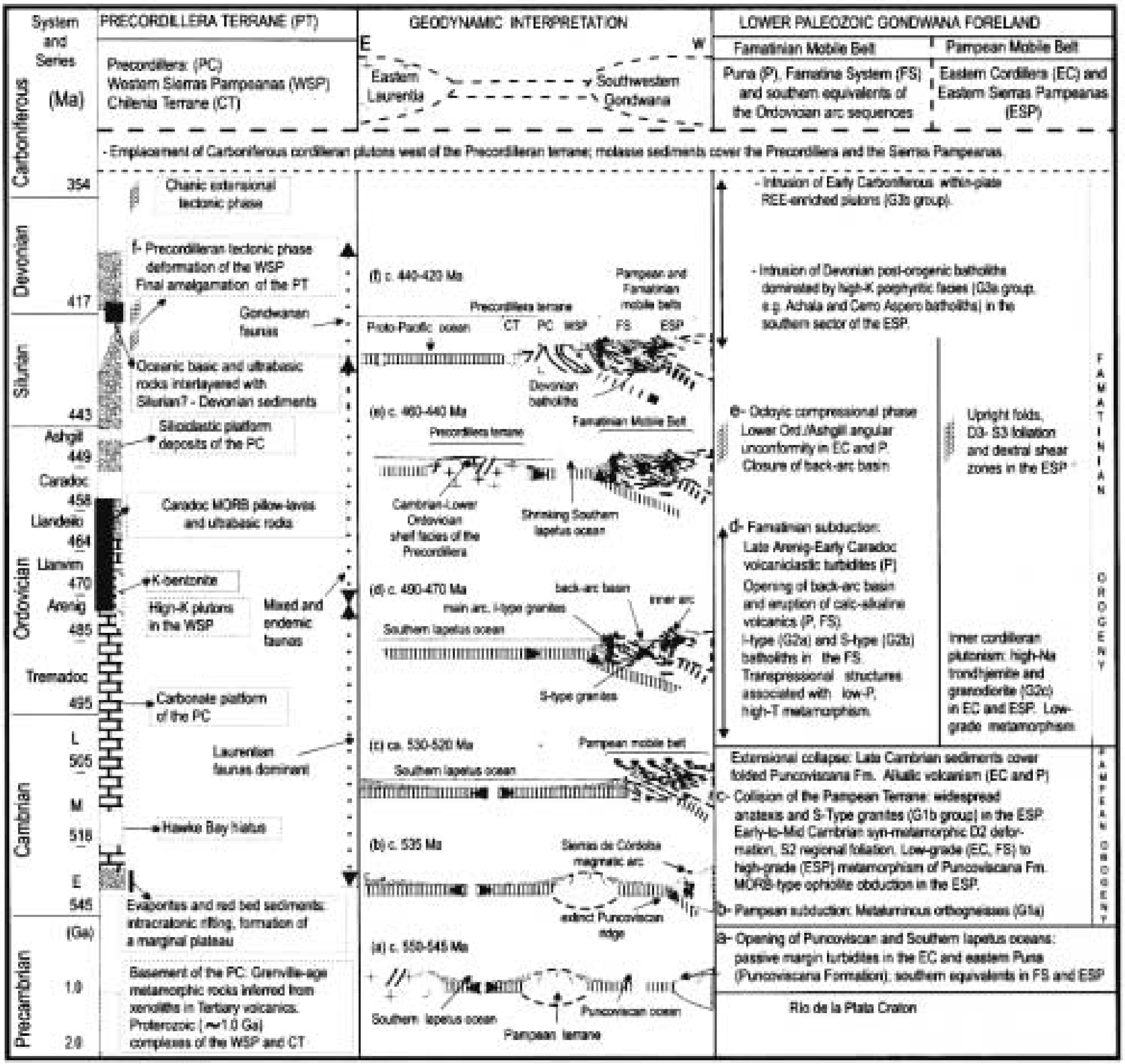


FIG. 3. Sequence and characteristics of Paleozoic orogenic events in the proto-Andean margin of South America (modified after Rapela et al., 1998b). Tectono-magmatic evolution and geochronological control for Gondwana are after Pankhurst et al. (1998) and Rapela et al. (1998a, 1999); time scale of Gradstein and Ogg (1996). Sedimentary sequences, faunas and tectonic phases in the Precordillera are simplified and slightly modified after Astini et al. (1995), Benedetto (1998), and Keller et al. (1998); plate tectonic stages of the Precordillera terrane are after Keller et al. (1998). Acronyms as in Figure 1.

Famatinian episode. Local generation of S-type granite intrusions during high-T, low-P metamorphism in the Sierra de Chepes occurred also in Early Ordovician times (479 ± 4 Ma; 483 ± 12 Ma, Rapela *et al.*, 1999).

I-and S-type Famatinian granites intruded the westernmost passive margin sequences overlying the Pampean basement. They exhibit identical Sm/Nd crustal residence ages to those of the Cambrian plutonic rocks of the Sierras de Córdoba (1.75 - 1.55 Ga), significantly older than those of the Precordillera Terrane (Pankhurst *et al.*, 1998). The Famatinian Magmatic Arc was clearly continental, not an island arc. Although roughly coeval, the Famatinian granitoid intrusives show a regular geographical distribution of compositional types (Fig. 2): (a) medium to small size bodies of high-Na tonalite, trondhjemite and granodiorite were emplaced in the easternmost side of the orogen, along the older Pampean Belt; (b) low-Ca peraluminous batholiths and small bodies of cordierite-bearing monzogranite were emplaced *in situ* at shallow depths, in high-T, low-P metamorphic rocks cropping out along the central sector of the Famatinian Belt; and (c) I-type, calcic metaluminous suites, from gabbro to high silica leucogranite, were emplaced along the western side of the Famatinian Belt. Volcanic rocks are also conspicuous to the N of this belt.

A wide back-arc basin filled by Tremadocian–Llanvirnian shallow-marine volcaniclastic and/or volcanic rocks opened between the inner and main arcs (Fig. 3). It was underlain by oceanic crust in the southern Puna (Bahlburg and Hervé, 1997), and opened progressively southwards, reaching the Sierras de San Luis at 33° S (Sims *et al.*, 1998). The basin closed during the Late Ordovician *Ocloyic tectonic phase, producing westward thrusting and* mylonite belts in the high-grade Pampean basement of the Sierras de Córdoba (D3 deformation). Pegmatites associated with the dextral shear zones of this stage yield K/Ar ages of 447 - 435 Ma (Rapela *et al.*, 1998a).

The passive margin of the Precordillera Terrane

Evolution of the Gondwana foreland at this time sharply contrasts with that of the obviously exotic Precordillera (Fig. 3). The Precordillera has a typical passive margin, Early Cambrian-Arenigian, carbonate shelf sequence, which resembles that of the southern Appalachians and similarly overlies Grenville-age metamorphic rocks (Astini *et al.*, 1995; Dalla Salda *et al.*, 1992, 1998; Kay *et al.*, 1996). The Grenville age of Western Sierras Pampeanas orthogneiss considered part of the Precordillera Terrane basement is established by Rb/Sr whole-rock isochrons of 1.03 ± 0.03 Ga, initial Sr = 0.70258 ± 0.00028 (Sierra de Umango; Varela *et al.*, 1996) and 1.021 ± 0.012 Ga, initial Sr = 0.7045 ± 0.0003 (Sierra de Pie de Palo; Pankhurst and Rapela, 1998).

Ordovician high-K quartz monzonite and monzogranite intruding the Grenville rocks of the Sierra de Pie de Palo (481 ± 6 Ma, SHRIMP U/Pb zircon, Pankhurst and Rapela, 1998) were tentatively associated with volcanism (Rapela *et al.*, 1998b) represented by K-bentonite interbedded with the Arenig–Llanvirn carbonate units of the Precordillera (Huff *et al.*, 1998). This magmatism was coeval with an extensional collapse of the Precordillera carbonate shelf, indicated by Arenigian mass-flow turbidite deposits and giant olistoliths, and culminating with eruption of Caradocian mid-ocean ridge basalt and ultramafic rocks (Fig. 3; Astini *et al.*, 1995). Ocloyic compressive tectonism at the Gondwana margin (Fig. 3; Bahlburg and Hervé, 1997) may have been related to the Ashgillian closure of the Famatina back-arc basin and not to the accretion of the Precordillera Terrane, as often supposed.

Docking of the Precordillera Terrane

Timing of the collision of the Precordillera Terrane with Gondwana is still unresolved. Despite many other differences, most recent geodynamic models have considered that this event occurred during the Middle Ordovician Famatinian Orogeny (Dalla Salda *et al.*, 1992; Astini *et al.*, 1995; Dalziel, 1997; Ramos *et al.*, 1998). *However, there is increasing evidence for later accretion*, during Silurian time (Fig. 3): (1) there is no evidence for Late Cambrian subduction on the Gondwana margin, which started *c.* 490 Ma, leaving little time to close the southern part of the Iapetus Ocean before the Middle Ordovician; (2) reactivation of the Grenville basement in the Sierra de Pie de Palo occurred between 432 and 394 Ma (Ar/Ar ages on hornblende and muscovite, Ramos *et al.*, 1998); (3) the first contraction of the Precordillera sedimentary sequences occurred in Late Silurian-Early Devonian times (Astini, 1996), which is consistent with K/Ar data obtained in the western sedimentary sequences of the Precordillera, suggesting a thermal overprint during the 425 - 400 Ma interval (Buggisch *et al.*, 1994); (4) the first unequivocal evidence for close proximity of the Precordillera to Gondwana is the Early Wenlockian (*c.* 425 Ma) *Clarkeia* fauna (Benedetto, 1998), and; (5) intrusion of high-K batholiths in the southern sector of the Eastern Sierras Pampeanas and the formation of transpressional shear zones continued well into Devonian times (Sims *et al.*, 1998).

A Silurian accretion of the Precordillera terrane overlaps with that of the postulated Chilenia terrane (Ramos and Basei, 1997), so that the small outcrops from which Chilenia has been identified might really be slivers of the Precordillera Terrane basement, displaced westwards during the collision, rather than a separate terrane (such an origin for the Frontal Cordillera basement outcrops was considered by Ramos and Basei, 1997). If the Precordillera Terrane was sutured to Gondwana in Late Ordovician to Silurian times, this cannot be the tectonic cause of the Early Ordovician Famatinian Orogeny. A more obvious tectonic setting for the generation of Famatinian magmas is the opening and closing history of the Famatina back-arc basin.

Conclusions

The crustal framework of the Southern Andes was formed during Early Paleozoic continental collisions. The first was the Early to Middle Cambrian (530 - 520 Ma) collision of the Pampean Terrane against the newly-formed passive margin of Gondwana. Subduction restarted at *c.* 490

Ma (Tremadocian?), forming a wide continental arc and an ensialic back-arc basin floored by oceanic crust. Accretion of the Precordillera Terrane to Gondwana was complex and protracted, starting with closure of the Famatinian back-arc basin during the Late Ordovician (450 - 440 Ma) and concluding with collision in Silurian times (440 - 420 Ma), followed by intrusion of large post-orogenic Devonian batholiths in the southern Sierras Pampeanas.

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REFERENCES

Aceñolaza, F.G., Miller, H. and Toselli, A.J., eds. (1990). *El Ciclo Pampeano en el Noroeste Argentino*. Universidad Nacional de Tucumán, Argentina, Serie Correlación Geológica, 4, 227 p.

Astini, R.A. (1996). Las fases diastróficas del Paleozoico Medio de la Precordillera del oeste argentino -evidencias estratigráficas. *XIII Congreso Geológico Argentino y III Congreso de Exploración de Hidrocarburos, Buenos Aires, Actas*, **5**, 509-526.

Astini, R.A., Benedetto, J.L., and Vaccari, N.E. (1995). The early Paleozoic evolution of the Argentine Precordillera as a Laurentian rifted, drifted and collided terrane: A geodynamic model. *Geological Society of America Bulletin*, **107**, 253-273.

Bahlburg, H., and Hervé, F. (1997). Geodynamic evolution and tectonostratigraphic terranes of northwestern Argentina and northern Chile. *Geological Society of America Bulletin* **109**, 869-884.

Benedetto, J.P. (1998). Early Palaeozoic brachiopods and associated shelly faunas from western Gondwana: Its bearing on the geodynamic history of the pre-Andean margin. In: *The Proto-Andean margin of South America*, eds. Pankhurst, R.J., and Rapela, C.W., pp. 57-83. Geological Society, London, Special Publication, 142.

Buggisch, W., von Gosen, W., Henjes-Kunst, F. and Krumm, S. (1994). The age of Early Paleozoic deformation and metamorphism in the Argentine Precordillera-Evidence from K-Ar data. *Zlb. Geol. Paläont. Teil I*, H.1-2, 275-286.

Caminos, R. (1973). Some granites, gneisses and metamorphites of Argentina. In: *Symposium on the Granites, Gneisses and Related Rocks of Rhodesia*. Special Publication Geological Society of South Africa, **3**, 333-338.

Caminos, R. (1979). Sierras Pampeanas Noroccidentales Salta, Tucumán, Catamarca, La Rioja y San Juan. In: *Segundo Simposio de Geología Regional Argentina*, Volumen I, ed. Turner, J.C.M., pp. 225-291. Academia Nacional de Ciencias, Córdoba, 869 p.

Dalla Salda, L., Cingolani, C.A., and Varela, R. (1992). Early Paleozoic belt of the Andes and southwestern South America: Result of Laurentia-Gondwana collision? *Geology*, **20**, 617-620.

Dalla Salda, L.H., López de Luchi, M.G., Cingolani, C.A., and Varela, R. (1998). Laurentia-Gondwana collision: the origin of the Famatinian-Appalachian orogenic belt (a review). In: *The Proto-Andean margin of South America*, eds. Pankhurst, R.J., and Rapela, C.W., pp. 219-234. Geological Society, London, Special Publication, 142.

Dalziel, I.W.D. (1997). Overview. Neoproterozoic-Paleozoic geography and tectonics: Review, hypothesis, environmental speculation. *Geological Society of America Bulletin*, **109**, 16-42.

Durand, F.R. (1996). La transición Precámbrico-Cámbrico en el sur de Sudamérica. In: *Early Paleozoic evolution in NW Gondwana*, eds. Baldis, B. and Aceñolaza, F.G., pp. 195-205. Universidad Nacional de Tucumán, Argentina, Serie Correlación Geológica 12.

Gradstein, F.M., and Ogg, J. (1996). A Phanerozoic time scale. *Episodes*, **19**, 3-5.

Huff, W.D., Bergström, S.M., Kolata, D.R., Cingolani, C.A. and Astini, R.A. (1998). Ordovician K-bentonitesin the Argentine Precordillera: relations to Gondwana margin evolution. In: *The Proto-Andean margin of South America*, eds. Pankhurst, R.J. and Rapela, C.W., pp. 107-126. Geological Society, London, Special Publication, 142.

Jeek, P. (1990). Análisis sedimentológico de la Formación Puncoviscana. In: *El Ciclo Pampeano en el Noroeste Argentino*, eds. Aceñolaza, F.G., Miller, H., and Toselli, A.J., pp. 9-36. Universidad Nacional de Tucumán, Argentina, Serie Correlación Geológica 4, 227 p.

Kay, S.M., Orrell, S., and Abruzzi, J.M. (1996). Zircon and whole-rock Nd-Pb isotopic evidence for a Grenville age and Laurentian origin for the basement of the Precordillera in Argentina. *Journal of Geology*, **104**, 637-648.

Keller, M., Buggisch, W., and Lehnert, O. (1998). The stratigraphic record of the Argentine Precordillera and its plate-tectonic background. In: *The Proto-Andean margin of South America*, eds Pankhurst, R.J., and Rapela, C.W., pp. 35-56. Geological Society, London, Special Publication, 142.

Lira, R., Millone, H. A., Kirschbaum, A. M., and Moreno, R. S. (1996). Calc-alkaline arc granitoid activity in the Sierra Norte-Ambargasta Ranges, central Argentina. *Journal of South American Earth Sciences* **10**, 157-177.

McDonough, M. R., Ramos, V. A., Isachsen, C. E., Bowring, S. A., and Vujovich, G. I. (1993). Edades preliminares de circones del basamento de la Sierra de Pie de Palo, Sierras Pampeanas Occidentales de San Juan: sus implicancias para el supercontinente Proterozoico de Rodinia. In: *XII Congreso Geológico Argentino y II Congreso de Exploración de Hidrocarburos, Mendoza, Actas*, **3**, 340-342.

Pankhurst, R.J., and Rapela, C.W. (1998). The proto-Andean margin of Gondwana: An introduction. In: *The Proto-Andean margin of South America*, eds. Pankhurst, R. J., and Rapela, C. W., pp. 1-9. Geological Society, London, Special Publication, 142.

Pankhurst, R.J., Rapela, C.W., Saavedra, J., Baldo, E., Dahlquist, J., Pascua, I. and Fanning, C.M. (1998). The Famatinian magmatic arc in the central Sierras Pampeanas. In: *The Proto-Andean margin of South America*, eds. Pankhurst, R. J., and Rapela, C.W., pp. 343-367. Geological Society, London, Special Publication, 142.

Ramos, V. A., and Basei, M. (1997). The basement of Chilenia: an exotic continental terrane to Gondwana during the early Paleozoic. In: *Terrane Dynamics-97: International Conference on Terrane Geology*, eds. Bradshaw, J. D., and Weaver, S. D., pp. 140-143. University of Canterbury, Christchurch, Conference Abstracts.

Ramos, V. A., Dallmeyer, R. D. and Vujovich, G .I. (1998). Time constraints on the Early Palaeozoic docking of the Precordillera, central Argentina. In: *The Proto-Andean margin of South America*, eds. Pankhurst, R.J. and Rapela, C.W., pp. 143-158. Geological Society, London, Special Publication, 142.

Rapela, C. W., Pankhurst, R. J., Casquet. C., Baldo, E., Saavedra, J., Galindo, C. and Fanning, C.M. (1998a). The Pampean orogeny of the southern proto-Andes: evidence for Cambrian continental collision in the Sierras de Córdoba. In: *The Proto-Andean margin of South America*, eds. Pankhurst, R.J., and Rapela, C.W., pp. 181-217. Geological Society, London, Special Publication, 142.

Rapela, C. W., Pankhurst, R. J., Casquet. C., Baldo, E., Saavedra, J., Galindo, C. (1998b). Early evolution of the proto-Andean margin of South America. *Geology* **26**, 707-710.

Rapela, C. W., Pankhurst, R. J., Dahlquist, J. and Fanning, C.M. (1999). U-Pb SHRIMP ages of Famatinian granites: new constraints on the timing and tectonic setting of I- and S-type magmas in an ensialic arc. In: *II South American Symposium on Isotope Geology*, Carlos Paz, Argentina, Conference abstracts (in press).

Rossi, J. N., Toselli, A. .J. and Durand, F.R. (1992). Metamorfismo de baja presión, su relación con el desarrollo de la cuenca Puncoviscana, plutonismo y régimen tectónico. *Estudios Geológicos*, **48**, 279-287.

Sims, J.P., Ireland, T.R., Camacho, A., Lyons, P., Pieters, P.E., Skirrow, R.G., and Stuart-Smith, P.G. (1998). U–Pb, Th–Pb and Ar–Ar geochronology from the southern Sierras Pampeanas, Argentina: implications for the Palaeozoic tectonic evolution of the western Gondwana margin. In: *The Proto-Andean margin of South America*, eds Pankhurst, R. J., and Rapela, C. W., pp. 259-281. Geological Society, London, Special Publication 142.

Turner, J.C.M., ed. (1979). *Segundo Simposio de Geología Regional Argentina*, Volumen 1. Academia Nacional de Ciencias, Córdoba, 869 p.

Varela, R., López de Luchi, M., Cingolani, C., and Dalla Salda, L. (1996). Geocronología de gneises y granitoides de la Sierra de Umango, La Rioja. Implicancias tectónicas. *XIII Congreso Geológico Argentino y III Congreso de Exploración de Hidrocarburos, Buenos Aires, Actas*, **3**, 519-528.

Sedimentary Basins of South America

Edison José Milani and Antonio Thomaz Filho

Sedimentary domains of the South American Plate

Regarding the distribution of Phanerozoic sedimentary successions and particular plate-tectonic regimes, four major domains are recognized in the South American Plate: the continental interior, a region of a long and complex geological history constituting a vast realm where thick sequences of Paleozoic and Mesozoic sediments accumulated; the western, convergent margin of the plate, along which the continental lithosphere of South America confronts the oceanic floor of the Pacific Ocean, creating a large orogenic belt, the Andes, with several sedimentary basins associated; the eastern side, a more than 10 000 km-long divergent margin originated by the break-up of the Gondwana paleocontinent and the separation of the South American and African plates since the Mesozoic; and the northern and southern margins of the plate, marked by a regional transcurrent tectonic regime along major transform faults that define the active contact of the South American Plate with, respectively, the Caribbean and the Scotia plates (Fig. 1). This review attempts include the most representatives sedimentary provinces that occur along these large-scale tectono-sedimentary domains of the continent; being the text on the Brazilian intracontinental sags and failed rift basins, a condensed version of Milani and Zalán (1998).

The interior region of the plate is known as the South American Platform (Almeida *et al.*, 1977; Almeida and Hasui, 1984), and is composed mainly of metamorphic and igneous complexes of Archean age. This domain forms the core of the plate, and consolidated by Late Proterozoic-Early Paleozoic times after the collage of the Brasiliano-Panafrican Cycle. This first-order tectono-magmatic cycle corresponds to an ancient amalgamation of several distinct landmasses and closure of several seaways situated among them, originating in this way the nucleus of Gondwana. Crystalline rocks of Precambrian age are extensively exposed in three major shield areas known as Guiana, Central Brazil, and Atlantic. Some of the zones of crustal weakness of the Brasiliano-Panafrican framework were used afterwards as major lines of control over the evolution of the Paleozoic intracratonic basins (Tankard *et al.*, 1995) as well as over the rupture lines during the Mesozoic rifting and opening of the South Atlantic Ocean.

Phanerozoic cratonic sequences developed extensively from Late Ordovician time until the Cretaceous, and the five major synclises (Fig. 2) were formed: Solimões, Amazonas, Parnaíba and Paraná in Brazil, and Chaco-Paraná in Argentina, Paraguay and Uruguay. These occupy the about 3.5 M km^2 of South America, as large remnants of a certainly much wider original area of sedimentation. The intracratonic sequences in the South American Platform configure a series of unconformity-bounded units (Sloss, 1963; Soares *et al.*, 1978) that resulted from successive phases of subsidence and accumulation of sedimentary rocks, the record of which was interrupted during several periods of widespread erosion.

Towards the SW, a persistently active convergent setting developed during almost all the Phanerozoic era along the margin of the Gondwana, defining a deformational belt known as the Gondwanides (Keidel, 1916; De Wit *et al.*, 1988). Along this mobile belt, sedimentation, tectonics, and magmatism kept through time a similar style to that observed today in the Andean Chain and related basins. Collisional episodes due to terrane accretion along the margin were remarkable (Ramos, 1988), and originated several orogenic cycles. In other words, the present-day tectonic regime and physiographic configuration of the western margin of the continent perpetuates the long-lived convergent character of that realm.

The eastern margin of the South American Plate is younger, having evolved from Mesozoic times onward. The development of the Atlantic Ocean since the break-up of the Gondwana left a large divergent margin basin encroached on the continent. The tectono-sedimentary evolution of the passive margin (Asmus and Ponte, 1973) included a rift stage, during which lacustrine sediments were deposited. In Aptian times, a large salt basin developed and marked the transitional stage that was characterized by restricted circulation followed by a cycle of open marine sedimentation under the rapid invasion by waters of the South Atlantic Ocean. Aborted Mesozoic interior rift basins indented the eastern margin in several places, and the Atlantic rifting resulted in a pervasive intrusion of dykes and sills and extrusion of huge volumes of basic lavas in the adjacent Paleozoic synclises.

To the N, the margin of the South American Plate lies in contact with the Caribbean Plate along a large transform boundary, the Oca-San Sebastian-El Pilar Fault, since the early Cenozoic. Sedimentation and tectonics in that region were greatly influenced by right-lateral wrenching. Similar strike-slip conditions are found along the southern border of the South American Plate, where it make contact with the Scotia Plate along the North Scotia Ridge, a left-lateral transform zone.

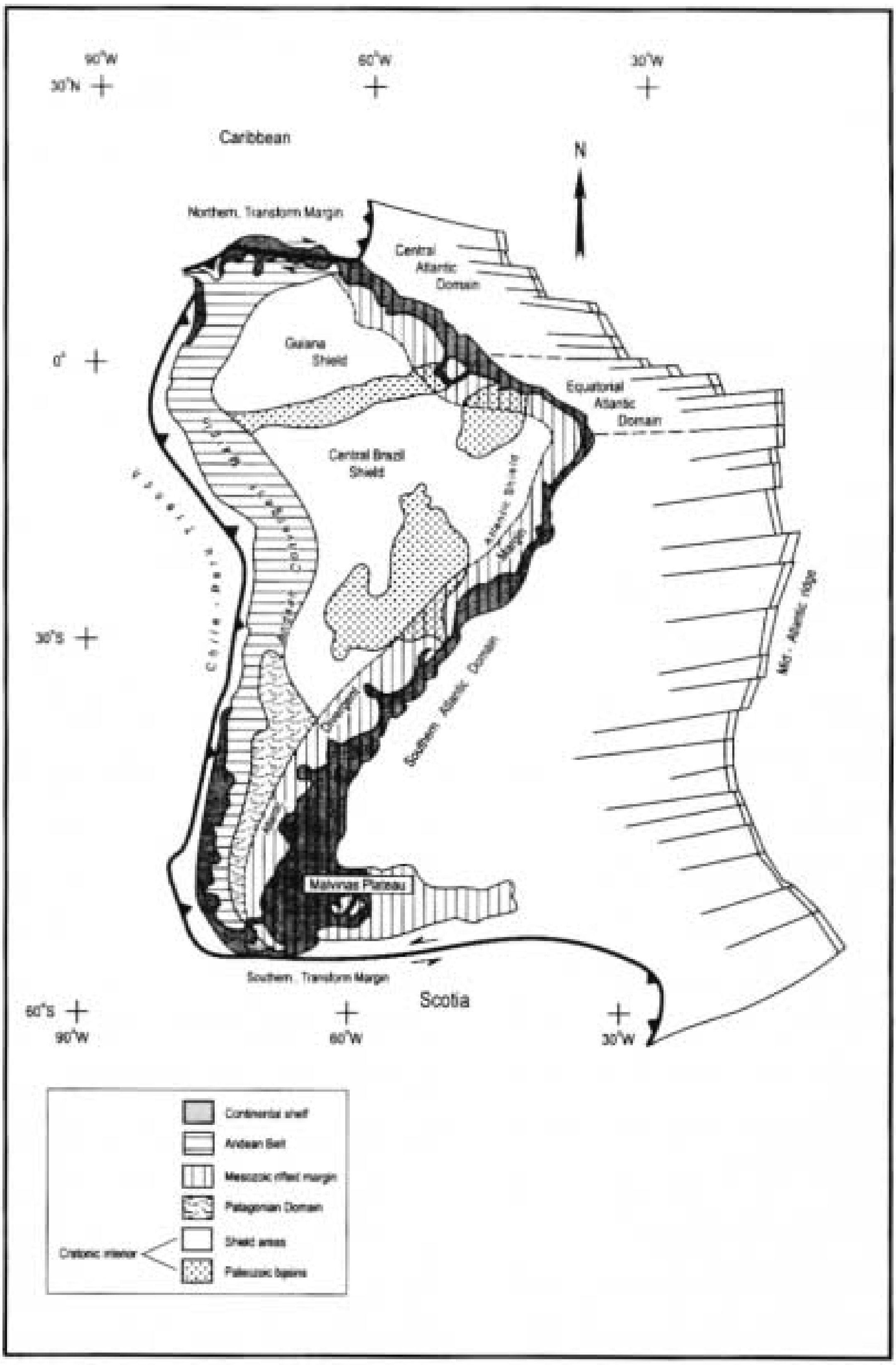
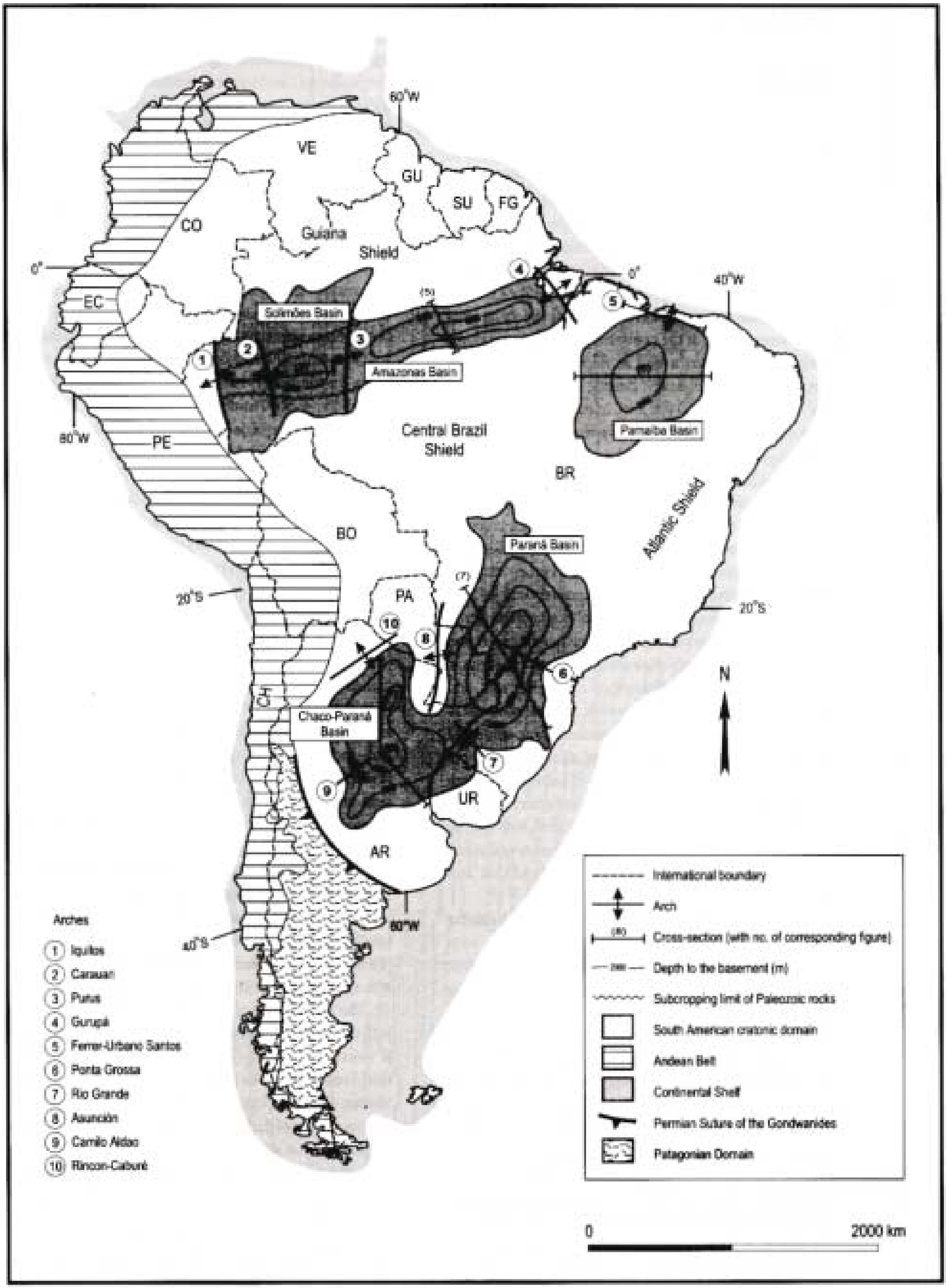


FIGURE 1 - The South American Plate and its major tectono-sedimentary domains.

FIGURE 2 - Situation map of the Paleozoic intracontinental sags of South America (compiled from various authors). The South American cratonic domain represents the core of the continent, and this large region achieved a final stable condition during the late Proterozoic to earliest Paleozoic Brasiliano Orogenic Cycle (Almeida and Hasui, 1984). Countries: FG-French Guiana, SU-Surinam, GU-Guiana, VE-Venezuela, CO-Colombia, EQ-Equator, PE-Peru, BO-Bolivia, CH-Chile, AR-Argentina, PA-Paraguay, UR-Uruguay, and BR-Brazil.


60°W
VE
GU
SU
FG
CO
Guiana
Shield
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EC
Solimões Basin
Amazonas Basin
40°W
Parnaíba Basin
80°W
PE
Central Brazil
Shield
BR
Atlantic Shield
BO
Paraná Basin
PA
20°S
CH
Chaco-Paraná
Basin
UR
AR
N
Arches
1 Iquitos
2 Carauari
3 Purus
4 Gurupá
5 Ferrer-Urbano Santos
6 Ponta Grossa
7 Rio Grande
8 Asunción
9 Camilo Aldao
10 Rincon-Caburé
40°S
International boundary
Arch
Cross-section (with no. of corresponding figure)
Depth to the basement (m)
Subcropping limit of Paleozoic rocks
South American cratonic domain
Andean Belt
Continental Shelf
Permian Suture of the Gondwanides
Patagonian Domain
0
2000 km

Cratonic Interior

Paleozoic intracontinental sags

Paleozoic intracontinental sedimentary successions and related magmatic rocks in South America are preserved in five individual depocenters (Milani and Zalán, 1999), four of them in Brazil, and one in Argentina (Fig. 2). The Brazilian basins are the Solimões, Amazonas, Parnaíba and Paraná, the names of which are derived from the major rivers that flow along the main axis of each. The sedimentary successions of the Chaco-Paraná Basin of Argentina lie below the Cenozoic deposits in the sub-Andean flats that developed along most of the western part of the continent. The interior cratonic basins of South America have an elliptical to semi-circular shape with individual basins ranging in size from 500 000 km^2 to more than 1 000 000 km^2. In common, these basins have such features as a relatively simple structural and stratigraphic framework and the presence of enormous volumes of Mesozoic basaltic magma.

Intracratonic subsidence is still a matter of debate (Milani, 1999). To date no convincing mechanism has been proposed to explain the subsidence and to justify how these basins had their subsidence history marked by several episodes of reactivation, giving rise to a pile of sedimentary and volcanic strata that can reach a thickness of several thousands of metres. Rifts are known below some of these intracratonic basins, and some of them are certainly related to the origin of the Paleozoic synedises. However, sometimes their much older ages, and sometimes their restricted distribution, are clear obstacles to the hypothesis of considering an extensional, fault-controlled mechanism of subsidence as the general, direct predecessor of the overlying sag. Additional cycles of subsidence in the intracratonic basins could have been achieved either by different events of lithosphere stretching (Gonzaga *et al.*, 1997), or, by lithosphere flexure due to cratonward influence of distant, plate-margin orogenies (Johnson, 1971; Milani and Ramos, 1998).

Whatever the driving mechanisms of the subsidence events, the depressions were filled essentially with siliciclastic rocks related to successive Paleozoic transgressive-regressive cycles, with the remarkable exception of an evaporite-carbonate cycle in the Pennsylvanian-Permian record of the Solimões, Amazonas and Parnaíba basins. In most of these basins, the first cycle of sediment accumulation took place during the Late Ordovician (Fig. 3), when the invasion of seas over Gondwana was widespread. Glacial influence over the sedimentary environment is recorded at various ages (Caputo and Crowell, 1985), between the latest Ordovician in the Paraná Basin and the Devonian in the Solimões and Amazonas basins. In the Paraná and Chaco-Paraná basins, the glacial influence returned, and was extreme in the Mississipian with the development of ice caps over the area. The progressive retreat of the glaciers during the Pennsylvanian allowed the development of the third transgressive-regressive cycle of sedimentation.

The last Paleozoic depositional cycle in all these five basins terminates with Late Permian to Early Triassic continental red beds, marking the drying out of the intracratonic basins, and the final disappearance of the seas from the interior areas of Western Gondwana. These arid conditions proceeded during the Mesozoic and eolian deserts dominated the Late Jurassic depositional scenery, associated with large volumes of magmatic material already related to the mechanisms that led to the break-up of Gondwana.

Solimões Basin

The Solimões Basin, comprising an area of over 600 000 km^2, is situated in northern Brazil and is entirely covered by the Amazonas forest. The occurrence of Paleozoic packages is restricted to two thirds of the total area of the basin, and these are not exposed, being onlapped and covered by sandy, Cretaceous to Recent continental sediments. Together the Solimões Basin and its eastern counterpart, the Amazonas Basin, constitute an E-W elongated, 2500 km-long, 500 km-wide and up to 5000 m-deep interior basin (Fig. 2).

The Solimões Basin is separated from the Acre Basin to the W and from the Amazonas Basin to the E by, respectively, the Iquitos and Purus arches (Fig. 2). The structural framework of the Solimões Basin shows a prominent NW-SE-striking positive feature, the Carauari Arch (Fig. 4), that define two sub-basins: Jandiatuba and Juruá. Mainly during pre-Pennsylvanian times, this positive feature had a decisive control on the distribution of sedimentary thickness and facies. The total sedimentary sequence in the deeper, Juruá Sub-Basin attains 3800 m.

The structural framework of the Solimões Basin is marked by the presence of the Solimões Megashear Zone (Caputo and Silva, 1990). This constitutes a classical example of an intraplate deformation belt, comprising a fault-and-fold system that strikes N70°-80°E, over a distance of about 1 000 km. The belt, caused by Jurassic to Cretaceous right-lateral wrenching, is defined by regularly spaced, right-stepping *en échelon* reverse faults trending N30°-40°E with associated hangingwall anticlines.

The stratigraphic framework of the Solimões Basin (Fig. 3) is defined by four Paleozoic supersequences covered by two units, one of Cretaceous and the other of Cenozoic age. During Late Triassic to Early Jurassic times, the Penatecaua Magmatism (Eiras *et al.*, 1994a) intruded the Paleozoic sedimentary units with a large volume of diabase sills and dykes.

The sedimentary record of the Solimões Basin starts with the Ordovician Supersequence (Eiras *et al*, 1994a), that is constituted by the continental to shallow marine siliciclastic rocks of the Benjamin Constant Formation. The Silurian-Devonian Supersequence follows and is constituted by a succession of dark grey marine shales named the Jutaí Formation, that grade eastwards into coastal sandstone units and micaceous siltstone of the Biá Member, marking the facies change due to the onlap of the sequence over the Carauari High.

The silexite and sandstone beds of the Uerê Formation, together with the black shale, rich in organic material, diamictite and sandstone of the Jandiatuba Formation, span the Emsian-Tournaisian range and constitute the Devonian-Carboniferous Supersequence. The section, up to 300 m thick, is considered of shallow marine origin, transgressive up to the Frasnian beds and remarkably regressive upwards.

From Carboniferous times onwards, the sediments

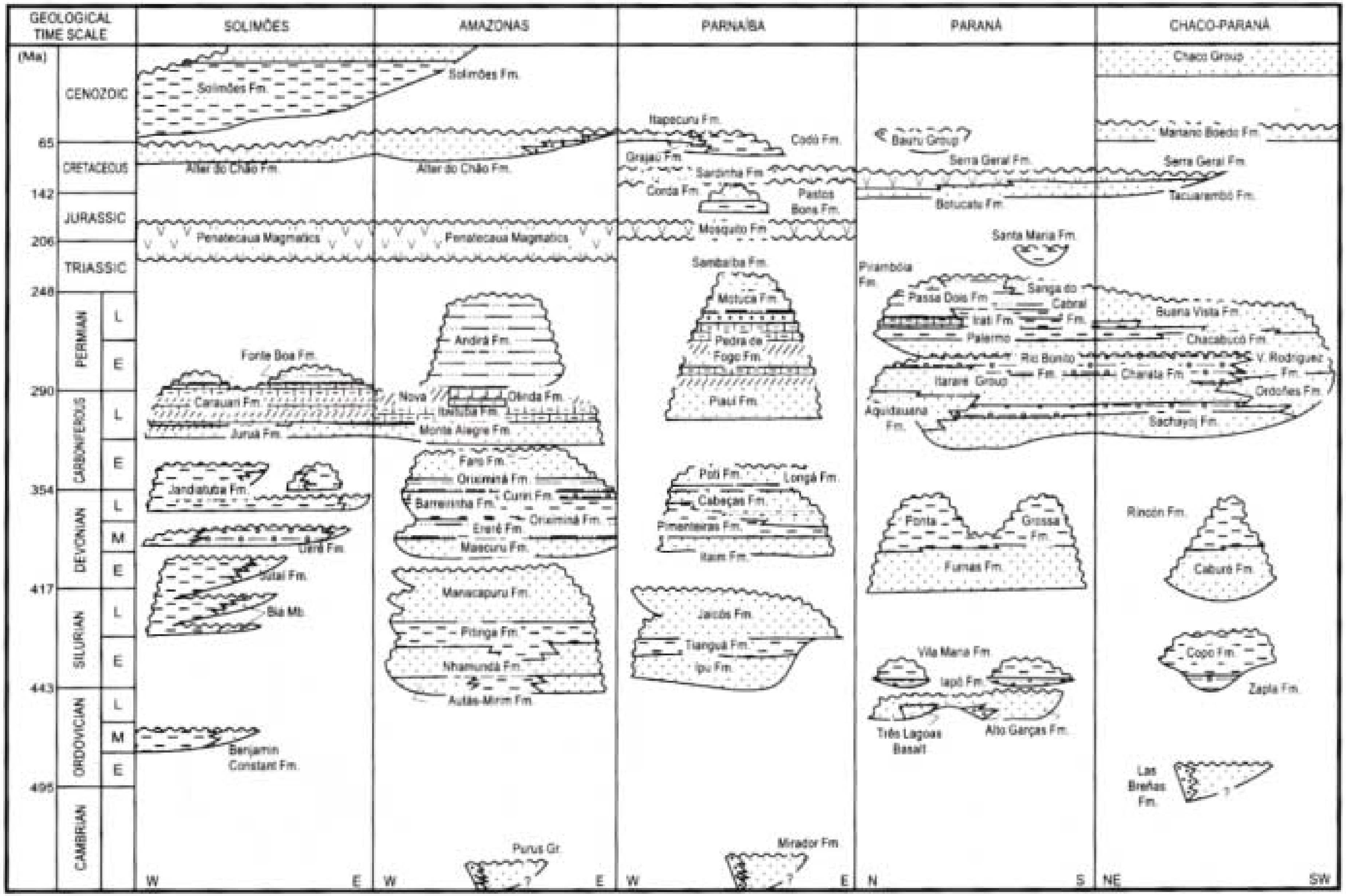


FIGURE 3 - Stratigraphic summary of the Paleozoic intracontinental sags of South America. Main sources of information were Eiras et al. (1994a), Solimões Basin; Cunha et al. (1994), Amazonas Basin; Góes and Feijó (1994), Parnaíba Basin; Milani et al. (1994), Milani (1997), Paraná Basin; Russo et al. (1987), Jager (1997), Pezzi and Mozetic (1989), Chaco-Paraná Basin. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.

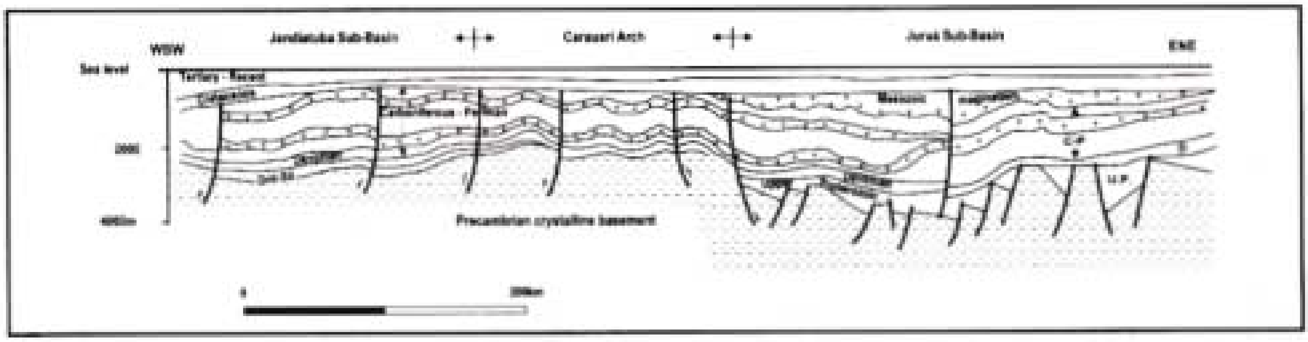


FIGURE 4 - Cross-section of the Solimões Basin (after Eiras, 1996; permission to reprint granted by the author).

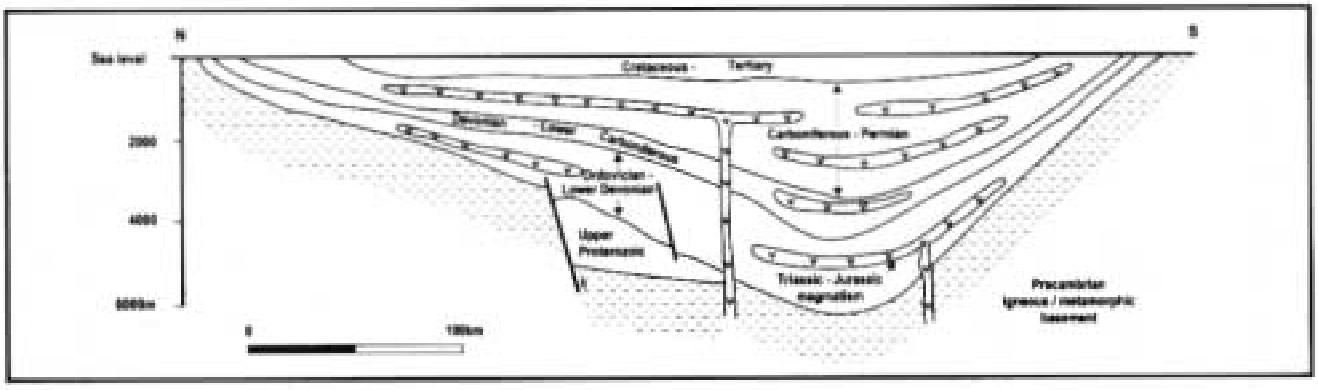


FIGURE 5 - Cross-section of the Amazonas Basin (after Gonzaga et al., 1997; permission to reprint granted by the authors).

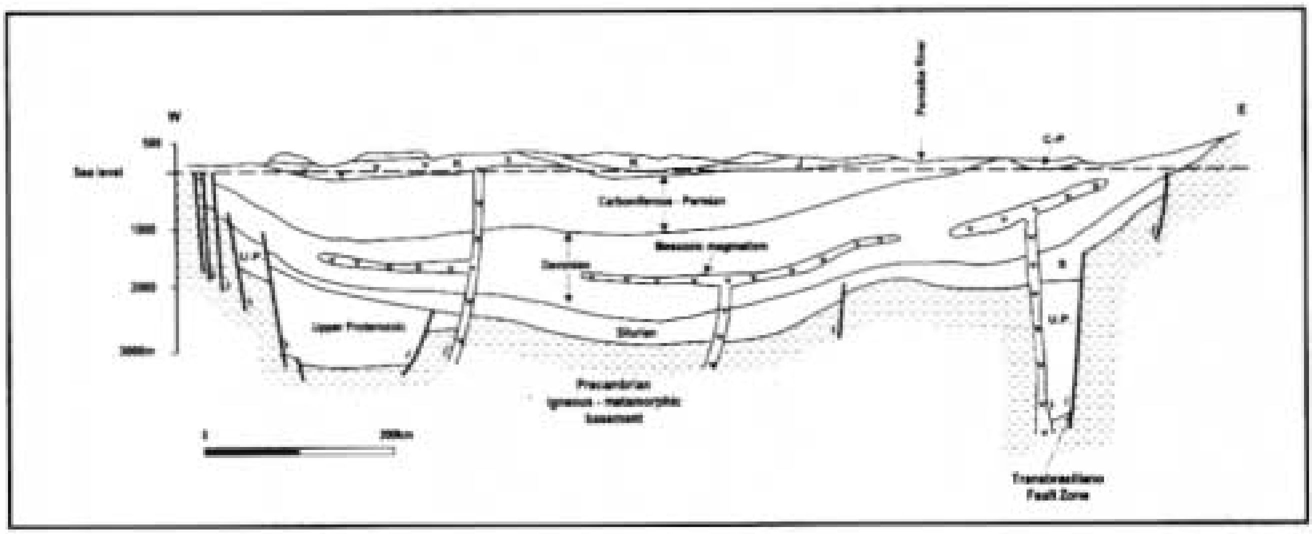


FIGURE 6 - Cross-section of the Parnaíba Basin (after Góes et al., 1990; permission to reprint granted by the authors).

covered the barrier once formed by the Purus Arch and so Amazonas and Solimões basins became united into a single depositional basin (Caputo and Silva, 1990). The accumulation of the initial units of the Carboniferous-Permian Supersequence occurred under conditions of severe aridity, with partial continentalization of the entire region, allowing deposition of a blanket of eolian sandstone beds known as the Juruá Formation. This formation grades upwards into an up to 1300 m thick package of rhythmic carbonate and evaporite units of the Carauari Formation. The sequence is terminated by a section of fluvial and lacustrine red beds named the Fonte Boa Formation.

A regional unconformity, with a mild angularity along the belt of deformation of the Solimões Megashear Zone, separates the Paleozoic record from the younger strata. The Javari Group (Eiras *et al.*, 1994a), including the Alter do Chão and Solimões formations, represents, respectively, the Cretaceous and Cenozoic Supersequences of the Solimões Basin, the former spreading into the Amazonas Basin, and the latter developing mostly towards the W and reaching the Acre Basin. The Alter do Chão Formation consists of coarse-grained ferruginous sandstone beds that accumulated in an alluvial-fluvial system. The Solimões Formation is an argillaceous unit, including beds of coquina and lignite, holding abundant fauna and flora of Miocene/ Pliocene age.

Amazonas Basin

This basin covers about 500 000 km² of northern Brazil, lying completely under the Amazonas forest. The major axis trends E-W (Fig. 2), and its sedimentary strata onlap the Precambrian provinces of the Guiana Shield to the N and of the Central Brazil Shield to the S, cropping out along both these sides of the basin. To the E, a Mesozoic rift shoulder, the Gurupá Arch, isolates the Amazonas Basin from a Mesozoic failed rift, the Marajó Basin. To the W, the Amazonas Basin is separated from the Solimões Basin by the roughly N-S-trending Purus Arch. The depocenter of the basin holds a stratigraphic record up to 5000 m thick (Fig. 5).

The basement of the Amazonas Basin consists mainly of igneous and metamorphic complexes, but some Proterozoic rift successions of alluvial, fluvial and lacustrine sedimentary rocks occur in its western region, comprising the Purus Group. The existence of strong positive gravity anomalies coincident with the axis of the syneclise, suggesting shallow ultrabasic bodies (Linsser, 1974), together with the underlying aulacogenic strata, led to the interpretation of crustal stretching as the driving mechanism for the initial subsidence of the Amazonas Basin. More recent work (Gonzaga *et al.*, 1997) interpreted four extensional events during the evolution of the Amazonas Basin, spanning the ranges Ordovician-Early Devonian, Devonian-Early Carboniferous, Middle Carboniferous-Permian and Cretaceous-Cenozoic. Each of these events was manifest by relatively high sedimentary accumulation rates during the interpreted extensional phases, succeeded by periods of low sedimentary accumulation rates driven by thermal subsidence.

The stratigraphic framework of the Amazonas Basin (Fig. 3) includes three Paleozoic supersequences covered by Cretaceous to Recent continental clastic sediments (Cunha *et al.*, 1994). The Ordovician-Devonian Supersequence or Trombetas Group, is composed of marine sandstone included in the Autás-Mirim, Nhamundá and Manacapuru formations, and shale and diamictite beds of the Pitinga Formation. These sediments in part accumulated under the influence of the Late Ordovician-Early Silurian glaciation, that affected large parts of Gondwana.

A second transgressive-regressive cycle is recorded in the Amazonas Basin by the sedimentary rocks of the Urupadi and Curuá groups that together comprise the Devonian-Carboniferous Supersequence. This large stratigraphic unit includes the Maecuru and Ererê fomations (Early Devonian), consisting of coastal sandstone and shallow marine shale units at the base, followed by an up to 250 m thick sequence of laminated black shale beds named the Barreirinha Formation, of Frasnian age. The Barreirinha Formation is considered to indicate the maximum paleobathymetric conditions during the history of the basin. A controversial Famennian glacial cycle (Daemon and Contreiras, 1971) is interpreted from the record of the Curiri Formation consisting of diamictite with associated shale and siltstone units. The upper part of this sequence, Oriximiná and Faro formations, record the regressive part of the cycle. Regional erosive processes affected the basin during the final stages of the Mississipian.

A renewed cycle of subsidence and sediment accommodation took place from the Pennsylvanian onwards. The lowermost beds of the Carboniferous-Permian Supersequence, or Tapajós Group, is a blanket of eolian sandstone included in the Monte Alegre Formation, that is covered by a sequence of carbonate and evaporite units with subordinated sandstone and shale beds that define the Itaituba and Nova Olinda formations. These stratigraphic units attain a maximum thickness of 1600 m. The cycle closed by the continental red beds of the Andirá Formation, of Late Permian age. E-W regional extension allowed a pervasive intrusion of magmatic bodies during Late Triassic and Early Jurassic times; the Penatecaua Magmatism (Mizusaki *et al.*, 1992).

Parnaíba Basin

The Parnaíba Basin is located in the western portion of the northeastern region of Brazil, occupying an area as large as 600 000 km². The Parnaíba Basin is a circular sag with a total sedimentary section of about 3500 m of thickness in its depocenter, and is framed by an almost continuous belt of outcrops (Fig. 2). The Ferrer-Urbano Santos Arch, a positive flexural feature related to the Mesozoic opening of the Equatorial Atlantic Ocean, defines the northern limit for the Parnaíba Basin, where the Atlantic rifting broke the previous connection of this huge syneclise with analogous basins that lie today in northwestern Africa.

Geophysical data, together with subsurface information, revealed the existence of a basal sequence of arkose confined inside Late Proterozoic/Early Cambrian N-S-trending grabens lodged by the crystalline basement. This sandy section of poorly constrained age, known as Mirador Formation, was interpreted as being the rift sequence that initiated the Parnaíba Basin (Góes *et al.*, 1990). A flexural sag developed afterwards (Fig. 6), but an influence of the initial rift architecture on the location of cratonic sedimentation depocenters was felt up to the Early Carboniferous.

A remarkable feature in the structural framework of the Parnaíba Basin is the Transbrasiliano Fault Zone, that cuts its eastern/southern portion and produced important structures in that domain. This fault zone, a deep-seated structural element in central Brazil, runs across the continent for 3000 km, from N to S in a NE-SW trend, starting in the Equatorial Atlantic and reaching Paraguay, crossing also the northernmost portion of the interior Paraná Basin. Within the Parnaíba Basin the fault zone is expressed as a wide band of surface and subsurface deformation that controlled the positioning of old aulacogens and Paleozoic depocenters. Outside the limits of the Transbrasiliano Fault Zone, the structural style in the Parnaíba Basin is almost featureless, being mostly represented by subtle fault-block tectonics. Intrusion of magmatic bodies during the Mesozoic was widespread and was also responsible for typical related deformation of the host sedimentary rocks.

The stratigraphic framework of the Parnaíba Basin (Fig. 3) encompasses three major Paleozoic and two Mesozoic supersequences (Góes and Feijó, 1994). The rocks of the Serra Grande Group, mainly of Silurian age, represent the older strata, accommodated by the initial subsidence of the syneclise. These rocks include sandstone with subordinate siltstone, shale and diamictite beds named the Ipu Formation; grey shale, siltstone and micaceous coastal sandstone of Wenlock age known as the Tianguá Formation; and coarse-grained sandstone typical of braided stream deposits known as the Jaicós Formation. The entire sequence represents a complete transgressive-regressive cycle of sedimentation.

The Devonian Supersequence or Canindé Group, of Eifelian-Tournaisian age, has tide and storm-related sediments in the Itaim Formation; black shale, rich in organic mater, up to 300 m thick known as the Pimenteiras Formation; fine-grained shallow marine sandstone, locally with diamictite beds, named the Cabeças Formation; fluvial sandstone and shale in the Longá Formation; and an Early Carboniferous heterolythic sequence of fine-grained sandstone and shale units known as the Poti Formation.

The beginning of the Pennsylvanian sedimentation in the Parnaíba Basin was marked by significant modifications in both environmental and tectonic characteristics (Góes and Feijó, 1994). Depocenters changed from an elongated graben-controlled geometry to a circular configuration, and the condition of a temperate, open marine basin turned into a restricted sea under progressive aridity (Caputo, 1984). The Balsas Group, or Carboniferous-Triassic Supersequence, consists of a clastic-evaporitic succession of sedimentary rocks with fine- to medium-grained sandstone, brown shale and subordinated limestone units, the Piauí Formation; rhythmite with alternating oolitic limestone units, white anhydrite and yellow sandstone beds in the Pedra de Fogo Formation; brown siltstone, medium-grained sandstone, locally with anhydrite beds named Motuca Formation; and eolian sandstone in the Sambaíba Formation.

Two main pulses of magmatic activity took place in the Parnaíba Basin during the Mesozoic. Both intrusive emplacements and volcanic flows occurred, the former being preferentially found inside the rocks of the Canindé Group. The first magmatic cycle, from the Triassic-Jurassic and correlated to the Penatecaua Magmatism of the Solimões and Amazonas basins, is known as the Mosquito Formation, and is related to the rifting of the Central Atlantic. The second pulse, aged Early Cretaceous and related to the rifting of the South Atlantic, constitutes the Sardinha Formation.

Between the two magmatic pulses, a phase of continental sedimentation developed, constituting the Mearim Group of Jurassic age. This package, of restricted area of occurrence, includes green to red argillite and siltstone in the Pastos Bons Formation and sandstone in the Corda Formation, being the section related to an eolian environment. Also related to the development of the Equatorial Atlantic are the Codó, Grajaú and Itapicuru formations, that constitute the Cretaceous Supersequence of the Parnaíba Basin. The Aptian Codó Formation is made of black shale rich in organic material, accumulated during a brief incursion of Mesozoic oceanic waters over the northern part of South America.

Paraná Basin

The Paraná Basin is located in southern Brazil, where it covers an area exceeding 1 000 000 km^2, extending into the neighbouring countries of Paraguay, Argentina and Uruguay, and there it occupies another 400 000 km^2. The basin has a NNE-SSW-trending elliptical shape with two thirds of its surface covered by basaltic lava of Mesozoic age (Fig. 2). The infilling sedimentary strata appear along a 5500 km-long belt of outcrops shaped during Mesozoic-Cenozoic times, encircling almost completely, the syneclise. The stratigraphic record of this huge basin, including sedimentary and magmatic rocks, surpasses the thickness of 7000 m in the central part of the basin, just in the region where runs the river that lends its name to the basin (Fig. 7).

The eastern flank of the Paraná Basin corresponds to a crustal region that was deeply affected by the South Atlantic rifting and the opening of the ocean. Uplift and erosion since the Early Cretaceous has been responsible for the removal of large amounts of Paleozoic sedimentary rocks from that area. The Asunción Arch defines the western border of the basin, a flexural bulge related to the loading of the Cenozoic Andean thrust sheets along the western margin of the plate. To the N and to the S, the strata of the Paraná Basin onlap crystalline rocks of Precambrian cratonic provinces of the basement.

The inception and further evolution of the Paraná Basin in the continental interior of Gondwana, since the Late Ordovician, had a close relationship to the development of the Gondwanides, an extensive Phanerozoic mobile belt marked by a succession of orogenic cycles (Ramos *et al.*, 1986). Such orogenic stresses exerted a decisive influence on the creation of accommodation space for the various supersequences that fill the Paraná Basin (Milani, 1997), as well as on the deformational history of the basin (Zalán *et al.*, 1990) by means of intraplate reactivation of NE-SW trending preexisting zones of crustal weakness, inherited from the Brasiliano Orogeny or cratonward propagation of regional flexural subsidence.

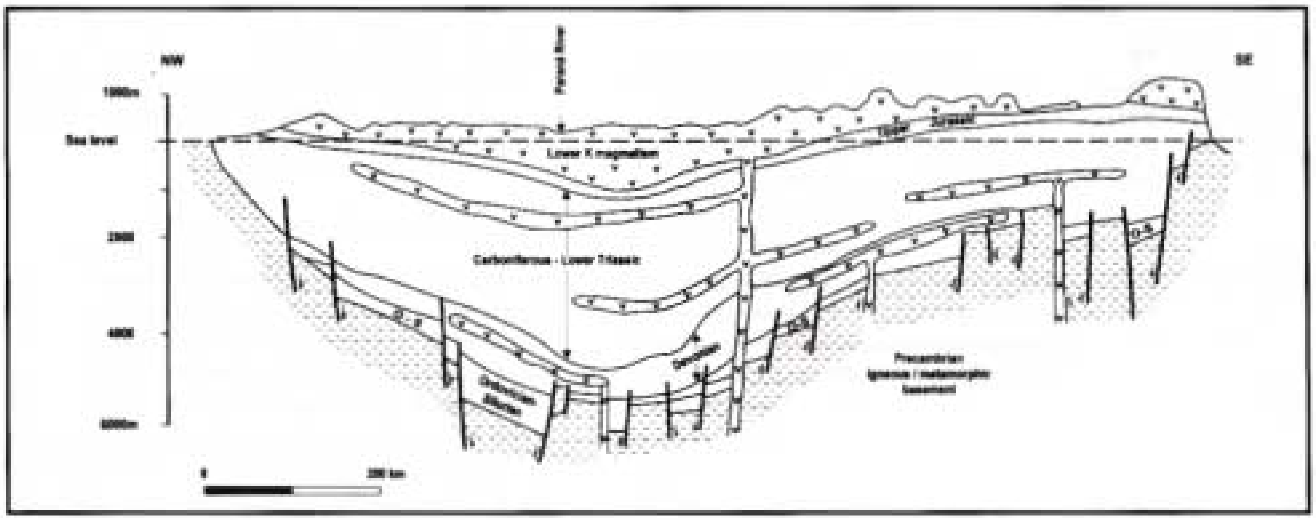


FIGURE 7 - Cross-section of the Paraná Basin (after Milani and Zalán, 1998).

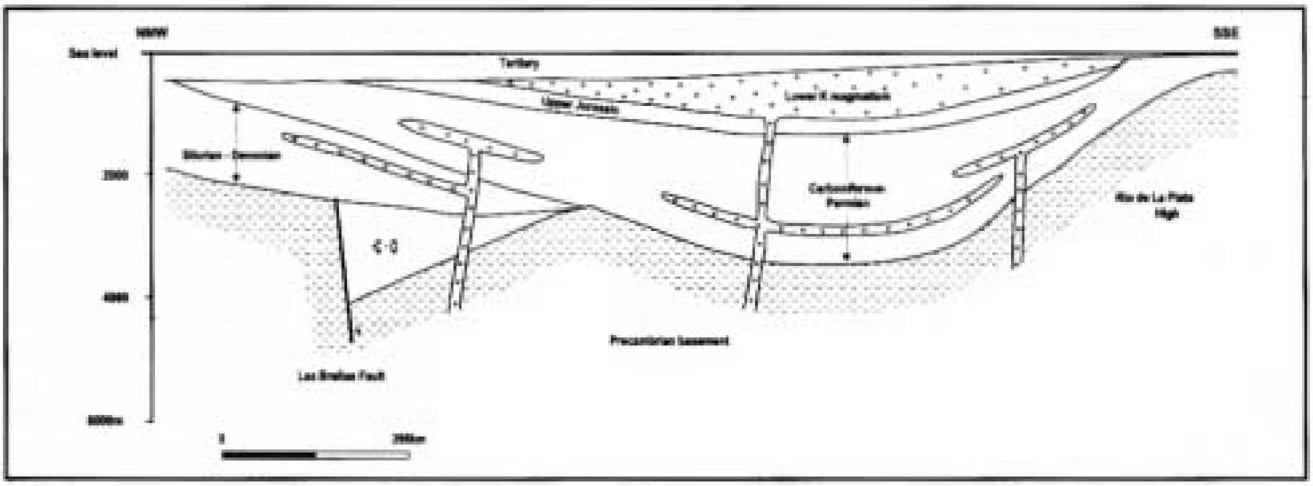


FIGURE 8 - Conceptual cross-section of the Chaco-Paraná Basin (based on information from Pezzi and Mozetic, 1989; Russo et al., 1987).

The stratigraphic record of the Paraná Basin (Fig. 3) includes six supersequences (Milani *et al.*, 1994; Milani, 1997): Rio Ivaí (Caradocian-Llandoverian), Paraná (Lochkovian-Frasnian), Gondwana I (Westphalian-Scythian), Gondwana II (Anisian-Norian), Gondwana III (Late Jurassic-Berriasian) and Bauru (Aptian-Maastrichtian). Three of these second order allostratigraphic units correspond to Paleozoic transgressive-regressive cycles, and the others are Mesozoic continental sedimentary sequences with associated igneous rocks.

The Rio Ivaí Supersequence begins with a basal sandy unit known as Alto Garças Formation, up to 300 m thick, followed by some tens of metres of diamictite beds defining the Iapó Formation that record the Late Ordovician-Early Silurian glaciation of Gondwana. The uppermost unit of the supersequence, the Vila Maria Formation, consists of micaceous shale and fine-grained sandstone containing Llandovery-age macro and microfossils. Some of the best reference sections for the Ordovician-Silurian sequence of the Paraná Basin can be found in its Paraguayan sector, where this interval is known as Caacupé and Itacurubi groups.

The Devonian sedimentary beds of the Paraná Basin, also described as the Paraná Supersequence, spans the range Pragian-Frasnian and are composed of a basal blanket of coarse-grained, kaolinite-rich sandstone beds, the Furnas Formation, covered by a shaly section with subordinate siltstone and deltaic sandstone units that constitute the Ponta Grossa Formation. The Devonian shale beds, particularly its lowermost section of Emsian-Eifelian age, hold the endemic Malvinokaffric fauna of invertebrates (Melo, 1988). Bioturbation is a dominant characteristic in these shales, but some intervals of black laminated rocks also occur, recording maximum paleobathymetric conditions of the Devonian sea. In Uruguay, the Devonian section is named Durazno Group.

The Gondwana I Supersequence is the most voluminous packet of sedimentary rocks in the Paraná Basin among all its second-order allostratigraphic units, with a thickness that reaches a maximum of about 2500 m. This is a sedimentologically heterogeneous and complex section that records the contrasting paleoenvironmental changes through time that occurred in Gondwana from the glacial period in the Pennsylvanian to the dry and arid times during the Triassic.

The lowermost section of the Gondwana I Supersequence is represented by a sequence of glaciogenic deposits included in the Itararé Group and Aquidauana Formation in Brazil, the Coronel Oviedo Formation in Paraguay, and the San Gregório Formation in Uruguay. These rocks are composed of diamictite, turbiditic sandstone units, conglomerate, subordinate varvite and tillite beds (França and Potter, 1988). The maximum thickness is about 1500 m. This section represents the thickest and stratigraphically most complete record of the Late Carboniferous-Early Permian glaciation-related sedimentary rocks known in Gondwana, spanning the Westphalian to Artinskian range.

The glaciogenic deposits were followed by a transgressive section known as Guatá Group, that includes deltaic sandstone and coal beds of Artinskian-Kungurian age, the Rio Bonito Formation; and siltstone and shale in the Palermo Formation, the latter representing the maximum flooding stage for the entire Gondwana I Supersequence (Milani, 1997). In the Kazanian, a section of bituminous shale and limestone beds with subordinated evaporite units named the Irati Formation was deposited. The Irati black shale exhibits an organic carbon content as high as 23%, a value that is among the highest of known sedimentary rocks the world over. The shales also carry the characteristic reptilian *Mesosaurus* fauna, that early in this century permitted the correlation of the Gondwana strata of southern Brazil with those in southern Africa (Du Toit, 1927), thus supporting the nascent theory of continental drift.

The Gondwana I Supersequence is terminated by the deposits included in the upper portion of the Passa Dois Group, represented by the sequence of red beds of the Rio do Rasto Formation, accumulated through the dawn of the Mesozoic. Continental facies up to several hundreds of metres thick encroached the remnants of the Rio do Rasto Basin from the northern and southern borders, constituting stratigraphic units named, respectively, Pirambóia and Sanga do Cabral formations. In Uruguay and northeastern Argentina, these continental successions are even thicker and are named Buena Vista Formation.

Lacustrine red beds of local occurrence and associated fluvial and eolian deposits constitute the Gondwana II Supersequence of the Paraná Basin. This section holds a remarkable Middle to Late Triassic tetrapod fauna (Barberena *et al.*, 1985). Huge seas of eolian dunes, the Botucatu Formation, of widespread regional distribution, dominated the Late Jurassic to Early Cretaceous scenario of the Paraná Basin, being followed by the Early Cretaceous magmatic rocks of the Serra Geral Formation. The Serra Geral Magmatism is predominantly basaltic in nature. However, rocks of intermediate and acid composition have been observed. Taken together, the Botucatu and Serra Geral formations comprise the Gondwana III Supersequence, also known as São Bento Group. The Serra Geral Magmatism peaked between 137±0.7 Ma and 126.8±2.0 Ma (Turner *et al.*, 1994), and the remnant pile of lava observed at the surface at may reach a thickness of 2000 m.

The Upper Cretaceous Bauru Supersequence is an up to 280 m thick package of alluvial, fluvial and eolian sedimentary rocks that ended the depositional history of the Paraná Basin. It occupied the flexural sag originated by the load of the volcanic pile; the depocenter of the Bauru Supersequence being coincident with the region of maximum thickness of lava. It includes conglomerate deposits with clasts of various rock-types, including sandstone, siltstone, argillite and calcrete. It contains a Senonian fauna including *Chelonia*, *Crocodilia* and *Dinosauria* (Huene, 1939, *apud* Fernandes and Coimbra, 1996).

Chaco-Paraná Basin

The Chaco-Paraná Basin is situated in northeastern Argentina, being an elliptical-shaped basin with the main axis oriented ENE-WSW, and encompassing an area of about 500 000 km^2 (Fig. 2). Its surface expression appears as an almost featureless flat plain covered by Quaternary to Recent fluvial deposits. In its northeastern part is remarkable the presence of Lower Cretaceous basalt flows, a lithological unit that is fully developed in its Brazilian counterpart, the Paraná Basin. To the W, the Chaco-Paraná Basin is limited by the Pampean Ranges, a system of thrusted Precambrian basement rocks uplifted in the Neogene by the Andean Orogeny. To the N the basin is separated from the Tarija Basin of southeastern Bolívia and Paraguay by the Rincón-Caburé Arch.

Subsidence and sediment accumulation in the Chaco-Paraná Basin began during Ordovician(?)-Silurian times. It seems that a deep SW-NE striking depocenter named the Las Breñas Low represented the initial site of sedimentary deposition in the region. This depocenter occurs in the central portion of the basin (Pezzi and Mozetic, 1989), and appears as an asymmetric graben with structural dip of the basement to the NW, against its master bordering normal fault (Fig. 8). The nature and age of the rocks contained within the graben are unknown. However, seismic information indicate that these are probably both sedimentary and igneous, and may be up to 5000 m thick. In spite of being the nature and thickness of the rocks inside the Las Breñas depocenter still speculative topics, a reasonable hypotesis based on seismo-stratigraphic analysis is that of a basal sequence of immature siliciclastic sediments, covered by a mixed quartzose-carbonate marine beds and culminating in deeper marine shale units (Pezzi and Mozetic, 1989). The complete sequence is probably confined to the Early Paleozoic.

Milani (1997) and Milani and Ramos (1998) suggested that there may be a relationship between the Las Breñas Low and analogous features in the Brazilian Paraná Basin, situated to the NE, where the Lower Paleozoic depocenters are related to the beggining of the sedimentation and which retain the record of the first transgressive-regressive cycle, the Ordovician-Silurian Supersequence.

The Silurian-Devonian Supersequence of the Chaco-Paraná Basin (Fig. 3) surpassed the depositional area of the previous unit, and locally includes remnants of the Late Ordovician-Early Silurian glaciation in the Zapla Formation. The Zapla Formation is followed by a sequence of micaceous, dark grey shale units of Silurian age known as the Copo Formation. There then follows the classic Devonian succession found in the basins of southwestern Gondwana consisting of a basal section of continental to shallow marine sandstone beds known as Caburé Formation that grade into the marine black shale section of the Rincón Formation. The total thickness of the Devonian sequence exceeds 1200 m.

The Permian-Carboniferous Supersequence is up to

2300 m thick in the Chaco-Paraná Basin (Russo *et al.*, 1987), and occur in two distinct sub-basins. In the western, Alhuampa Sub-Basin the lowermost unit, the Stephanian-Asselian Sachayoj Formation, consists of medium-grained sandstone units interbedded with black shale, followed by diamictite. This sequence is up to 550 m thick and is known as the Charata Formation, the characteristics of which indicate a cold and humid, glacial to periglacial environment. A similar sedimentary succession appears in the eastern sub-basin, but there the lower stratigraphic unit, known as the Ordoñez Formation, consists of about 1500 m of white sandstone beds, and grey to black shale with associated diamictite beds. The upper unit consists of some 650 m sandstone, siltstone and shale, and is known as Victoriano Rodríguez Formation.

There followed the transitional Chacabuco Formation, composed of siltstone and shale, locally bituminous, and *lenticular beds of limestone. These deposits grade upwards* into a thick sequence of continental sandstone beds, known as the Buena Vista Formation. These sandstone units invaded the remnants of the Permian seas and closed the Paleozoic history of sedimentation in the Chaco-Paraná Basin. Such continental sedimentation persisted up to the Early Triassic.

During the Jurassic, the Chaco-Paraná Basin was the site of a broad crustal upwarp, reflecting the initial tectono-thermal events related to South Atlantic rifting. Fluvial and eolian sediments up to 400 m thick accumulated throughout the region and constitute the Tacuarembó Formation. There then followed the intrusion and extrusion of the Early Cretaceous diabase and basalt of the Serra Geral Formation. In the Paleocene, large parts of the Chaco-Paraná Basin were invaded by waters of the South Atlantic Ocean, resulting in the accumulation of some 350 m of grey calciferous mudstone and sandstone beds, with subordinated beds of gypsum, a sequence known as the Mariano Boedo Formation. From Miocene times onwards, a continental sequence of sands encroached on the Chaco-Paraná Basin from the W, constituting a wedge of poorly consolidated post-orogenic sediments derived from the Andean Belt, known as Chaco Group. These sediments exceed 2000 m in thickness.

Atlantic, Divergent Margin

Continental margin basins

The divergent margin of South America (Fig. 1), can be followed over a distance of about 10 500 km from the Orinoco Delta in easternmost Venezuela, to Tierra del Fuego, at the southern tip of Argentina. It is a large geological province that originated by the mechanisms that caused the break-up of Western Gondwana and the resulting drift of the African and South American plates since the Mesozoic, forming in this way, the South Atlantic Ocean and related marginal sedimentary basins.

Considering the nature and orientation of the regional stress fields during rifting and the following dynamics of the divergent motion between African and South American plates during drifting, three diverse domains are recognized along this margin. These are the dominantly extensional domains from southernmost Argentina up to the northeastern corner of Brazil. Secondly, a transform segment corresponding to the margin along the Equatorial Atlantic; and thirdly the region to the N of the mouth of the Amazonas River.

In the southern extensional domain, the structural pattern of the rift phase was marked by dominantly dip-slip normal faults. This style can be observed in each individual basin along the margin (Ojeda, 1982; Chang *et al.*, 1992; Urien and Zambrano, 1996). Transfer faults occur at high angles to the regional strike of normal faults. The set of normal faults evolved and defined the regional trend of the newly growing passive margin, and the former transfer faults nucleated large transform zones in the oceanic floor, the projection of which against the continental border intercepts it in almost at right angles.

The tectonic style in the Equatorial Atlantic Domain was different (Zalán, 1985; Szatmari *et al.*, 1987; Mascle *et al.*, 1988; Matos, 2000). Right lateral wrenching was the mechanism responsible for crustal rupture, creating a pattern of high-angle oblique faults that controlled rifting and resulted in the development of large-scale parallel-to-shore fracture zones (Gorini, 1977). These fracture zones include those of Fernando de Noronha, Chain, Romanche, and Saint Paul.

A third region of the divergent margin of South America, lies to the N of the mouth of the Amazonas River (Fig. 1). This constitutes part of another extensional domain, the Central Atlantic Ocean. This is the older segment of the extensional margin of the plate, in which an early phase of rifting occurred in the Triassic.

Individual sectors or "basins" along the continuous divergent margin of South America share some fundamental characteristics, including the classic rift, transitional and open marine tectonic and sedimentological patterns (Asmus and Ponte, 1973; Asmus, 1984). However, there are differences, too, and in some cases these are significant differences, justifying their consideration as individual geological provinces.

Rifting, and the subsequent opening of the Atlantic Ocean, was a remarkably diachronous event along the divergent margin of South America. This was of Late Triassic age in the northernmost sector of the Brazilian Margin (Brandão and Feijó, 1994a); Early Jurassic in southern Argentina (Baldi and Nevistic, 1996); Neocomian along the eastern Brazilian margin (Chang *et al.*, 1992), and Aptian along the Equatorial Atlantic (Matos and Waick, 1998), a range of time that spans about 90 Ma. The existing evidence suggests that there occurred an early opening along the Guiana and in the Caciporé-Foz do Amazonas basins, related to Central Atlantic rifting. In the Jurassic, crustal fracturing started to define the southern margin of Argentina, and this event propagated northwards to northeastern Brazil during the Neocomian. By the Aptian, the final delimitation of the South American and African plates was close to an end with the breaking along the Equatorial Atlantic. The Pernambuco-Paraíba Basin is recognized as the last sector of the divergent margin of South America to undergo rifting (Matos, 1998), as late as Late Aptian times (Feijó, 1994b).

Central Atlantic Domain

Rifting in the Central Atlantic region occurred during Late Triassic to Early Jurassic times, and led to the separation of the North American and African plates. The tectonic mechanisms that created a divergent plate margin in the region were influenced to large extent by the fabric of the basement, which was printed during Late Paleozoic orogenies (Manspeizer, 1988). The Variscan-Alleghenian orogenic belt formed by the coalescence of the Gondwana and Laurasian cratons (Rast, 1988). Along both margins of the Central Atlantic Ocean the presence of remnants of the Variscan-Alleghenian deformation belts is remarkable, including the Mauritanides Orogen in Western Africa and its American counterpart, the Appalachian Belt; to the S, those ancient belts can be projected towards the northernmost region of South America.

The Newark Supergroup fills one of the several Triassic to Early Jurassic rift basins related to the Atlantic margin of the North American and African plates (Manspeizer, 1988). This might be used as representative for correlation with the equivalent units in some of the northernmost Mesozoic basins of South America (Fig. 9). The Newark Supergroup is a sequence some up to 9000 m thick of lacustrine conglomerate, arkosic and lithic sandstone units, siltstone and red-brown mudstone. Syn-rift sedimentation in the Newark Basin began during Anisian to Carnian times and extended up to the Liassic.

The first incursion of the sea in the Central Atlantic Basin occurred in the Late Triassic, probably during the Carnian (Manspeizer, 1988), when Tethyan waters invaded westward through the Gibraltar Fracture Zone or southwards from eastern Greenland. By the Early Jurassic, close to the time of the post-rift unconformity, the evolving margins of the Central Atlantic were submitted to tectonic reactivation. Basins were uplifted and eroded, the sea retreated, and tholeiitic volcanism associated with lacustine strata followed (200 Ma to 175 Ma; Sutter, 1985, *apud* Manspeizer, 1988). This distinctive evolutive phase signaled the end of syn-rift times and the onset of sea-floor spreading in the Central Atlantic Ocean.

Guiana Basin

The Guiana Marginal Basin is situated in the northernmost part of the Atlantic margin of South America, encompassing the territorial waters of easternmost Venezuela, the Guiana, Surinam and French Guiana, with a total area of more than 400 000 km^2 (Fig. 9). The basin contains an estimated 10 000 m of Mesozoic to Cenozoic sediments (Krook, 1975), assigned to the Corantijn Group (Fig. 10) in the Surinam coastal region. The initial stages of crustal rifting in that area were marked by the intrusion of the Apatoe Dolerites, of Late Triassic age (227±10 Ma).

The sediments of the Corantijn Group were deposited during a succession of transgressive-regressive sedimentary cycles, starting with the sandy Nickerie Formation of Late Cretaceous age, followed by the Calcutta Formation (Paleocene to early Miocene) and the Tambaredjo Formation (late Miocene to Pleistocene) (Noorthoorn van de Kruijff, 1970, *apud* Krook, 1975). A significant stratigraphic hiatus occurs between the Upper Cretaceous and the Paleocene, in the upper Eocene-lower Miocene interval, and in the mid-lower Miocene. Holocene clays constitute the Demerara Formation, mostly supplied by the Amazonas River and the Caribbean oceanic current.

To the NE of the continental shelf of Surinam, a particular feature exists (Fig. 9); in that area, the size of the shelf is much larger than its mean size, defining the Demerara or Guiana Marginal Plateau (Colette *et al.*, 1971, *apud* Krook, 1975). The Demerara Plateau is an almost undisturbed depositional feature with an asymmetric profile, being steep-sided to the N and NE and gentle to the W. The oldest rock already sampled from the Demerara Plateau is a sandstone of Late Jurassic age.

Caciporé-Foz do Amazonas Basin

The Caciporé-Foz do Amazonas Basin covers about 270 000 km^2 of the northernmost continental margin of Brazil, adjacent to the mouth of the Amazonas River and along the coast of the State of Amapá (Fig. 9). The offshore basin is delimited by a NNW-SSE-trending dyke swarm of Triassic age that cuts through the outcropping area of the Guiana Shield and dates the initial rifting in that region, thus related to the opening of the Central Atlantic.

The first rift phase in the Caciporé-Foz do Amazonas Basin (Fig. 10) is represented by the Calçoene Formation, a sequence of tholeiitic volcanic rocks with radiometric ages ranging from 186 to 222 Ma (Brandão and Feijó, 1994a). The volcanic rocks are intercalated with sandy red beds of alluvial, fluvial and lacustrine sediments. This rift sequence, that has an estimated thickness exceeding 1000 m, was tilted inside the half grabens where it is confined, displaying an angular unconformity with the overlying sediments. With respect to rifting in the Central Atlantic region, this sequence occupies a similar tectonic setting and chronostratigraphic position of that of the Newark Supergroup, previously described.

A second and even more important phase of rifting in the area occurred during the Valanginian-Middle Albian interval, when the Caciporé Formation, consisting of up to 7000 m of syn-tectonic grey shale with intercalated sandstone beds, locally with volcanic rocks (125 Ma) accumulated in deep grabens (Fig. 11). This formation is followed unconformably by transgressive shale and prograding marginal marine beds of sandstone, known as the Limoeiro Formation, which accumulated between the Late Albian and the early Paleocene, and marks the initial stages of the post-rift phase. The upper Paleocene to middle Miocene sequence is represented by the Marajó Formation. This formation is up to 2000 m thick and consists of coarse-grained marginal marine sandstone beds that interfinger seawards with the limestone units of the Amapá Formation. The Amapá Formation consists of a great variety of carbonate rocks, and grades towards the continental slope into the shaly facies of the Travosas Formation.

The post-middle Miocene record in the Caciporé-Foz do Amazonas Basin is represented by the siliciclastic rocks of the Pará Group, including the coarse-grained sandstone beds of the Tucunaré Formation, the fine-grained sandstone and siltstone units of the Pirarucu Formation, and the deep-water shales and turbidites of the Orange Formation. The

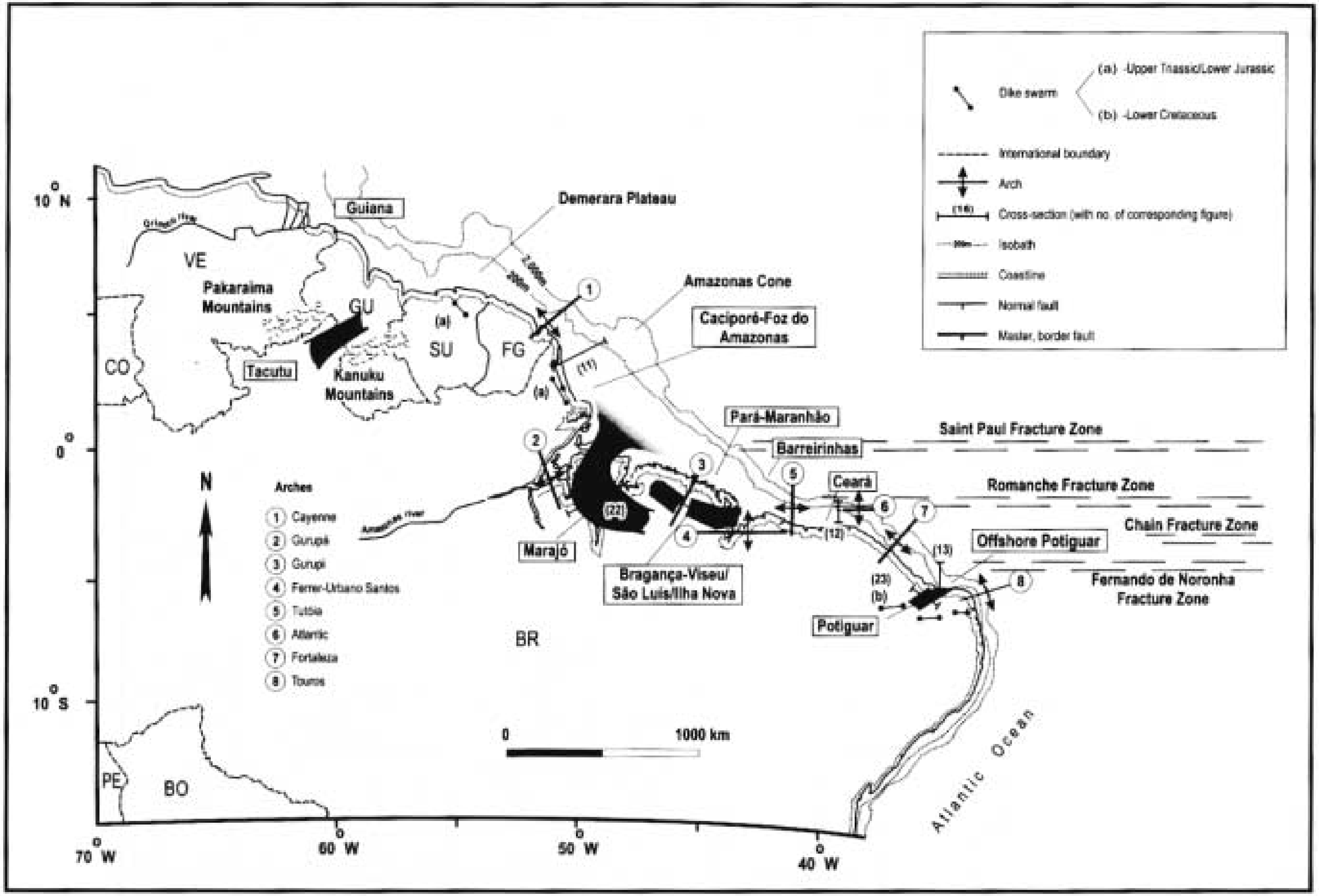


FIGURE 9 - The Central Atlantic and Equatorial segments of the divergent margin of South America (compiled from various sources). Countries: see Fig. 2.

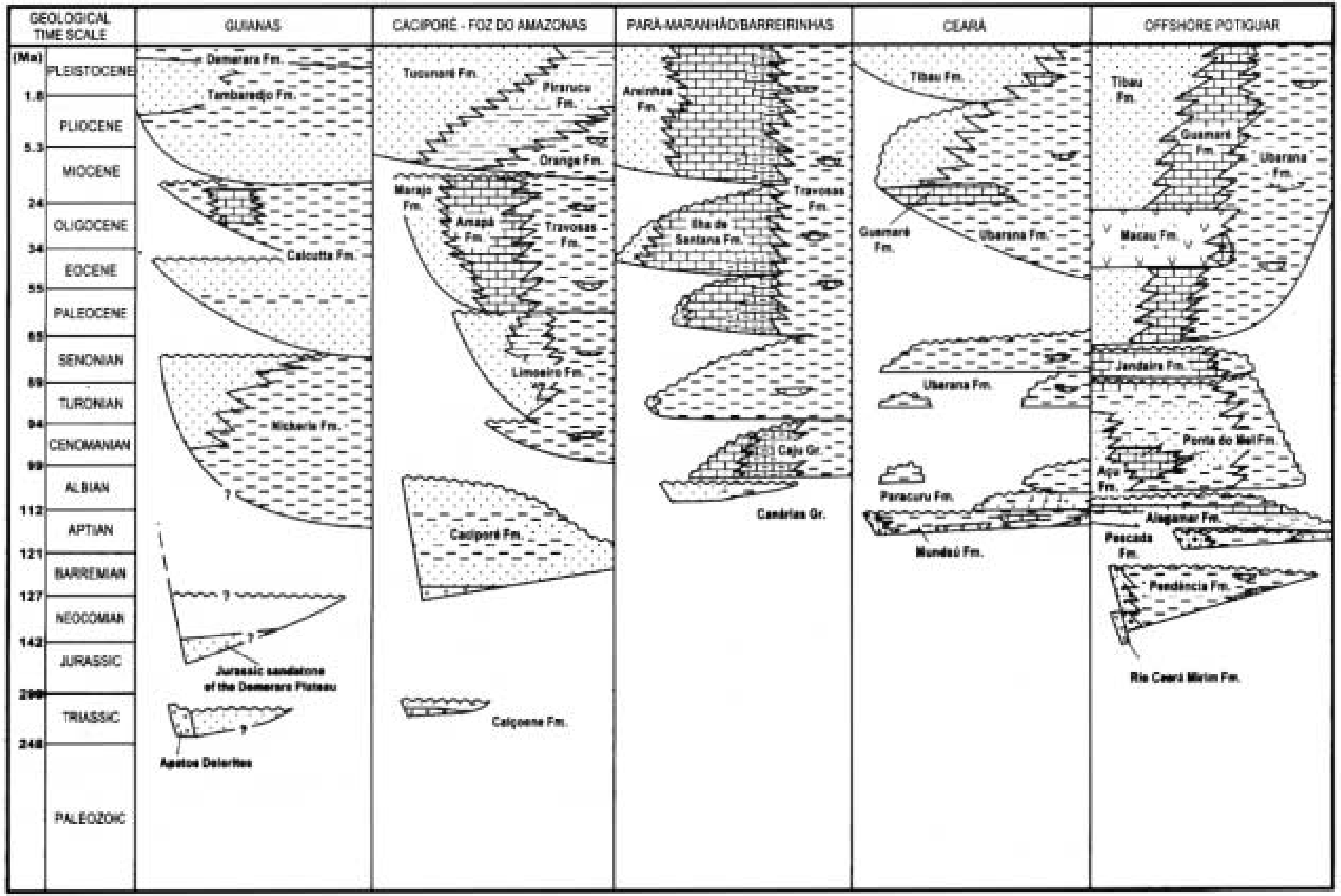


FIGURE 10 - Stratigraphic summary of the Central Atlantic and Equatorial Atlantic marginal basins of South America. On each display, continent is at the left side. Chart for the Guiana Basin is mostly conceptual, and corresponds to its coastal domain. Chart for the Ceará Basin is more representative of its central portion, the Acaraú-Icaraí Sub-Basin; to the west, in the Piauí-Camocim Sub-Basin, the Late Cretaceous to Early Cenozoic lacuna is even greater (Late Cenomanian to Late Eocene hiatus), and to the east, in the Mundaú Sub-Basin, it is of almost imperceptible amplitude. Main sources of information were Krook (1975), Guiana Basin; Brandão and Feijó (1994a), Caciporé-Foz do Amazonas Basin; Brandão and Feijó (1994b), Feijó (1994), Pará-Maranhão/ Barreirinhas basins; Beltrami et al. (1994), Ceará Basin; Araripe and Feijó (1994), Offshore Potiguar Basin. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.

Orange Formation includes a peculiar feature: the Amazonas Cone. This gigantic deep-water fan is several-kilometres thick, and consists of partially unconsolidated shale deposited by gravity-controlled bottom flows and intercalated turbidite bodies with clay to gravel-sized particles (Damuth, 1976). Its characteristic feature is the adiastrophic structures like diapirs and growth faults (Castro *et al.*, 1978, *apud* Brandão and Feijó, 1994a).

Equatorial Atlantic Domain

Equatorial Atlantic transform tectonics imposed particular characteristics on the dynamics of rifting and the final geometry of each of the related marginal basins. The conventional architecture of rift basins, in which subsidence is controlled by a master border fault and where the basement tilts towards it, with fans of coarse clastic sediments developing adjacent to the faulted margin, is not a standard pattern that can be easily recognized along the Equatorial Atlantic marginal basins (Matos and Waick, 1998).

Specific patterns of transtensional rifting along a continental-scale transform plate boundary led to strongly segmented development along the Equatorial Atlantic (Fig. 9), with significant differences in the thermal history, facies distribution, magmatism, uplift and deformational styles from one basin to another. Some regions of the Equatorial Atlantic margin experienced almost continuous subsidence, resembling that of the purely extensional Eastern Brazilian margin, whereas other sectors underwent strong episodes of inversion and erosion that created large lacunas in their stratigraphic record.

The available data set indicates that rifting along the Equatorial Atlantic margin occurred between the Aptian and the Cenomanian. Matos and Waick (1998) suggested an almost instantaneous, widespread Aptian rift phase that fractured the entire Equatorial margin by a right-lateral strike-slip motion, a contrasting mechanism with respect to the longer, Late Jurassic to Barremian rift stage commonly observed along the extensional margin of the South Atlantic. In such a particular tectonic context of rapid overall crustal rupture and mechanical subsidence, marine incursion was facilitated and at some places influenced sedimentation already during the rift phase.

Pará-Maranhão/Barreirinhas basins

The Pará-Maranhão Basin (Fig. 9) is situated in the offshore region of the homonymous states of northern Brazil, covering an area of about 50 000 km². The Barreirinhas Basin is the eastward continuation of the former, covering an area of around 20 000 km² in the offshore regions of the State of Maranhão, a small part of this being on land. The striking characteristic in the stratigraphic record of this part of the Equatorial Atlantic margin (Fig. 10) is the persistent development of a carbonate platform, known as the Ilha de Santana Formation, ranging in age from Maastrichtian to Recent.

The Barreirinhas Basin is developed on Paleozoic to Neocomian siliciclastic and magmatic rocks that correspond to the stratigraphic sequences of the adjacent Parnaíba Interior Basin (Feijó, 1994a). These sequences are the Serra Grande Group (Silurian), the Canindé Group (Devonian), the Balsas Group (Carboniferous to Triassic), the Mearim Group (Jurassic), and Sardinha Volcanics (Neocomian). Rifting in this part of the margin occurred during the Early and Middle Albian (Brandão and Feijó, 1994b), and is represented by a sequence of immature lithic sandstone beds, siltstone and green shale, comprising the Canárias Group. These beds are interpreted as deltaic facies that were deposited in a prematurely marine rift basin. Shallow marine conditions dominated during the Late Albian, with the deposition of the Caju Group. The Caju Group consists of a basal sequence of quartzose sandstone units named the Periá Formation followed by bioclastic limestone units, the Bonfim Formation, and shales of the Preguiças Formation.

The open marine section, from Cenomanian time onwards, is represented by the Humberto de Campos Group, a classic coast-platform-slope-deep basin depositional complex. The proximal facies are the quartzose sandstone beds of the Areinhas Formation that interfinger with the high-energy carbonates of the Ilha de Santana Formation. Deeper water facies include shale and turbiditic sandstone units known as the Travosas Formation.

Ceará Basin

This basin is situated in the offshore part of the State of Ceará (Fig. 9), and covers about 35 000 km². Due to prominent differences in the tectonic and stratigraphic styles, three distinct domains are recognized in the Ceará Basin. From W to E these are the Piauí-Camocim, Acaraú-Icaraí and Mundaú sub-basins. The sub-basins are separated by important positive transversal features, corresponding to basement highs, large volcanic bodies or huge anticlines produced by syn-sedimentary structural inversion (Zalán, 1985). Compressional faults and folds were first described in the Piauí-Camocim Sub-Basin and in the adjacent Barreirinhas Basin by Miura and Barbosa (1972), at a time when the presence of compressional tectonics in the context of Atlantic-type marginal basins was considered anomalous and often contested.

Rifting occurred during Early to Middle Aptian times in the Ceará Basin (Beltrami *et al.*, 1994). This stage accommodated a sequence of more than 4000 m of continental sedimentary rocks, corresponding to the alluvial, fluvial and lacustrine facies of the Mundaú Formation (Fig. 10). In the Late Aptian, the transitional sequence of the Paracuru Formation accumulated, to form up to 500 m of sandstone, chalk, coarse-grained limestone and black shale in marginal marine to shallow marine environments. The top of this section is defined in the Ceará Basin by an important regional unconformity (Fig. 12).

From the Albian time onwards, a thick slope shaly facies started to develop in the entire Ceará Basin, corresponding to the Ubarana Formation, with subordinate beds of sandstone and limestone. However, the kinematics of this transform margin left important differences in the stratigraphic record preserved in its sub-basins, particularly that deposited in Late Cretaceous to early Cenozoic times. In the Piauí-Camocim sub-basin, an important lacuna was recorded spanning the interval between the Late Cenomanian-late Eocene, due to a strong episode of basin inversion. At this time, the Atlantic high, an E-W oriented positive structure related to a large, deep-seated transpressive wrench fault

(Zalán, 1985), was originated. The corresponding hiatus is of lesser extent but still important in the adjacent Acaraú-Icaraí Sub-Basin, and almost imperceptible in the easternmost, NW-SE-trending Mundaú Sub-Basin.

From the Eocene-Oligocene to Recent, coarse-grained clastic sediments accumulated in the proximal areas of a fully developed depositional system situated at the continent margin. These are the beds of the Tibau Formation that grade into the carbonate rocks of the Guamaré Formation deposited in the shallow waters and to shale beds in the more distal areas, the latter being known as Ubarana Formation. Another aspect observed in the Ceará Basin that certainly reflects the presence of an oceanic fracture zones cutting its floor, is the abundance of guyots and seamounts in the deep-water regions. Volcanic rocks also occur as flows and intrusions, affecting all stratigraphic levels.

Offshore Potiguar Basin

The Potiguar Basin is situated in the easternmost part of the Brazilian equatorial margin, in the northeastern corner of Brazil (Fig. 9), occupying some regions of the states of Ceará and Rio Grande do Norte. It includes an offshore segment having an area of about 27 000 km^2 and an onshore segment that covers 22 000 km^2; the latter being discussed in the section on Failed Rift Basins.

The Potiguar Basin underwent a complex evolution, merging elements from both the Equatorial and the Southern Atlantic tectonic zones. Rifting started in the Late Berriasian as result of a WNW-ESE regional extensional stress field (Cremonini *et al.*, 1998) that persisted up to the Early Barremian. During this stage, a sequence of shale beds, rich in organic matter, and sandstone units, including alluvial, fluvial, deltaic and lacustrine facies, named Pendência Formation (Fig. 10), filled fault-bounded depressions (Fig. 13). A second rift phase caused by E-W extension and only in the offshore regions of the basin was active during the Late Barremian to Early Aptian, and expanded the deformation to the W. During this phase the Pescada Formation was deposited. This formation consists of alluvial-fluvial sandstone beds with associated siltstone and shale units. Basic magmatism of the Rio Ceará Mirim Formation constituted a persistent association to rifting in the domain of the Potiguar Basin.

The transitional-to-marine cycle is represented by the Alagamar Formation, of Late Aptian age (Araripe and Feijó, 1994), fully developed in the offshore domain of the basin, where it contains about 600 m of sediments. This formation is composed of a basal sandy sequence, known as the Upanema Member; an intermediate section of black shale and associated limestone beds, informally named the Ponta do Tubarão Beds; and an upper sequence of black shales, included in the Galinhos Member. From Albian time onwards, sedimentation assumed the standard marginal basin configuration of proximal sandy facies, corresponding to the Açu and Tibau formations; carbonate units included in the Ponta do Mel, Jandaíra and Guamaré formations; and a distal shaly facies known as the Ubarana Formation.

In the Late Campanian, right-lateral strike-slip deformation affected the Potiguar Basin, creating a WNW-ESE-trending pattern of structures and causing the development of a regional unconformity of significance in the offshore part of the basin (Cremonini *et al.*, 1998). In the early Cenozoic, subaquous olivine-basaltic magmatism known as Macau Formation took place. Final tectonic events in the region of the Potiguar Basin are represented by Oligocene to Recent, E-W compression released along pre-existing faults trending NE-SW, making this one of the most seismically actives regions of Brazil.

Southern Atlantic Domain

Late Jurassic time marked the beginning of rifting in the southernmost part of the South Atlantic (Uliana and Biddle, 1988; Urien and Zambrano, 1996). However, since the Late Triassic to Early Jurassic, a series of geological phenomena linked to the main phase of break-up, but of more localized occurrence, preceded the overall rupture. In southern Argentina, two active phases of rifting are recognized as preceding the final, Late Jurassic rupture: the Chon Aike Phase (Late Triassic-Early Jurassic) and the Lonco Trapial-Aguada Bandera Phase (Middle Jurassic) (Keeley and Light, 1993). The remarkable 180 Ma complete isotope resetting of the Paleozoic sedimentary rocks of the Paraná Basin (Thomaz Filho *et al.*, 1976), indicating a time of anomalously high heat flow, was another phenomenon related to this Jurassic, early phase of rifting. The consequent elevated continental freeboard just before rifting may explain the dominance of continental sedimentation in the Jurassic, pre-rift sequences. The presence of magmatism and the earliest rifting in southernmost Argentina was the precursor to the imminent widespread break-up.

The pre-existing structural grain of Gondwana, including Precambrian and Paleozoic to Middle Triassic trends, played an important role during Mesozoic rifting. Large sectors of the South Atlantic rupture were accommodated as normal faults upon a much older fabric. However, the existence of prominent transversal structures created complications to this simple picture of propagating rifts, as was the case of the Gondwanides (Urien and Zambrano, 1996), in which the southern margin of the paleocontinent evolved as a collisional belt facing the Panthalassa during the Late Paleozoic.

The southernmost part of the eastern continental margin of South America that developed over the Gondwanides domain, is marked by the presence of a large offshore region of continental crust. This is a complex geological domain striking E-W known as the Malvinas Plateau that extends for more than 2000 km from the shoreline of southern Argentina (Fig. 14). The western, cratonward side of this transversal feature merges both morphologically and structurally with the normal-sized passive margin of the continent (Biddle *et al.*, 1996). To the E and SE, far away from the shoreline, the plateau developed a rifted transitional margin to oceanic crust. Important accumulation of sedimentary rocks, up to 7000 m thick, occurred on the southeastern flank of the plateau in the Malvinas Oriental Basin (Kress *et al.*, 1996).

The northern border of the Malvinas Plateau coincides with the important E-W trending Malvinas (Falklands) fracture zone, along which continental and oceanic crust abruptly keep contact and define the Malvinas (Falklands) Escarpment. This submarine escarpment is almost 3000 m

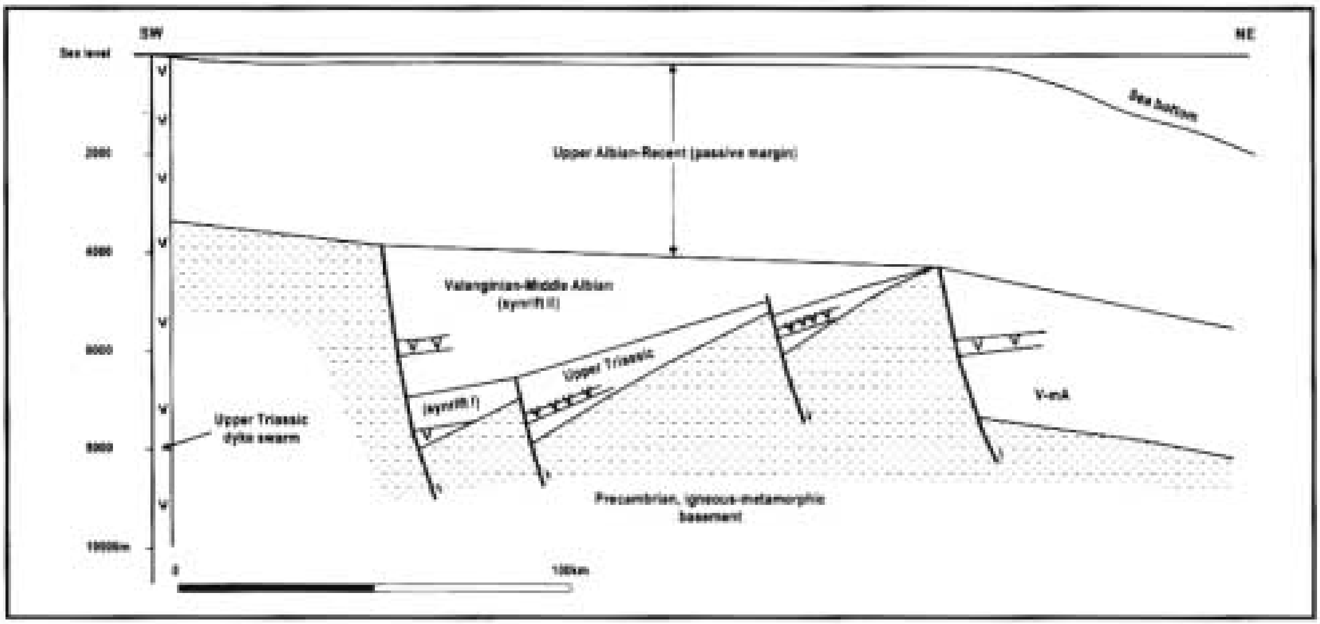


FIGURE 11 - Conceptual cross-section of the Caciporé/Foz do Amazonas Basin (based on information from Brandão and Feijó, 1994a).

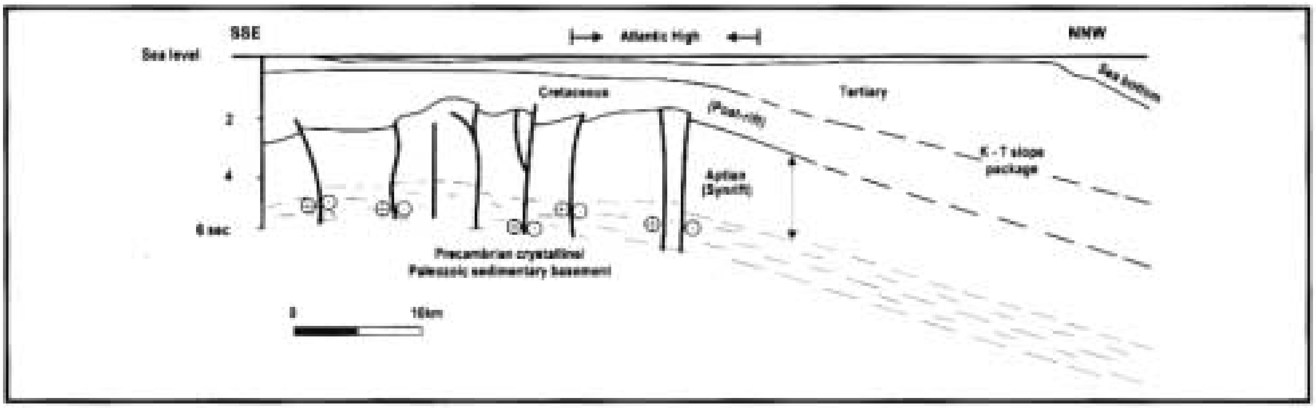


FIGURE 12 - Geoseismic cross-section of the Piauí-Camocim Sub-Basin of the Ceará Basin (after Zalán, 1985; permission to reprint granted by the author).

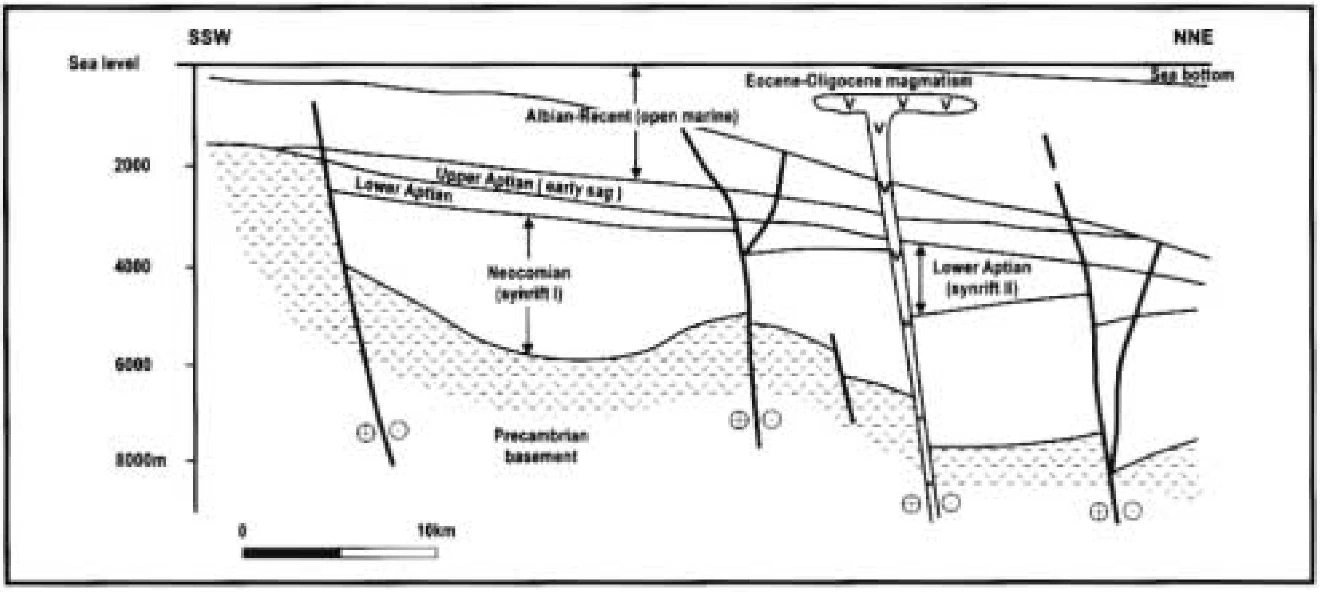


FIGURE 13 - Cross-section of the Offshore Potiguar Basin (after Cremonini et al., 1998; permission to reprint granted by the authors).

high, and the North Scotia Ridge, the active transform boundary between South American and Scotia plates, marks the southern limit of this large continental salient.

Developed on the Malvinas Plateau are a Mesozoic to Recent sedimentary package estimated to have a maximum thickness of about 4000 m (Biddle *et al.*, 1996). The western part of the plateau is marked by the Malvinas Islands, a remote archipelago where there occurs a complete succession of Paleozoic sediments overlying Precambrian basement, the Cape Meredith Complex dated at 1.124 Ga (Nullo *et al.*, 1996). The Paleozoic sequence includes elements that support its correlation with those of Gondwana in South America and Africa, showing the classic succession of a lower unit, the Gran Malvina Group of Silurian-Devonian age, and an upper section, the Isla Soledad Group of Late Carboniferous-Permian age. Between these two groups of rocks there appears a regional unconformity that corresponds to a major break, encompassing a lacuna as large as the Mississipian. The Paleozoic sediments were intruded by Early Jurassic diabase dykes (180 Ma to 192 Ma; Musset and Taylor, 1994, *apud* Nullo *et al.*, 1996).

To the N of the Malvinas Plateau, from the southern Argentinean margin, up to the Santos Basin in southeastern Brazil, the continental margin of South America is about 3000 km long and strikes SW-NE (Fig. 14). At the northeastern end of this straight part of the margin lies the São Paulo Ridge in the deep water realm, and the Florianópolis Platform in the shallower area. The continuation of these features towards the continent coincides with the Ponta Grossa Arch, a prominent NW-SE trending magmatic rift where abundant Early Cretaceous tholeiitic intrusions (128 Ma to 134 Ma; Turner *et al.*, 1994), affecting the Paleozoic successions of the Paraná Basin and surrounding basement, define the largest dyke swarm of the continent.

The São Paulo Ridge also marks the southern limit of the São Paulo Plateau, where the extended crust of the continental margin widens considerably compared to adjacent regions, reaching more than 400 km of width between the 2000 m and 3000 m isobaths (Mascle and Renard, 1976), and holding the Santos and Campos basins. On its seaward side, the São Paulo Plateau is limited by a steep scarp that marks also the eastern border of the area of salt occurrence in these marginal basins. To the NW, the São Paulo Plateau is delimited by the Serra do Mar, another prominent feature along the divergent margin of South America. The present-day configuration of these ranges is given by a chain of scarps, about 1000 km long and up to 2200 m high. Its regional NE-SW trend follows the structural grain of the Precambrian rocks of the Atlantic Shield (Almeida and Hasui, 1984). The origin and development of this uncommonly high topographic feature along the eastern margin of the continent was interpreted by Almeida and Carneiro (1998) as the result of Paleocene to Recent scarp retreat upon the uplifted block of the Santos faultline, a large synthetic normal fault that was progressively buried and today lies offshore.

The divergent margin of South America, to the N of the São Paulo Plateau, extends N-S for about 1200 km. In its northern extremity, this sector reaches the Salvador Deflection (Szatmari *et al.*, 1984), the point where the marginal rift bifurcates into an inland branch, the Recôncavo-Tucano-Jatobá Rift System, and a coastal branch that extends along the length of the Jacuípe Basin, Sergipe-Alagoas Basin, and Pernambuco-Paraíba Basin to its northernmost limit, the Touros Platform, in the northeastern corner of Brazil.

Pernambuco-Paraíba Basin

The northernmost segment of the occidental, extensional margin of the South American continent is known as the Pernambuco-Paraíba Basin. It lies mainly offshore and the total area is about 35 000 km^2, of which some 9000 km^2 lies on land (Fig. 14). With respect to the entire eastern margin, this is the region where rifting occurred last, the rupture having been prevented up to Early Aptian times by the nature and high rigidity of the cratonic rocks that constitute the Precambrian basement (Matos, 1998). The sedimentary accumulation in the offshore areas of the basin has not yet been drilled, and its nature and stratigraphy are inferred from seismic data.

Rifting occurred very quickly in this region and lasted for about 5 Ma (Matos, 1998) during the Late Aptian. A sequence at least 3000 m thick of conglomerate and sandstone units, the Cabo Formation (Fig. 15), was deposited. These sedimentary rocks are associated with alkaline and calc-alkaline volcanic rocks, named the Ipojuca Formation, and filled fault-bounded depocenters during this phase. There followed a sag phase, when the Albian to Cenomanian carbonates of the Estiva Formation were deposited in a shallow marine basin (Feijó, 1994b). Senonian times witnessed the encroachment of the marginal coarse-grained clastic sediments of the Beberibe Formation, and the fine-grained clastics of the Itamaracá Formation upon the shallow marine basin. During the Latest Cretaceous to Recent, carbonate units developed in the shallow marginal basin, a sequence that is included in the Gramame and Maria Farinha formations, that grade seawards into the shale beds of the Calumbi Formation.

An unusual aspect of the geology of the Pernambuco-Paraíba Basin is the existence of good exposures of the K/T boundary, not found elsewhere along the continental margin of Brazil. This interval was studied by Albertão (1995). Here the K/T boundary corresponds to a zone situated about 60 cm above the contact between the Gramame Formation, of Maastrichtian age, and the Maria Farinha Formation of Paleocene to Oligocene age. This contact is expressed as an erosive surface separating a lower sequence of biomicritic marlstone beds of neritic environment from a sequence of limestone units and shale beds accumulated under deeper neritic conditions. These show hummocky cross stratification, interpreted as the result of the action of a tsunami. A bed of marlstone that shows anomalously high values for iridium and total organic carbon marks the K/T boundary itself. Here, as elsewhere, worlwide, we see a remarkable level of extinction of pre-existing biota that may be due to the impact of an extraterrestrial body (Alvarez *et al.*, 1980, *apud* Albertão, 1995).

Sergipe-Alagoas Basin

The Sergipe-Alagoas Basin is situated in northeastern Brazil, and covers an area of about 35 000 km^2, two thirds of which lies offshore (Fig. 14). Along the entire eastern margin

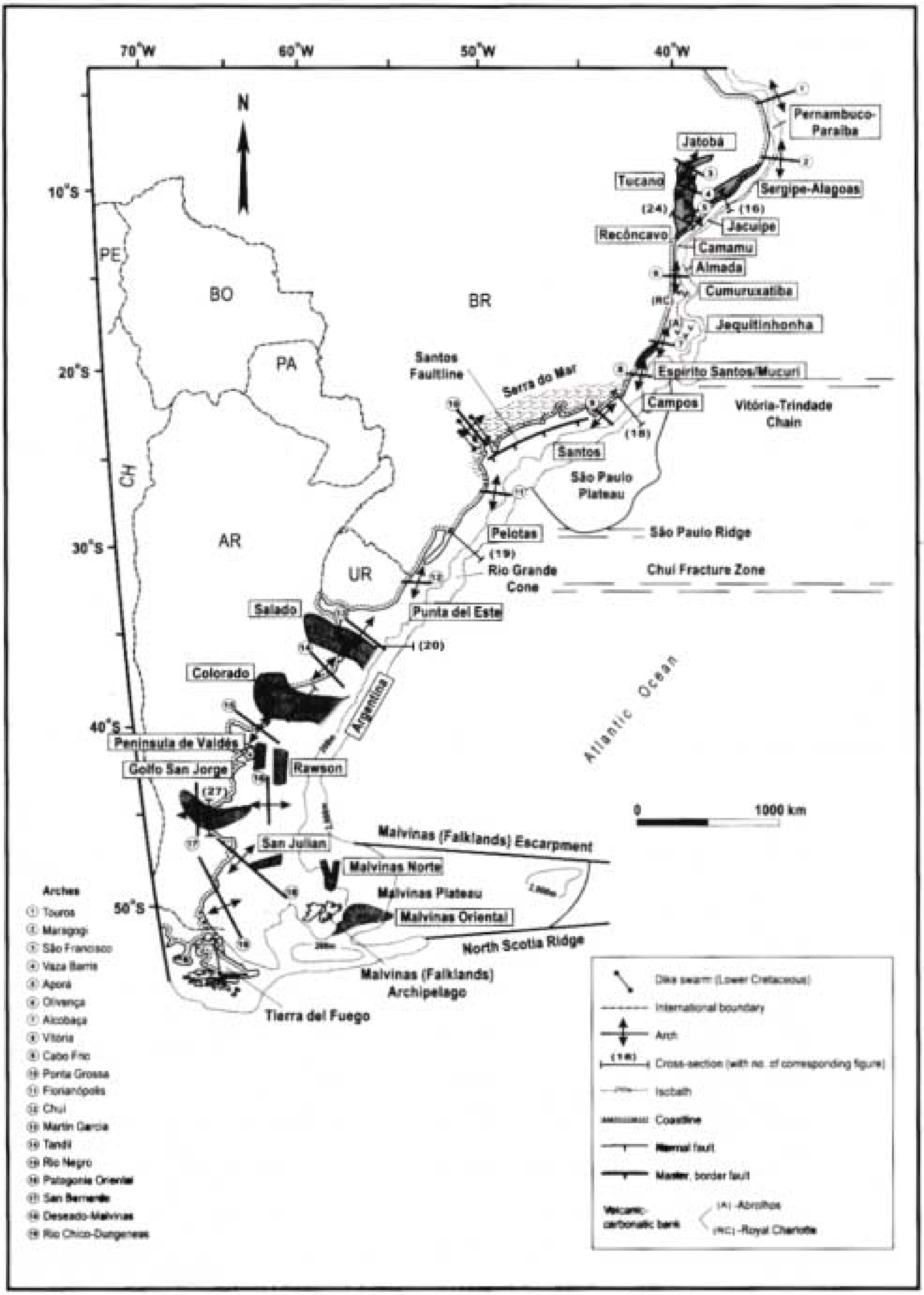
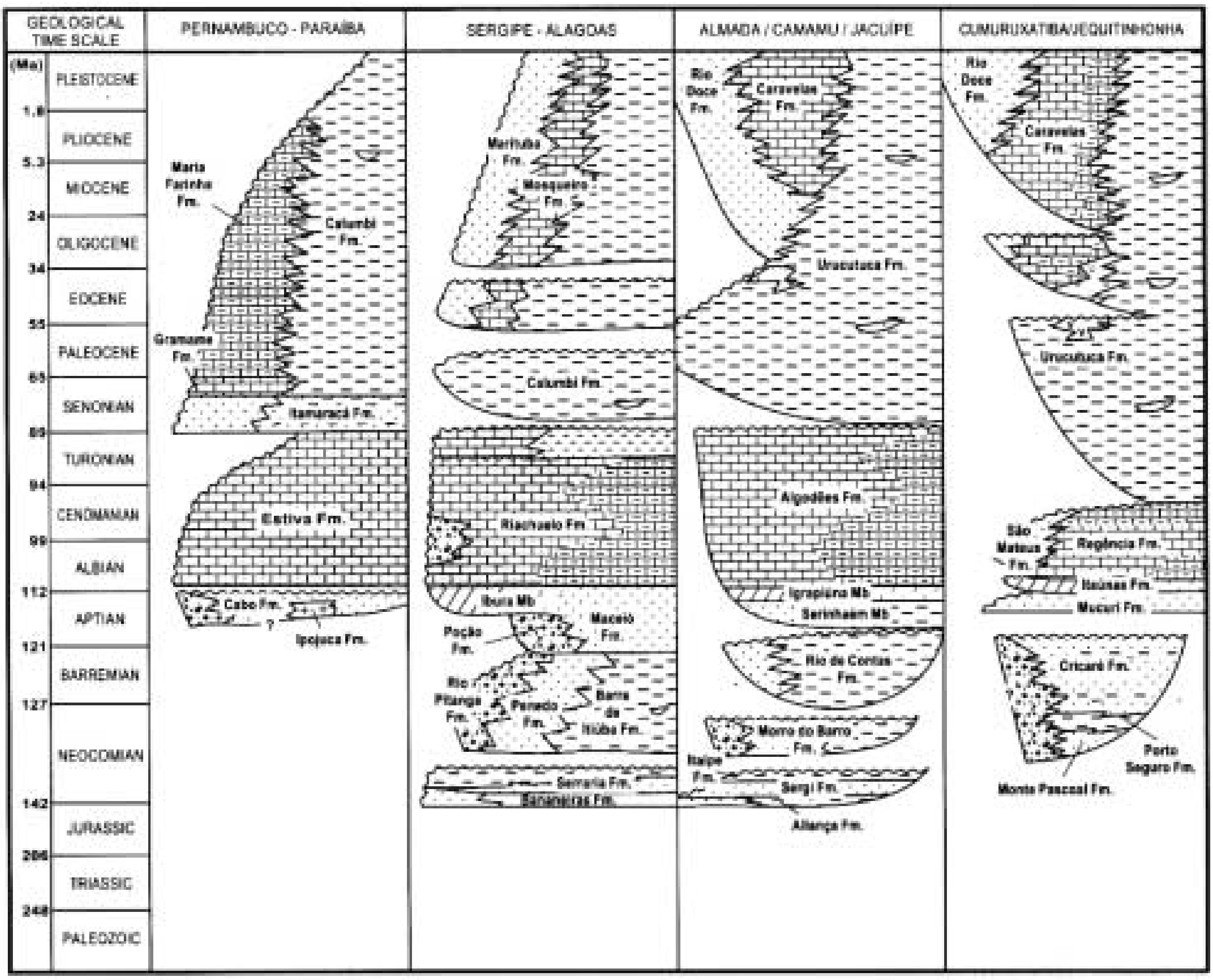


FIGURE 14 - The Southern Atlantic segment of the divergent margin of South America (compiled from various sources). Countries: see Fig. 2.



of Brazil, the Sergipe-Alagoas Basin has the most complete stratigraphic record, including remnants of Late Proterozoic and Paleozoic sedimentary successions, a fully developed Jurassic to Early Cretaceous pre-rift sequence, and the classic Mesozoic-Cenozoic rift and post-rift sequences (Fig. 15).

Northeastern Brazil was the site of extensive sedimentation during the Paleozoic and probably during the Late Proterozoic as well, a fact that is supported by the countless occurrences of these strata throughout the region. These beds are invariably preserved from erosion by local subsidence due to Mesozoic extensional tectonics. The Estância Formation is the oldest of these rock units. It consists of an undeformed, very low-metamorphic grade succession of conglomerate, sandstone, shale, limestone and dolomite beds of Late Proterozoic to Cambrian age that are exposed in the southern part of the basin.

Paleozoic remnants in the Sergipe-Alagoas Basin are included in the Igreja Nova Group (Feijó, 1994c). The Batinga Formation, of Late Carboniferous age, consists of diamictite beds, coarse-grained sandstone, and laminated siltstone deposited under glaciomarine conditions. The Aracaré Formation consists of red sandstone, shale, limestone and silexite of Early Permian age.

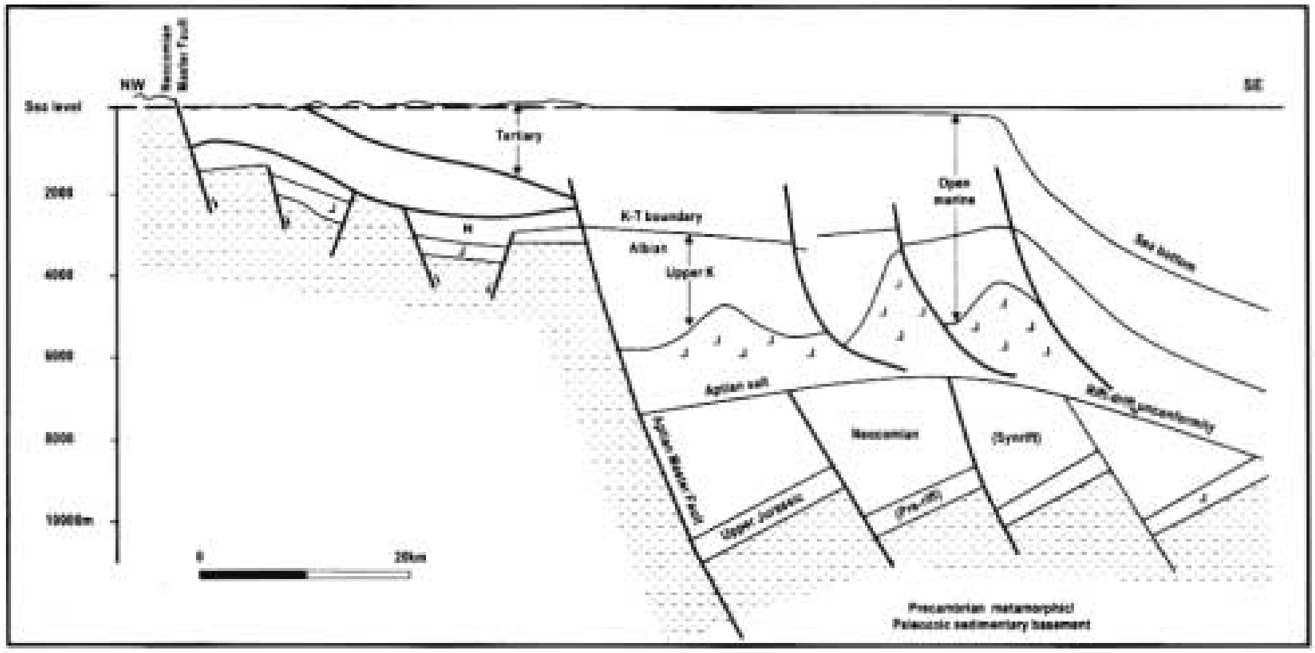
The Late Jurassic to Early Cretaceous pre-rift sequence of the Sergipe-Alagoas Basin corresponds to the Perucaba Group (Feijó, 1994b). It consists of fine-grained red sandstone beds at the base, the Candeeiro Formation; followed by red shale in the Bananeiras Formation; and medium to coarse-grained white and reddish sandstone beds with exuberant cross-bedding, the Serraria Formation.

In the Sergipe-Alagoas region the initial crustal rupture was in response to transtensional forces (Milani *et al.*, 1988; Milani and Szatmari, 1998; Szatmari and Milani, 1999), and rifting along this part of the margin was a process completely devoid of magmatism. A set of *en échelon* N-S trending grabens developed along the regional, SW-NE strike of the nascent margin, and were filled by the Neocomian continental sedimentary rocks represented by the lower part of the Coruripe Group. This group includes conglomerate accumulated as alluvial fans along the faulted margin, known as the Rio Pitanga Formation; the fluvial sandstone of the Penedo Formation; alternating shale and sandstone beds of deltaic origin, of the Coqueiro Seco Formation; and lacustrine shale of the Barra de Itiúba Formation.

A recurrent rift phase took place during the Early Aptian, promoted by a NW-SE extensional stress field. A large NE-SW trending fault-line running approximately along the present day shoreline and separating the onshore part of the basin from that offshore thus originated (Fig. 16). The onshore part of the basin displays a thin sedimentary record whereas the sequence that occurs offshore is known to attain 10 000 m. The second rift phase

FIGURE 15 - Stratigraphic summary of the Southern Atlantic marginal basins of South America, eastern and northeastern Brazilian sectors. Chart for the Sergipe-Alagoas Basin is more representative of its southern domain, the Sergipe Sub-Basin; and the Almada/Camamu/Jacuipe chart is more representative for the Camamu sector. On each display, continent is at the left side. Main souces of information were Santos et al. (1994), Cumuruxatiba/Jequitinhonha basins; Netto et al. (1994), Almada/Camamu/Jacuipe basins; Feijó (1994c), Sergipe-Alagoas basin; Feijó (1994b), Pernambuco-Paraíba Basin. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.

FIGURE 16 - Cross-section of the Sergipe-Alagoas Basin (after Cainelli and Mohriak, 1998; permission to reprint granted by the authors).



is marked by the conglomerate beds of the Poção Formation in the Alagoas Sub-Basin and by the lower part of the Maceió Formation in the Sergipe Sub-Basin. A sag phase followed, and Late Aptian evaporite beds that constitute the Ibura Member of the upper part of the Maceió Formation filled the depositional space.

Albian carbonates, the Riachuelo Formation, are fully developed in the Sergipe Sub-Basin, and grade to a sandy-conglomeratic proximal facies known as Angico Member. The Upper Cretaceous to Recent continental margin sequence is named the Piaçabuçu Group, and corresponds to the sandstone beds of the Marituba Formation, the carbonate units of the Mosqueiro Formation and the shale of the Calumbi Formation. According to interpretations of seismic data, there occur locally some magmatic bodies of probably Late Cretaceous to early Cenozoic age in the deep water realm of the Sergipe-Alagoas Basin, constituting seamounts and intrusions into the sedimentary rocks (Santos and Castro, 1992, *apud* Feijó, 1994c).

Almada/Camamu/Jacuipe basins

These basins correspond to the offshore parts of the continental margin of Brazil that is situated adjacent to the Todos os Santos Bay, in the State of Bahia covering an area of about 110 000 km^2, with a small fraction of this on land (Fig. 14). The Todos os Santos Bay corresponds to the southernmost termination of the aborted Recôncavo-Tucano-Jatobá Rift System, and the region where this interior rift links with the continental margin. This important tectonic junction, situated right at the mouth of the Todos os Santos Bay, is marked by a complex tectonic pattern. During the evolution of the northern part of the Camamu Basin, both compressional and extensional regimes were active, producing an intrincate framework with a mix of structural styles.

Some intervals in the stratigraphic succession of these basins show similarities with equivalent sections in adjacent basins, from where they had also obtained their stratigraphic names (Fig. 15). This is the case of the Paleozoic Afligidos Formation and the pre-rift Aliança and Sergi formations, classical units of the interior Recôncavo Basin. Above the sandstone units of the Sergi Formation lies a section of up to 500 m of thick grey and brown shale beds of lowermost Neocomian age known as the Itaípe Formation, representing the transition between the pre-rift stage and the rift stage (Netto *et al.*, 1994).

The syn-rift lacustrine basin was filled by shale named the Jiribatuba Member that grades into proximal coarse-grained sandstone, the Tinharé Member. Together these beds constitute the Morro do Barro Formation. The Barremian to Early Aptian times correspond to a sequence of mixed clastic and carbonate sedimentary rocks known

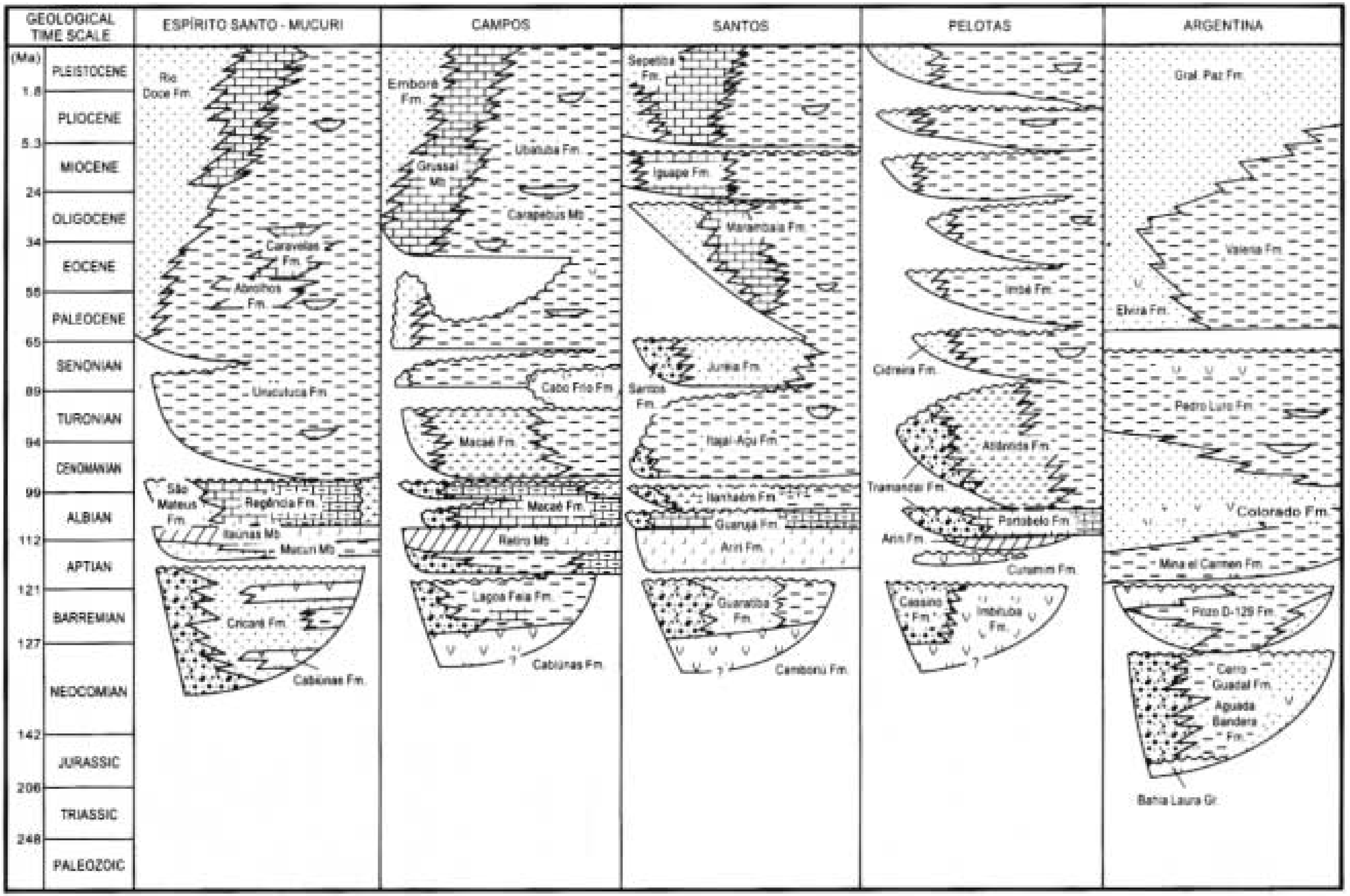


FIGURE 17 - Stratigraphic summary of the Southern Atlantic marginal basins of South America, from Argentina to southeastern Brazil. On each display, continent is at the left side. Chart for the Argentina Basin is a compilation of data from the various sub-basins of that domain. Main souces of information were Keeley and Light (1993), Urien and Zambrano (1996), Argentina Basin; Dias et al. (1994), Pelotas Basin; Pereira and Feijó (1994), Santos Basin; Rangel et al. (1994), Campos Basin; Vieira et al. (1994), Espírito Santo-Mucuri Basin. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.

as Rio de Contas Formation. These consist of grey and brown shale beds and fine-grained to conglomeratic sandstone that accumulated under deltaic-lacustrine conditions. In the Jacuípe Basin the Neocomian sequence is absent, and the rift sequence is represented by the Rio de Contas Formation, in that area spanning the range Barremian-Early Albian.

A section consisting of evaporite beds and associated clastic sedimentary rocks of Late Aptian age define the Taipus-Mirim Formation. This formation consists of very fine-grained sandstone intercalated with siltstone and black shale rich in organic mater, a sequence known as the Serinhaém Member. The Igrapiúna Member consists of carbonate (including dolomite), brown shale, halite and anhydrite. Locally, the anhydrite is replaced by a layer of barite in some regions of the Camamu and Almada basins. These rocks are the result of the first marine transgression into the basin that developed under arid conditions (Netto *et al.*, 1994). The classic Albian-Cenomanian carbonate sequence is known as the Algodões Formation. It is followed by the Late Cretaceous and Cenozoic sequence that in the Almada/Camamu/Jacuípe basins is lithologically similar and holds the same stratigraphic nomenclature of the equivalent rocks existent in the Espírito Santo-Mucuri Basin, that is Urucutuca, Caravelas and Rio Doce formations.

Cumuruxatiba/Jequitinhonha basins

These dominantly offshore basins correspond to a stretch of the Eastern Brazilian Margin that strikes approximately N-S and is about 350 km long (Fig. 14). These basins occupy an area of about 45 000 km^2 along the coast of the State of Bahia. The basement in the area consists of Precambrian granitic-gneissic terrains of the São Francisco Craton (Almeida and Hasui, 1984). The structural framework of the rift phase in these basins is defined by N-S and SW-NE-trending dominantly synthetic normal faults, locally interrupted by NW-SE-oriented transfer faults and accommodation zones.

Neocomian to Early Aptian sedimentation in these basins (Fig. 15) corresponds to the Cumuruxatiba Group and the Cricaré Formation. The former includes a basal sequence of clastic sedimentary rocks, the Monte Pascoal Formation, consisting of coarse to medium-grained sandstone beds intercalated with dark grey to black shale and conglomerate that may represent the pre-rift section. The syn-rift package is represented by the lacustrine shale of the Porto Seguro Formation (Santos *et al.*, 1994) and the Cricaré Formation. The Cricaré Formation, as well as the rest of the stratigraphic record in the Cumuruxatiba/Jequitinhonha basins, including the early Cenozoic magmatism of the Abrolhos Formation, exhibit similar lithological characteristics to the equivalent section found in the adjacent Espírito Santo-Mucuri Basin, also borrowing from that basin the names of the formal stratigraphic units.

Espírito Santo-Mucuri Basin

This basin is situated in the coastal (about 20 000 km^2) and the offshore regions (200 000 km^2) of the State of Espírito Santo and the southernmost part of the State of Bahia (Fig. 14). The southern part of the basin, in the deep water realm, is connected to the Campos Basin, but is separated from the adjacent basin over the continental platform and onshore domains by a positive feature at the level of the Precambrian basement, known as the Vitória High. To the N, the Mucuri Basin displays the marginal sag successions similar to that seen in basins along the eastern coast of Brazil.

The Espírito Santo-Mucuri Basin evolved over a complex association of Precambrian igneous and metamorphic rocks (Vieira *et al.*, 1994). The oldest indications of rifting in the region are NW-SE trending Jurassic dykes, known as the Fundão Suite. This early magmatic event was followed by a Neocomian-Barremian episode of extrusion of tholeiitic basalt and volcaniclastic beds during the rift phase known as the Cabiúnas Formation, where the magmatic bodies locally interfinger with the sediments.

Syn-rift sedimentation of the Cricaré Formation in the Espírito Santo-Mucuri Basin is estimated at 5000 m of Neocomian to Early Aptian continental conglomerate, sandstone, coquina, calcilutite and dolomite beds, with associated black shale with organic matter (Fig. 17). The sequence accumulated according to a pattern of N-S to NE-SW trending, dominantly synthetic normal fault-bounded blocks, and is terminated at the top by an important erosional unconformity, separating this lowermost unit from the transitional beds of the Mariricu Formation. The Mariricu Formation is represented by Late Aptian conglomerate, feldspathic coarse-grained sandstone and shale, the Mucuri Member, that grades upwards into a sequence of evaporite beds and black shales known as the Itaúnas Member.

The open marine sediments in the Espírito Santo-Mucuri Basin form the Barra Nova Group, of Albian age, and the Espírito Santo Group, spanning the Cenomanian to the Recent. The former consists of coarse-grained sandstone of marginal marine facies, the São Mateus Formation that grades seawards into the limestone units of the Regência Formation. The Espírito Santo Group is the classic sequence that defines the prograding continental platform, constituted by a proximal sandy facies known as Rio Doce Formation interfingering with carbonate beds, the Caravelas Formation; and becoming shaly at the slope and deep basin, the realm of the Urucutuca Formation.

Cenozoic times were marked in the Espírito Santo-Mucuri Basin by the important Abrolhos magmatic event that imposed profound structural rearrangements to the basin. The Abrolhos Formation (Vieira *et al.*, 1994) corresponds to the sub-alcaline to alcaline volcanic and volcaniclastic rocks that were extruded during late Paleocene-Eocene times. The magmatism peaked between 60 Ma and 40 Ma, and the resulting bodies are intercalated with the limestone units of the Caravelas Formation and with the shale beds of the Urucutuca Formation. The outpouring of large volumes of magma in the off-shore part of the basin added local complications to the conventional gravity-driven salt tectonics; reaching such a down-dip obstacle, halokinesis developed a distinctive pattern of compressional structures around the Abrolhos Complex.

Campos Basin

The Campos Basin is situated in the territorial waters of the State of Rio de Janeiro, southeastern Brasil. The basin covers an area of almost 100 000 km²; of which, only about 500 km² are onshore (Fig. 14). To the N, it is partially isolated from the Espírito Santo Basin onshore, and from the shallow water domain, by the Vitória High, an elevated block of Precambrian basement that coincides with the westward prolongation of the Vitória-Trindade Chain in the Atlantic seafloor. Crystalline basement has scarcely been sampled by drilling in the basin, and corresponds to the same Precambrian gneissic domain that crops out in the surrounding areas. To the S, the Campos Basin is limited with the Santos Basin along the Cabo Frio Arch, an area of persistently active magmatism during the post-break-up evolution of the margin. In that domain, Turonian to Campanian sediments are intercalated with volcaniclastic deposits, flood basalts and diabase sills of 80 Ma to 90 Ma. Recurrent magmatism formed volcanic cones during Eocene times (Mohriak *et al.*, 1995).

Rift sediments of the lower part of the Lagoa Feia Formation in the Campos Basin span the time interval Late Neocomian-Barremian, and overlie and interfinger with the basalt and volcaniclastic beds of the Cabiúnas Formation, dated at 120 Ma to 130 Ma (Dias *et al.*, 1990). The lower part of the Lagoa Feia Formation (Fig. 17) includes conglomerate with abundant clasts of basalt that form large fans along major border faults of tilted blocks, coarse to fine-grained sandstone, shale rich in organic material, siltstone and coquina, defining a complete sequence of lacustrine sedimentation. The coquinas are up to 50 m thick bioclastic accumulations of pelecypod shells, named the Coqueiros Member, constituting sedimentary rocks associated with structural syn-depositional highs, and representing a particular facies of porous rocks in this basin (Rangel *et al.*, 1994). The upper part of the Lagoa Feia Formation, lying above an expressive unconformity, is represented by a sequence of Aptian conglomerate beds and shaly red beds that grade upward into a section of Late Aptian evaporites known as the Retiro Member, with rhythmic intercalations of anhydrite and halite. The occurrence of the Aptian and older sediments is limited by a SW-NE trending synthetic, basement-involved fault-line nearly parallel and close to the coast (Fig. 18).

During Albian to Cenomanian times, marine conditions prevailed. The Macaé Formation consists of clastic and oolitic, coarse to medium-grained limestone beds, the Quissamã Member. Locally, these limestone beds are completely dolomitized. The vertical succession includes beds of calcilutite, marlstone and shale, a package known as Outeiro Member. The sandy turbiditic bodies that occur at some places are known as the Namorado Sandstone. In the proximal part of the basin, the Macaé Formation consists of conglomerate and poorly sorted sandstone beds that form the Goitacás Member. These beds grade basinward into the others facies of the Albian-Cenomanian sequence. The Campos Group unconformably overlies the Macaé Formation.

The Campos Group represents the post-Cenomanian filling-in of the marginal sag during the last thermal phase of subsidence and seaward tilting of the continental margin. It consists of two laterally interfingering units. The proximal is the conglomeratic, sandy and carbonate-bearing Emborê Formation; and the distal is the shaly Ubatuba Formation. The Ubatuba Formation consists of thousands of metres of shale and marlstone, with intercalated sandy turbiditic bodies, collectively known as Carapebus Member.

The post-rift tilting of the basin to the E also triggered adiastrophic structures caused by the voluminous salt flow (Demercian *et al.*, 1993). Salt tectonics and the resulting structural styles in the Campos Basin configure two particular cases: close to the coastline, an upper extensional, 100 to 200 km wide domain caused by horizontal extension above the Aptian salt layer; and a lower compressional domain, in the deep water regions, originated by downdip contraction of the thin-skinned, gliding package. Listric normal faults, including a basal synthetic shear for the entire rafted section, are features overall associated with salt tectonics.

Santos Basin

The Santos Basin is a large, NE-SW trending concave coastal basin that occupies about 200 000 km² of the continental margin of southeastern Brazil (Fig. 14). To the N, the basin is limited by the Cabo Frio Arch, and to the S by the Florianópolis Platform, both features lying as the westward prolongation of ocean fracture zones (Cainelli and Mohriak, 1998) and related to important magmatic activity associated with sedimentation. The Santos Basin overlies the São Paulo Plateau that along the Brazilian eastern margin is a unique feature of anomalous highly extended continental lithosphere.

To the W, the basin is limited by a belt of coastal ranges up to 2200 m high where Precambrian rocks crop out, the Serra do Mar, a physiographical feature that confines the marginal sag to the offshore domain. The border of the pre-Aptian package is marked by a synthetic normal fault, the trace of which runs parallel to the shoreline about 50 km seaward. This feature remained active up to Late Cretaceous times, and was onlapped by the Cenozoic sediments during the thermal subsidence phase. The maximum thickness of the Neocomian to Recent sediments is estimated at about 11 000 m.

Rifting in the Santos Basin occurred in the Neocomian-Early Aptian, when lacustrine sediments accumulated (Fig. 17). The rift sequence, named the Guaratiba Formation and hardly sampled in the basin, includes proximal reddish coarse siliciclastic wedges with fragments of basalt and quartz, and coquina deposited over, and probably interfingering with, the Lower Cretaceous basalt of the Camboriú Formation. Anoxic lacustrine deposits including shaly beds, rich in organic material, are supposed to exist in the deep-water region of the basin (Pereira and Feijó, 1994). The rift sequence is covered by the Aptian evaporites of the Ariri Formation, consisting of intercalations of anhydrite and halite. The downdip flow of the evaporites during the late evolution of the basin provided abundant halokinetic structures in the deep-water region of the Santos Basin. Large salt diapirs and walls are observed, attaining several kilometres in hight.

The Albian succession, known as Florianópolis, Guarujá and Itanhaém formations, consist of, respectively, fine-to coarse-grained red sandstone, carbonate and grey shale, a suite of rocks deposited under open marine conditions during progressive deepening of the basin. Maximum flooding conditions were reached during Cenomanian-

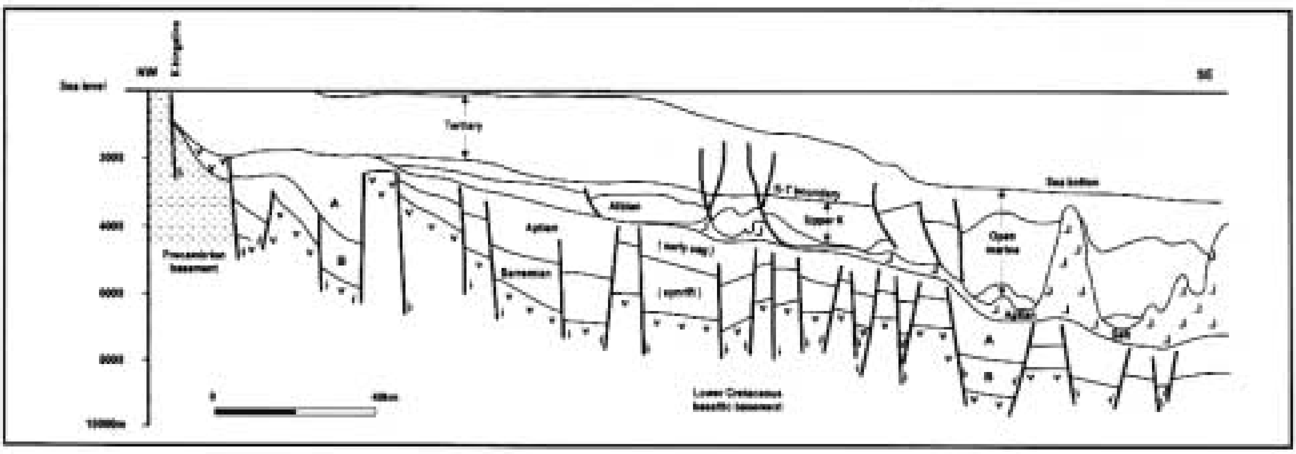


FIGURE 18 - Cross-section of the Campos Basin (after Cainelli and Mohriak, 1998; permission to reprint granted by the authors).

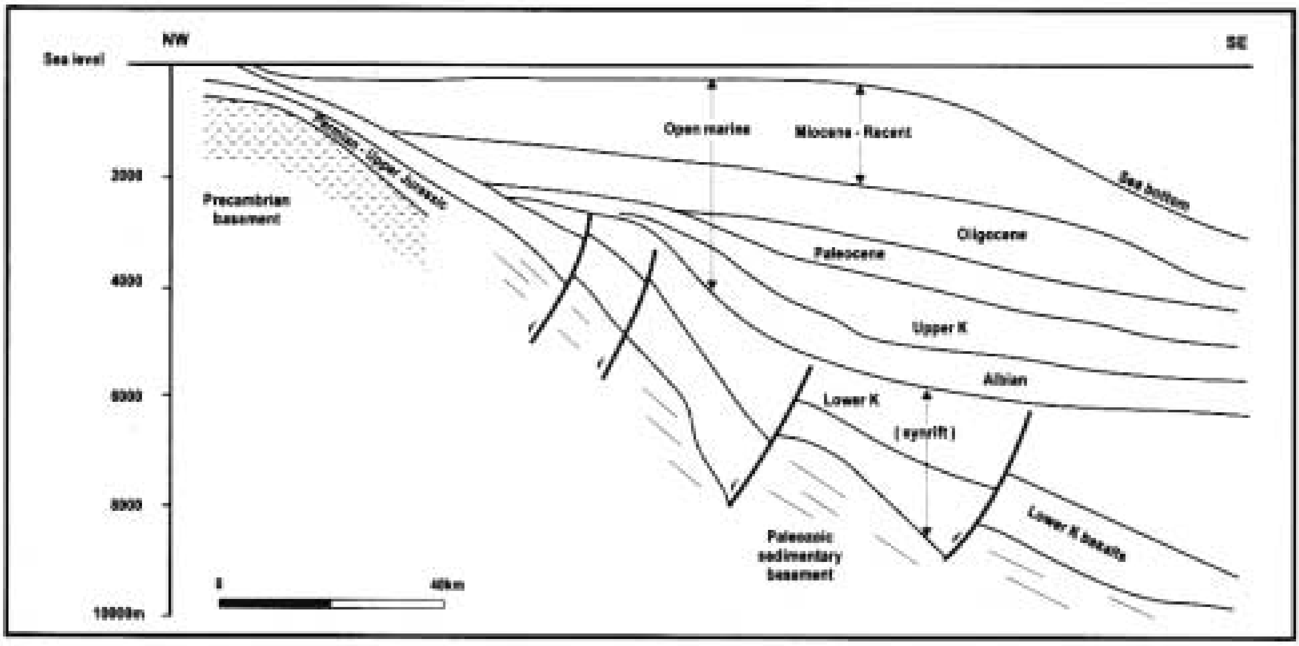


FIGURE 19 - Cross-section of the Pelotas Basin (after Dias et al., 1994, in Cainelli and Mohriak, 1998; permission to reprint granted by the authors).

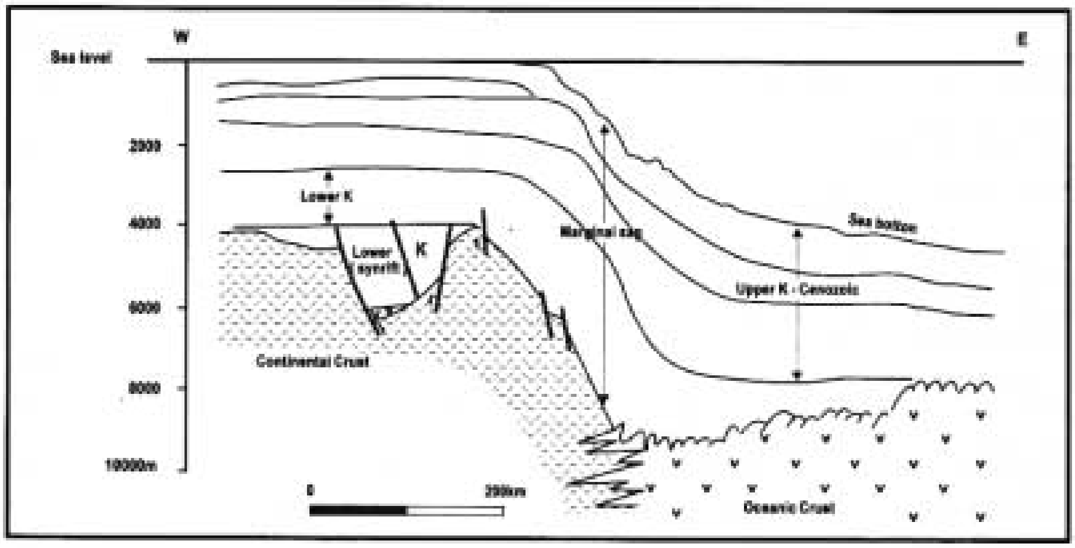


FIGURE 20 - Cross-section of the northern portion of the Argentina Basin (after Urien and Zambrano, 1996; permission to reproduce granted by Asociación Geológica Argentina).

Turonian times, when the dark grey shales of the Itajaí-Açu Formation were deposited. In the proximal regions, a thick wedge of conglomerate beds, the Santos Formation, and shallow marine sandstone beds, the Juréia Formation, spanning the Santonian-Maastrichtian interval, encroached on the basin in response to the uplift of the coastal Serra do Mar. This event is dated at 100 Ma to 80 Ma by apatite fission track analysis (Lelarge, 1993). Further evolution of the Santos Basin, during the Cenozoic, was marked by a basinward siliciclastic progradation, corresponding to the deposits of the Iguape and Sepetiba formations, against a large shaly platform/slope system included in the Marambaia Formation.

Pelotas Basin

The Pelotas Basin (Fig. 14) extends for about 160 000 km^2 in the offshore region of the southernmost Brazilian state, the State of Rio Grande do Sul. This basin also underlies about 40 000 km^2 of the coastal region of that province. In the onshore part, the basin is filled by a sequence of Cenozoic sandy siliciclastics sedimentary rocks as thick as 1800 m (Dias *et al.*, 1994). The post-Neocomian rocks that fill the Pelotas Basin directly overlie the crystalline Precambrian basement or over Paleozoic sequences that correspond to those of the Paraná Basin. The southern limit of the basin is the Chuy Fracture Zone, situated close to the international boundary with the Uruguay. To the N the Florianópolis Platform separates the Pelotas Basin from the Santos Basin.

Rifting occurred during Neocomian times (Fig. 17), when the basaltic flows of the Imbituba Formation and the coarse-grained siliciclastic rocks of the Cassino Formation were deposited. In the shallow-water region, where the rift structure is imaged by seismic data (Fig. 19), the half-grabens that define the rift blocks are controlled by high-angle antithetic faults (Cainelli and Mohriak, 1998). The basalt of the Imbituba Formation, with a characteristic array of seaward dipping wedges (Fontana, 1990), is 124±8.6 Ma old (K/Ar method, Dias *et al.*, 1994), an age that is close to the time when the voluminous intracontinental Serra Geral magmatism was terminating (126.6±2.9 Ma, Ar/Ar method, Milani, 1997; 126.8±2.0 Ma, Ar/Ar method, Turner *et al.*, 1994). This relationship seems to indicate a major shift in the volcanic centre from the interior of the plate towards its nascent margin.

The record of the continental-to-marine, evaporitic transitional phase is known only in the Florianópolis Platform, where there occurs a sequence, up to 50 m thick, consisting of halite, anhydrite, and carbonate beds included in the Ariri Formation, of Late Aptian age. These beds overlie the Curumim Formation, in which there occur trachy-andesites of Aptian age (113.2±0.1 Ma, Ar/Ar method, Dias *et al.*, 1994). The Florianópolis Platform also defines the southern limit of the great Aptian saline basin of the Southern Atlantic (Dias, 1998). The absence of evaporite beds gives the Pelotas Basin a monotonous, unfaulted and almost undeformed profile of a gently tilted monocline for its post-Aptian sequence, in sharp contrast with the standard configuration observed along the rest of the eastern Brazilian marginal basins.

From the Albian to the Recent, thermal subsidence and an essentially transgresive pattern of sedimentation dominated. The Early Albian is represented by the carbonate sequence of the Portobelo Formation that is covered by Late Albian to Recent proximal sandstone beds known as Tramandaí and Cidreira formations. These units grade to siltstone and fine-grained limestone of the Atlântida Formation, and to shale of the Imbé Formation towards the deep parts of the basin. The entire supra-Portobelo sequence is cut by a series of erosional surfaces that correspond to times of major regression. In the southern part of the Pelotas Basin, there occurs a sequence of post-Paleocene sediments up to 6000 m thick that define the Rio Grande Cone, a region with abundant structures related to gravitational movement of shale and that holds significant occurrences of gas hydrates in its upper levels (Fontana, 1989).

Argentina Basin

The Argentina continental margin basin is a 2300 km long, NE-SW trending straight segment of the South American eastern border (Fig. 14). It has an average width of 350 km to the 200 m isobath, and has a total area of about 800 000 km^2. Towards the N, in the Uruguaian continental waters, this segment of the margin is named Punta del Este Basin. Local depocenters, corresponding to Jurassic to Early Cretaceous aborted rift basins, and the intervening platform areas, all were draped by a sequence of Late Cretaceous to Cenozoic marine sedimentary rocks that may reach a thickness of 4000 m (Fig. 17).

The large offshore marginal sag of Argentina is divided into three domains (Ramos, 1996): the northern domain is characterized by the predominance of a transverse-to-shore structural pattern inherited from the Mesozoic rifting from the NW-SE pre-existing fabric in the area. In this northern sector the aborted rift structures of the Salado and Colorado basins developed, indenting cratonward from the margin (Fig. 20). Contrasting with the above, normal faults run parallel to the continental margin and define the framework of the rift phase in the central domain of the Argentina Basin, that includes the Rawson, Península Valdés, San Julian, and North Malvinas grabens. The southernmost domain is defined by the presence of the transverse structural feature of the Malvinas (Falklands) Islands Platform and its eastward prolongation, the Malvinas Plateau. This is a a large province of submarine continental crust bordered to the N by the prominent scarp of the Malvinas (Falklands) Fault Zone and to the S by the transform zone of the North Scotia Ridge.

The onset of extensional conditions and fault-driven subsidence in this area occurred during the Early Jurassic and marked the beginning of the development of the South Atlantic (Fig. 17). Magmatism was important during the initial stages of rupture (for example, dykes of basic rocks peaking between 180 Ma and 192 Ma in the Malvinas (Falklands) Islands (Musset and Taylor, 1994, *apud* Nullo *et al.*, 1996), and heralding large-scale break-up of Gondwana. The framework of rifting was to a large extent controlled by the pre-existing structural grain, and the rift phase accommodated continental deposits of lacustrine facies, including black shale, sandstone and limestone units, locally intercalated with abundant volcanic and volcaniclastic rocks.

As early as the Valanginian, the creation of oceanic crust began in the area (McLachlan and McMillan, 1979, *apud*

Light *et al.*, 1993), and sedimentation developed under incipient marine conditions associated with a phase of thermally-driven subsidence. However, up to Middle Cretaceous times the occurrence of widespread marine sedimentation in the Argentina Basin was made difficult by the presence of structural positive elements (Urien and Zambrano, 1996). Aptian times were marked by anoxic conditions and the local development of sediments rich in organic matter (Dias, 1998). A final stratigraphic pattern was established with the invasion of the Atlantic, transgressive from Late Cretaceous times to the Neogene. This was followed by a regressive tendency, when deltaic systems spreaded all over the shelf (Keeley and Light, 1993), offlapping the continental margin and prograding over the slope sequences, with abundant submarine erosion associated.

Failed rift basins

Extensional processes related to South Atlantic rifting led to the development of a series of grabens striking towards the craton from the continental margin (Figs. 9 and 14); most of which are restricted to the onshore domain of the continent in a function of the present-day freeboard along the margin. These rift basins have variable sizes, from few thousand to several ten thousand square kilometres; shapes, most usually being crescent or linear; position with respect to the margin, from parallel to normal-to-shore; and nature of its floor correspondent to different structural provinces of the South American Platform, lying directly over Precambrian crystalline basement or over older cratonic sedimentary cover.

The control exerted upon their positioning and evolution by basement weakness zones was also a variable factor: some were strongly controlled by pre-existing grain and others created their own. Rifting ages observed in these grabens reflect the age of that sector of the continental margin. This varied enormously and covered a time span of about 90 Ma. It is late Late Triassic along the northernmost sector of the Brazilian Margin (Marajó Basin; Milani and Zalán, 1998) and Early Jurassic in southern Argentina (Golfo San Jorge Basin; Baldi and Nevistic, 1996); Neocomian along the eastern Brazilian Margin (Recôncavo-Tucano-Jatobá basins; Caixeta *et al.*, 1994); and Aptian along the Equatorial Atlantic (São Luís/Bragança-Viseu/Ilha Nova basins; Lima *et al.*, 1994).

Each of these rift basins shows an evolutionary picture with a particular degree of completeness with respect to the full development of the continental margin of South America. Some of these rift basins became inland, aborted grabens still during the Cretaceous, and only contain lacustrine rocks. Others were partially or completely involved in the overall post-rift subsidence of the continental margin, and over their syn-tectonic sequences were deposited transitional and marine sedimentary beds.

Tacutu Basin

The Tacutu Basin (Fig. 9) includes areas of the State of Roraima (4500 km^2), in northernmost Brazil, and of the neighbouring country of Guiana (7000 km^2), where it is named the North Savannas Rift Valley (Berrangé, 1977). It is a NE-SW trending rift system about 300 km long and 30 to 50 km wide. It contains a sedimentary-volcanic fill of Jurassic to Recent age with a thickness that surpasses 7000 m (Fig. 21). The surface expression of this trough is essentially flat and dominated by tropical savanna. The faulted margins of the rift define highlands, the Pakaraima Mountains in the N and the Kanuku Mountains in the S.

The basement of the Tacutu Basin consists of Precambrian rocks of the Guiana Shield. Rifting was to a large extent controlled by NE-SW structural lines, pervasive in the ancient basement (Eiras and Kinoshita, 1990). The origin of the Tacutu Graben can be credited to differential rotation within the Guiana Shield during the opening of the Central Atlantic. The basin is limited by normal faults on both sides, but with an asymmetric transversal profile since that larger downthrows can be observed along the master faults of the southern border. An inversion in the sense of dip of the basement occurs when reaching the Guianan sector, where the master fault shifts to the northwestern side. An important E-W transfer zone known as the North Savannas Arch defines the region where the sense of dip of the basement inverts.

An interpreted pre-rift sequence, consisting of up to 1200 m of Jurassic basalts, the Apoteri Formation, and subordinate lacustrine red siltstone of the Manari Formation, underlies the Tacutu Basin (Eiras *et al.*, 1994b). The basalt was extruded mainly under subaerial conditions, but the existence of some pillows also indicates local subaquous volcanism. K/Ar dating indicates ages between 180 Ma and 150 Ma (Eiras and Kinoshita, 1990), a range consistent with the time of the opening of the southern part of the Central Atlantic Ocean.

During the initial stages of its evolution, the rift depression in the Tacutu Basin was filled by some 950 m of Tithonian evaporite beds and intercalated grey to brown shale and sandstone known as Pirara Formation. These beds interfinger towards the margins with a coarse-grained sandy and conglomeratic clastic wedge. The bromine content of these halites suggests marine influence, probably some short-lived incursions of the adjacent Central Atlantic waters.

Most of the rift sequence is aged Berriasian to Middle Albian, and is mainly composed of a thick succession of redbeds. The basal unit, the Tacutu Formation, up to 2700 m thick, consists of lacustrine siltstone and subordinated sandstone. The upper unit, the Tucano Formation, consists of fluvial and deltaic sandstone with a maximum thickness of 2200 m. Both syn-rift units interfinger with conglomerate and coarse-grained sandstone derived from footwall erosion along the southeastern faulted margin.

During the Cenozoic, the Tacutu Basin was submitted to a vigorous phase of inversion and structural rearrangement (Eiras and Kinoshita, 1988). The compressional stress field originated by the convergent motion between the South American Plate and the oceanic plates of Cocos, Nazca and the young-lived Caribbean Plate was released in the Tacutu area by transpressional motion along the major bounding faults of the graben. This gave rise to a series of positive features, including reverse faults, large anticlines and prominent flower structures. From the Pliocene onwards, a renewed phase of subsidence has affected the Tacutu Basin (Cordani *et al.*, 1984), that is

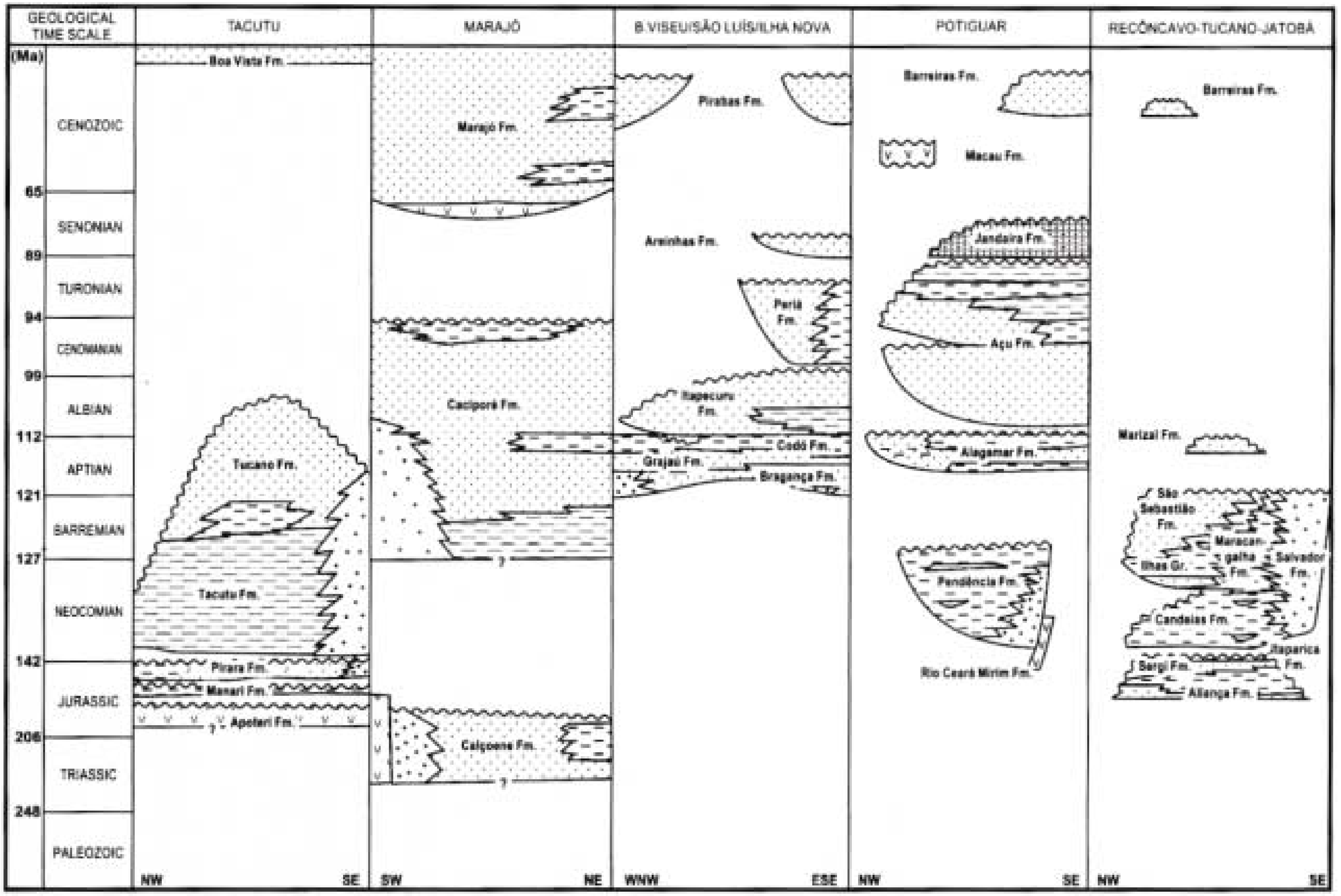


FIGURE 21 - Stratigraphic summary of the failed rift basins in the Brazilian sector of the divergent margin of South America. Chart for the Recôncavo-Tucano-Jatobá basins is representative of the southernmost portion of this rift system, the Recôncavo Basin. Chart for the Marajó Basins is mostly conceptual. Main sources of information were Eiras et al. (1994b), Tacutu Basin; Lima et al. (1994), Bragança-Viseu/São Luís/Ilha Nova Basins; Araripe and Feijó (1994), onshore Potiguar Basin; Caixeta et al. (1994), Recôncavo-Tucano-Jatobá basins. Geological time scale simplified after Gradstein and Ogg (1996). Lithological representations are of common use.

nowadays accommodating the fluvial-lacustrine system of the Boa Vista Formation.

Marajó Basin

The Marajó Basin is situated in the northern-equatorial region of Brazil, in the State of Pará. It is a boomerang-shaped, dominantly inland rift basin with a northeasternmost extension continuing below the Atlantic Ocean (Fig.9). Its on-shore area is about 52 000 km², mostly in the tropical flooded areas and swamps adjacent to the mouth of the Amazonas River.

The tectonic framework of the Marajó Basin includes three major structural domains or depocenters: the Mexiana, Limoeiro, and Cametá grabens. The Mexiana Graben trends SW-NE and links the onshore basin with the equatorial continental margin of Brazil, exhibiting a full graben profile and a complex pattern of normal and strike-slip faults, folds, and magmatism during the rift phase.

The SE-NW striking Limoeiro Graben is the largest sub-basin of the system, showing a SW dipping half-graben profile with a maximum depth to the basement of about 12 000 m. Large listric normal faults trending parallel to the margins of the basin define its structural framework. These normal faults exhibit large displacements between the adjacent blocks. Such downthrows associated with the low angle of the fault planes create huge gaps in map view, with separations that may surpass 10 km. In its central part, the Limoeiro Graben is marked by strong positive Bouguer anomalies and a regional basement high, both features parallel to the axis of the basin. The Cametá Sub-Basin is the southeastern end of the Marajó Basin, and it lies over a thick sequence of Paleozoic rocks corresponding to the same stratigraphic units that fill up the adjacent Parnaíba interior sag.

The Marajó Basin represents the southernmost extension of the Central Atlantic Rift, and its beginning occurred during the Triassic (Fig. 21), when an extensional tectonic regime dominated the region. However, no biostratigraphic information exists to support this statement, which is based solely on the presence of Late Triassic magmatic rocks as sills and dikes in the region. The initial phase of rifting accommodated a thick sequence of barren, dominantly fluvial and deltaic continental sandy sediments named the Calçoene Formation. A second, and in fact the most important, phase of rifting occurred during the Hauterivian(?)-Aptian interval, imposing a significant rearrangement in the pre-existing framework of the basin. This phase accommodated about 5000 m of sandy sediments in a large fault-bounded trough, a sequence known as the Caciporé Formation. Fault activity decreased with time, and an early sag configuration was achieved during the Cenomanian, when dominantly shaly sediments were deposited. This late rift section lies just below a regional unconformity that marks the change from the second rift stage to fully post-rift tectono-sedimentary conditions (Fig. 22).

The post-rift record, named the Marajó Formation, started in the Maastrichtian, and is also a sandy section of marginal marine fluvial and deltaic sedimentary rocks that filled up a large sag basin. Most of this sequence corresponds to the discharge of the Amazonas River drainage basin during the Cenozoic. Locally, this section exhibits indications of marine influence over sedimentation, such as the presence of planktonic foraminifera and glauconitic sandstone beds.

Bragança-Viseu/São Luís/Ilha Nova basins

The Bragança-Viseu/São Luís basins are located in northeastern Brazil (Fig. 9), in the states of Pará and Maranhão, covering an area of 25 000 km². They constitute an onshore aborted rift system that strikes NW-SE, accompanying the orientation of major structural discontinuities of the Precambrian basement (Aranha *et al.*, 1990). In part, they lodge Paleozoic rocks correlated with those of the adjacent Parnaíba Basin. In the offshore region of the State of Maranhão, the São Luís Basin continues into the Ilha Nova Graben, a WNW-ESE trending basin having an area of about 5000 km². Towards the E the Ilha Nova Basin merges with the Barreirinhas Basin, a segment of the open continental margin of Equatorial Atlantic.

The development of this rift system occurred as a result of an extensional stress field active during Aptian time, related to the opening of the Equatorial Atlantic continental margin. The maximum total depth of the basin is about 5000 m. Tectonics and sedimentation terminated at the end of rifting, and the onshore domain of these basins does not exhibits thermal subsidence phase-related deposits. Right-lateral wrenching, related to the regional sense of displacement between the African and South American plates, affected the Bragança-Viseu/São Luís/Ilha Nova basins during the Albian (Aranha *et al.*, 1990).

The rift phase of the Bragança-Viseu/São Luís/Ilha Nova basins is represented by a succession of clastic sedimentary rocks starting with Aptian grey, medium-grained sandstone with subordinate siltstone and conglomerate beds known as the Bragança Formation (Fig. 21), of alluvial and fluvial depositional systems (Lima *et al.*, 1994). This sequence has a maximum thickness of about 300 m, and is followed by a clastic-evaporitic section of Late Aptian age named the Grajaú and Codó formations, that includes fine-grained sandstone, bituminous black shale, anhydrite and calcilutite beds. The organic content of this section shows indications of an anoxic, transitional marine environment. A sequence of marginal marine, fluvial to deltaic reddish sandstone beds, the Itapecuru Formation, ends most of the stratigraphic record of the Bragança-Viseu/São Luís basins. In the Ilha Nova Basin, sedimentation continued up to the end of the Cretaceous through the Periá and Areinhas formations. This final phase of deposition was related to thermal subsidence. Sandy sediments recurred during the Miocene and constitute the Pirabas Formation.

Potiguar Basin

The onshore Potiguar Basin is situated in the north-easternmost corner of Brazil (Fig. 9). It covers an area of about 22 000 km² and it includes a non-outcropping, confined, aborted graben that strikes NE-SW, and is filled with Neocomian to Aptian sedimentary rocks. These rocks are completely covered by a sequence of sub-horizontal Aptian-Campanian strata of regional extent. The sediments are related to a post-rift sag phase, of which only the Albian-Campanian rocks crop out. Thickness in the centre of the buried rift basin attains 6000 m.

The structural framework of the onshore Potiguar Basin includes depocenters or grabens, internal highs and platforms. The grabens are asymmetric features developed along NE-SW normal faults (Fig. 23), with the basement dipping to the SE against the major boundary faults, the Carnaubais Fault being the most important of these. Block rotation along listric faults created some internal highs between adjacent grabens. Shallow platforms border the deep, confined Potiguar Basin on both its eastern and western sides, and do not include an Early Cretaceous record. E-W trending strike-slip faults affected the basin during the Barremian to Early Aptian. Persistent magmatic activity took place during the rift phase in the Potiguar Basin, creating a diabase dyke swarm along the margins of the basin. These igneous rocks, known as the Rio Ceará-Mirim Formation, span the range of 140 Ma - 120 Ma (Araripe and Feijó, 1994). Magmatic activity in the region of the Potiguar Basin is closely related to the active rifting mechanism and left a characteristic signature of local positive Bouguer anomalies (Milani and Latgé, 1987; Milani, 1991). These anomalies reflect deep-seated intrusive bodies, the E-W orientation of which defines a trend that in a pre-drift reconstruction continues into the Benué Trough in Nigeria (Adighije, 1981).

The rift sequence in the Potiguar Basin, up to 6000 m thick (Fig. 21), is represented by the Pendência Formation, consisting of lacustrine shale, fluvial, deltaic and turbiditic sandstone and conglomerate of Berriasian to Early Aptian age. These beds are confined inside the limits of the central graben, with no outcrops. The top of the sequence is marked by a regional unconformity, and the rift sequence is structurally tilted to the SE.

The Alagamar Formation, of Late Aptian age, covers the rift sequence. It represents the transitional, restricted marine phase of evolution of the basin, and consits of black shale, marlstone and calcilutite beds. This section is covered by fine to coarse-grained sandstone beds and subordinate green shale and red siltstone of Albian-Cenomanian age known as the Açu Formation. The stratigraphic record of the onshore Potiguar Basin includes also the Turonian to Campanian high-energy carbonate beds of the Jandaíra Formation. Eocene-Oligocene magmatic rocks constitute the Macau Formation. The last sedimentary deposits in the basin consist of alluvial sand known as Barreiras Formation, of Miocene age.

Recôncavo/Tucano/Jatobá basins

The Recôncavo-Tucano-Jatobá Rift System is situated in the State of Bahia in the northeastern region of Brazil (Fig. 14). This interior rift system strikes N-S, it is about 500 km long and covers an area of about 46 000 km^2. The Recôncavo-Tucano-Jatobá Rift System developed together with the initial stages of opening of the South Atlantic during the Neocomian. By the Aptian it became an aborted interior basin when the extensional stress field was concentrated in the adjacent continental margin of northeastern Brazil, and the South American and African plates started to move apart in that area (Lana and Milani, 1986).

The Recôncavo Basin covers an area of about 1_ 500 km^2 in the southern part of the interior rift system. It is a single graben that strikes in NE-SW, parallel to the Salvador Fault, the major tectonic element that forms the faulted southeastern margin of the rift. The Salvador Horst, situated to the E of the Salvador Fault, is a prominent block where Precambrian granulite is exposed, and separates the interior basin from the offshore Jacuípe Basin. The tectonic framework of the basin includes a set of both synthetic and antithetic NE-SW-trending normal faults that run parallel to the main border, in addition to some NW-SE-striking transfer zones that accommodate the lateral displacement between disrupted crustal blocks (Milani, 1985). The network of faults separates the more stable areas such as the platforms from the subsiding regions. The regional dip of the basement is to the SE. In the main depocenter, the Recôncavo Basin contains about 7000 m of sediments.

Neocomian syn-rift sedimentation in the Recôncavo Basin was preceded by the accumulation of a pre-rift succession of rocks (Fig. 21), indicating the initial, unfaulted stage of crustal stretching. This Late Jurassic pre-rift sequence, or Brotas Group, includes the fluvial-lacustrine red beds of the Aliança Formation and a blanket of grey and reddish fluvial-eolian sandstone, the Sergi Formation. The syn-rift sequence is composed of a dominantly lacustrine shaly section with sediments rich in organic matter and turbiditic sandstone beds in the basal part, included in the Candeias and Maracangalha formations. These beds are followed by prograding sediments forming deltas, the Ilhas Group, and fluvial sandstone beds, the São Sebastião Formation. The Salvador Formation is a thick wedge of conglomerate that occurs adjacent to the Salvador Border Fault (Fig. 24) and interfingers with the other stratigraphic units of the syn-rift sequence. A 250m thick succession of Late Aptian alluvial sedimentary rocks, the Marizal Formation, constitutes the limited vestige of a post-rift record related to thermal subsidence in this basin (Caixeta *et al.*,, 1994).

The Tucano and Jatobá basins cover an area of about 35 000 km^2, mostly in the arid, flat and dusty region of the *caatinga*, a characteristic biotic and climatic domain in the interior of this region of Brazil. These basins are completely surrounded and underlain by Precambrian rocks of the Atlantic Shield, the lines of structural weakness of which were followed during the Neocomian rifting.

The Tucano Basin is the major N-S striking segment of the rift (Fig. 14). It consists of three sub-basins: Southern, Central and Northern sub-basins, each separated by important transfer zones. Large normal boundary faults, striking in NE-SW, controlled the subsidence of the individual grabens of Southern Tucano and Central Tucano sub-basins, where the basement dips to the SE. The depocenter of Central Tucano Sub-Basin, expressed on gravity data as an impressive -140 mGal anomaly, is the deepest part of the rift system, being filled with over 12 000 m of sediments (Milani, 1987). An important inversion occurs between the Central and Northern Tucano sub-basins, along the Vaza-Barris Arch, and in the Northern Tucano Sub-Basin the basement dips to the W (Santos *et al.*, 1990). The Jatobá Basin is the northernmost, E-W-striking element of this rift system, and there the Ibimirim Fault controlled the subsidence, also due to Mesozoic reactivation of an ancient Precambrian shear zone.

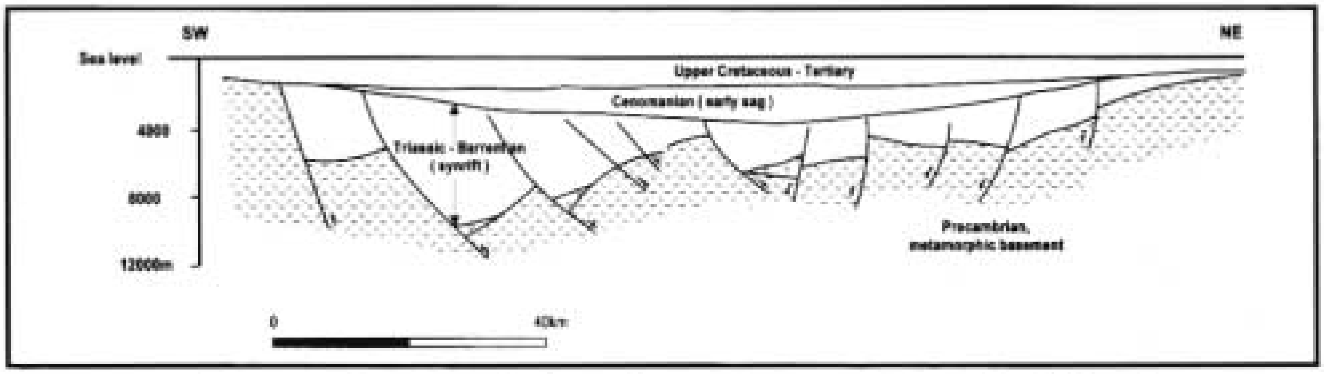


FIGURE 22 - Conceptual cross-section of the Limoeiro Graben, the main depocenter in the Marajó Basin (after Milani and Zalán, 1998).

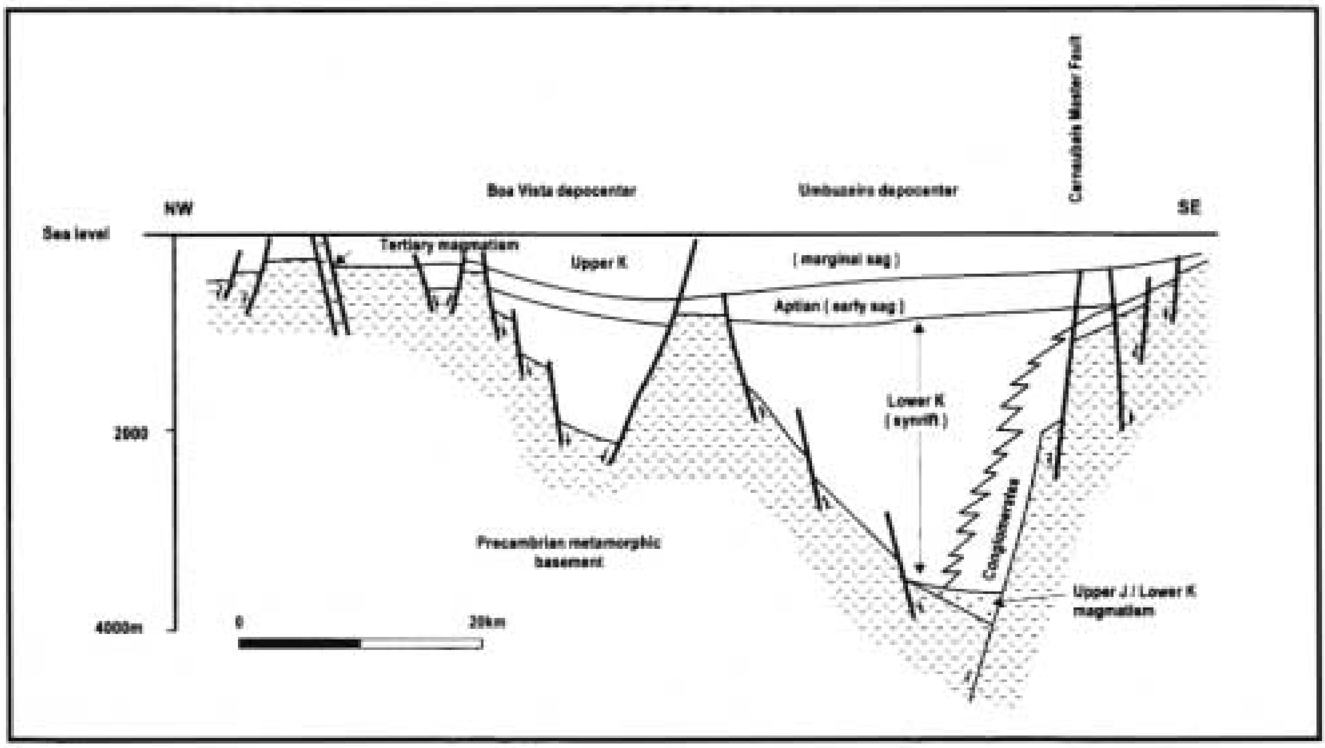


FIGURE 23 - Cross-section of the onshore portion of the Potiguar Basin (after Bertani et al., 1990; permission to reprint granted by the authors).

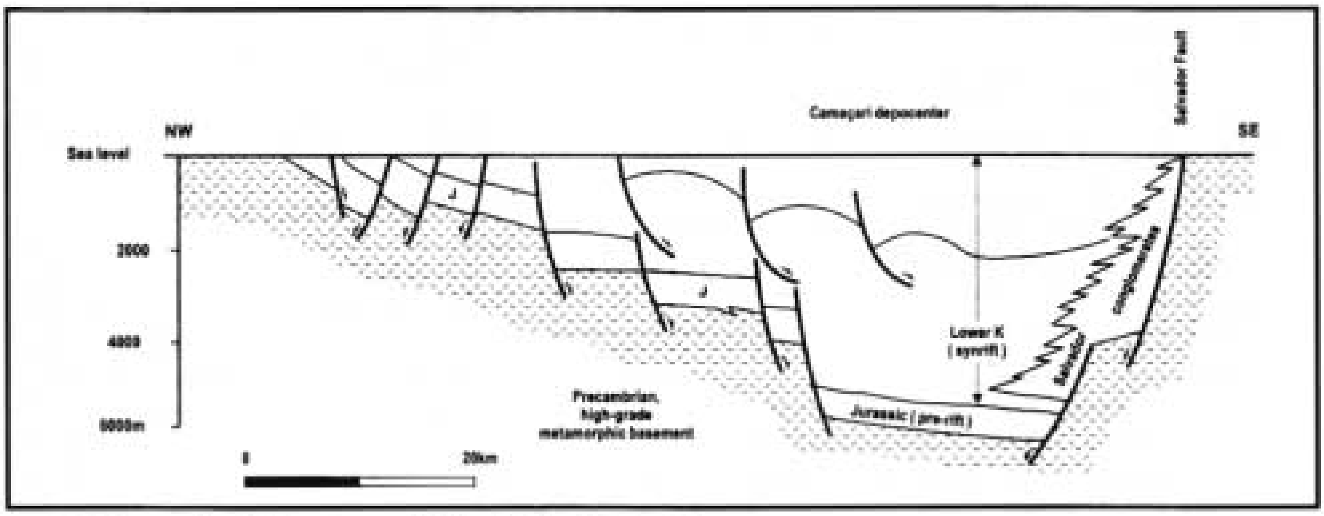


FIGURE 24 - Cross-section of the Recôncavo Basin (after Milani and Zalán, 1998).

Remnants of a previous, widely distributed Paleozoic cratonic sedimentation can be locally found below the Late Jurassic to Early Cretaceous sediments of the Tucano and Jatobá basins (Caixeta *et al.*, 1994), as well as elsewhere in the northeastern region of Brazil. In the Northern Tucano Sub-Basin and Jatobá Basin, Paleozoic strata are thicker and include Silurian to Devonian fluvial sandstone beds of the Tacaratu and Inajá formations, and Carboniferous to Permian sandstone, limestone, and evaporite units and bituminous black shale beds comprising the Curituba, Santa Brígida and Afligidos formations.

Pre-rift and syn-rift sedimentation in the Tucano and Jatobá basins resembles that of the Recôncavo Basin, including the same lithostratigraphic nomenclature. However, remarkable differences exist and are related basically to the overall character of the deposits, with a sandy component being dominant in the northern basins. This regional picture seems to represent the north-to-south sense of progradation of the axial fluvial-deltaic system that filled up the huge Recôncavo-Tucano-Jatobá interior rift system.

Salado Basin

The Salado Basin is located in the Province of Buenos Aires in Argentina, and occupies an area of 85 000 km², about half of it being in the onshore domain. The main axis of the basin is linear, and trends NW-SE, parallel and adjacent to the Rio de La Plata Estuary (Fig. 14).

Seismic information suggests that the basin overlies a sequence of siliciclastic sedimentary rocks of probable Paleozoic age (Tavella and Wright, 1996), equivalent to the sequences found in the inland Chaco-Paraná Basin. The rift stage is interpreted as having started in the Late Jurassic and peaked during the Neocomian, taking advantage of ancient weakness zones in the basement (Stoakes *et al.*, 1991). Volcanic rocks occur associated with the lacustrine sediments of the rift phase and are named the Rio Salado Formation (Fig. 25). These are essentially beds of red sandstone, conglomerate and argillite. A regional, angular unconformity separates the 1200 m thick rift sequence from the following, post-rift section.

The early post-rift sequence, with ages spanning the range Aptian-Senonian, is represented by a sequence of 2000 m of transitional sediments related to a phase of sagging, including alluvial, fluvial and deltaic depositional systems known as the Gral. Belgrano Formation, that interfingers with the shallow marine deposits of the Chilcas Formation. Fully passive margin conditions were reached by the Cenozoic, when the subsidence rate decreased, and the eustatically-controlled sedimentary successions of the Paz and Valeria formations were accumulated. These included platform shales, beds of deltaic sandstone and subordinate carbonate units.

Colorado Basin

The Colorado Basin is situated in the southeastern corner of the Province of Buenos Aires in Argentina, covering an area of about 90 000 km², mostly in the offshore domain (Fig. 14). The axis of the basin trends in E-W and the trough is confined between important positive features, the Rawson High-North Patagonian Massif to the S-SE and the Sierras Australes (or La Ventana Fold and Thrust Belt) to the NW (Fryklund *et al.*, 1996).

The rifting stage in the Colorado Basin started during Late Jurassic to Early Cretaceous times (Lesta *et al.*, 1980), over a basement of Paleozoic sedimentary strata and granitic rocks. The sedimentary basement corresponds to the offshore extension of the Claromecó Basin, a Permian foreland province related to the evolution of the Cape-La Ventana Fold Belt (López-Gamundí *et al.*, 1995).

A classical conglomeratic-sandy-shaly sequence up to 2000 m thick, the Fortín Formation, represents the rift phase in the Colorado Basin (Fig. 25). Its occurrence is confined to half-grabens, the development of which followed the WNW-ESE grain of the La Ventana Fold and Thrust Belt, especially along the northern border of the basin (Juan *et al.*, 1996). Uplift and erosion were important mechanisms during the final stages of the rift phase (Juan *et al.*, 1996).

Albian to Cenomanian siliciclastic sedimentary rocks several thousands of metres thick represent the sag phase, and are named the Colorado Formation, lying unconformably over the rift sequence. Open marine conditions prevailed onwards, starting with the deposition of the shaly sediments of the Pedro Luro Formation, of Cenomanian to Paleocene age. Recurrent tectonic activity is suggested by the presence of the Ranquel flood basalts that are associated with the Pedro Luro beds (Keeley and Light, 1993). A section of intercalating glauconitic sandstone beds and shale accumulated during Eocene times and constitutes the Elvira Formation. These rocks are overlain by the dominantly shaly Barranca Final Formation. Both formations are related to the passive margin stage of evolution of the continental margin of Argentina (Fig. 26).

Rawson/Península de Valdés basins

The Rawson/Península de Valdés basins are two parallel-to-coastline, fully offshore troughs with areas of, respectively, 35 000 km² and 20 000 km², linked along their northern boundaries by a system of E-W trending small grabens (Fig. 14). The Rawson Basin contains some 5000 m of sediments, and the Península de Valdés Basin contains a sequence around 3300 m thick. In like manner to observations made in some of the aborted rift basins of eastern Argentina, the Rawson and Península de Valdés basins overlie Paleozoic sedimentary rocks. In this case these are Silurian and Devonian strata of the Sierra Grand Formation (Yrigoyen, 1989) that dip at about 30° and exhibit the same Jurassic to Neocomian age of rifting observed in correlative basins.

The sedimentary record of the rift stage is made by fining upwards cycles that include basal fluvial sandstone beds grading to argillite (Marinelli and Franzin, 1996). Volcanic rocks of Jurassic age (181±10 Ma, Linares, 1977) are associated with the sequence of red beds that filled the rift basin (Fig. 25). The rift deposits were suceeded by the sag phase sequence, a package of transgressive glauconitic sandstone beds that passes upwards to marine shale, the deposition of which ended by Late Cretaceous times. The Cenozoic section consists of argillite, thin beds of glauconitic sandstone and fossiliferous limestone units. Effusive basaltic bodies of Paleocene age (Yrigoyen, 1989) are also found in the Rawson Basin, and occur along some unconformity surfaces that break the open marine section.

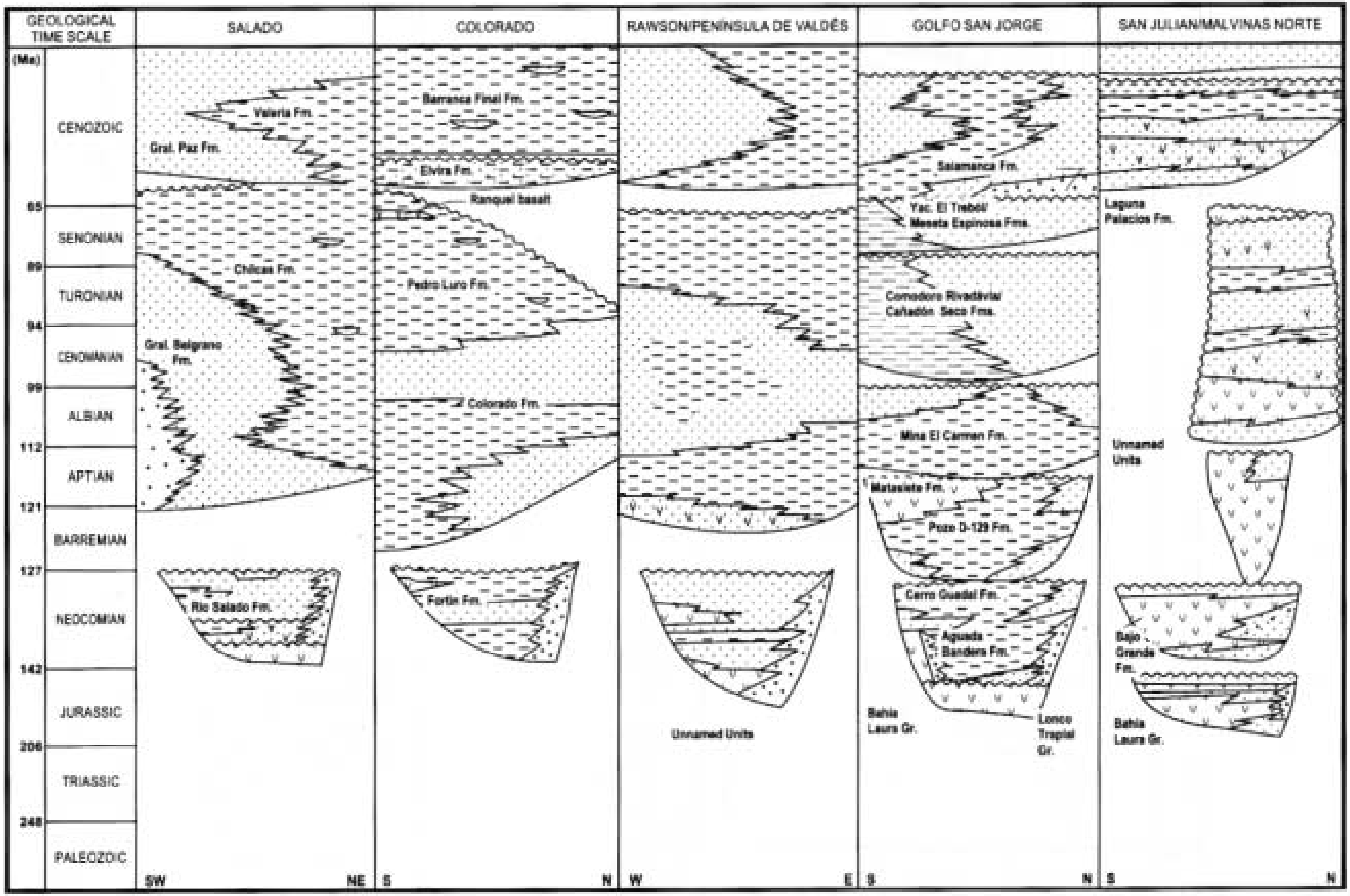


FIGURE 25 - *Stratigraphic summary of the failed rift basins in the Argentinean sector of the divergent margin of South America. Main sources of information were Tavella and Wright (1996), Salado Basin; Juan et al. (1996) Fryklund et al. (1996), Colorado Basin; Marinelli and Franzin (1996) Yrigoyen (1989), Rawson/Peninsula de Valdés Basins; Baldi and Nevistic (1996), Golfo San Jorge Basin; Figueiredo et al. (1996) Ross et al. (1996), San Julian/Malvinas Norte Basins. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.*

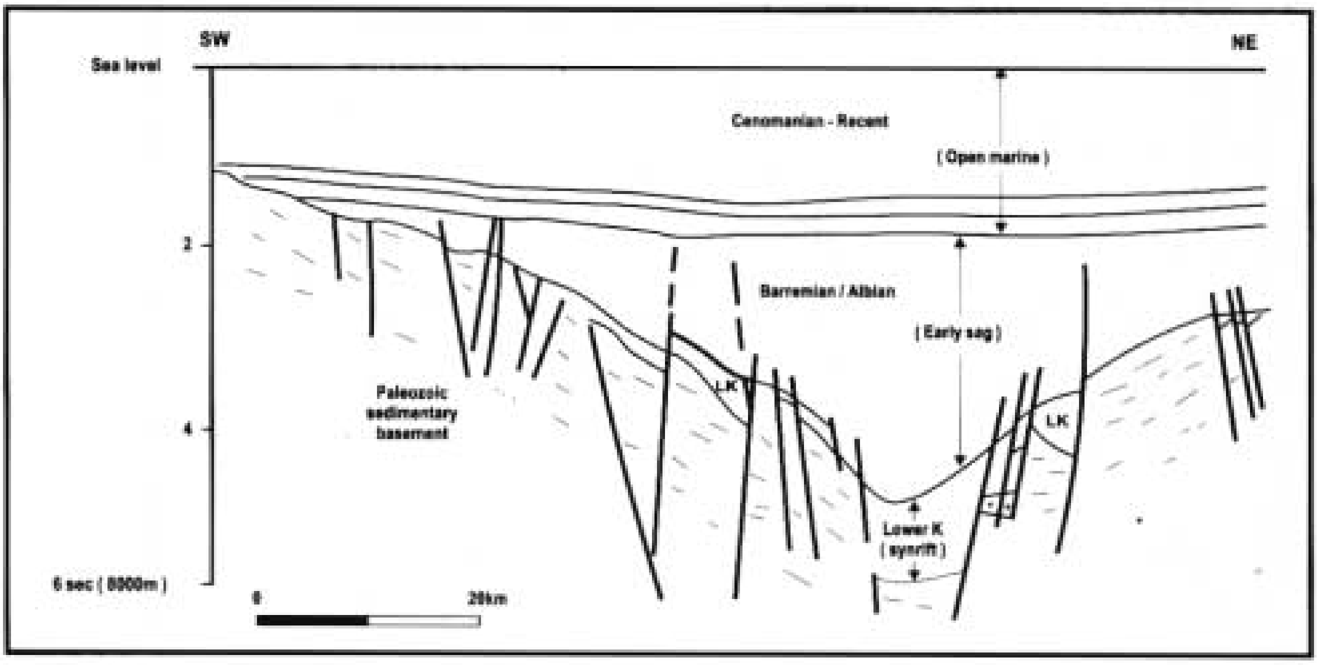


FIGURE 26 - Geoseismic cross-section of the Colorado Basin (based on seismic information and interpretation from Fryklund et al., 1996; permission to reproduce granted by Asociación Geológica Argentina).

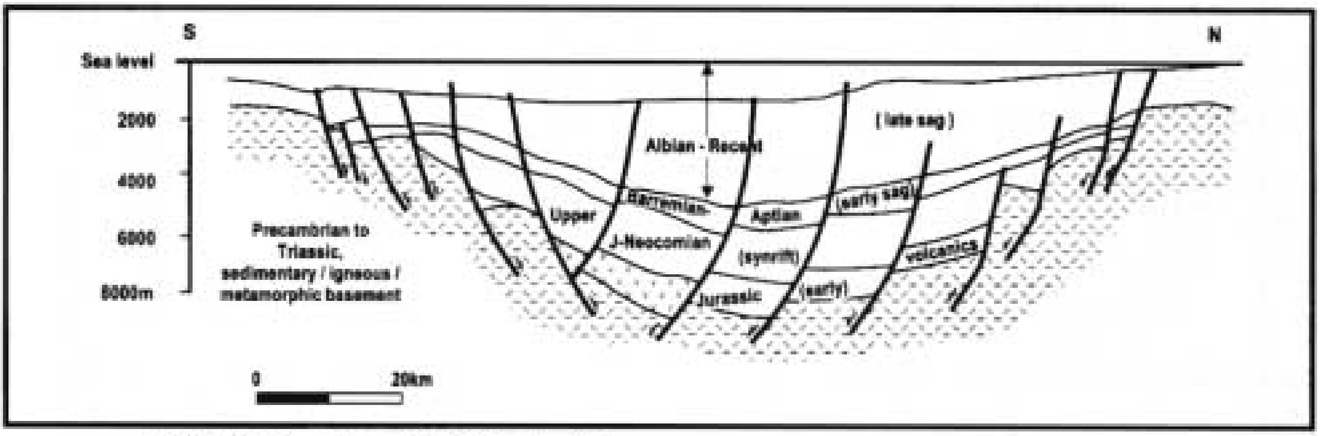


FIGURE 27 - Cross-section of the Golfo San Jorge Basin (after Baldi and Nevistic, 1996; permission to reproduce granted by Asociación Geológica Argentina).

Golfo San Jorge Basin

The Golfo San Jorge Basin has an area of about 170 000 km^2, being two thirds onshore, and is situated in the central sector of the Argentinean Patagonia (Fig. 14). The limits of the basin are defined by important tectonic elements of southern South America: the Andean Belt to the W, and the N-S-trending Patagonia Oriental High in offshore regions to the E (Urien and Zambrano, 1996). To the S and N, respectively, there occur the cratonic blocks of Deseado and Nord Patagonico. The basin mainly contains Late Jurassic-Cretaceous sediments that attain a maximun thickness of 7000 m, the depocenter being situated in the onshore domain of the basin.

The main extensional structures in the framework of the Golfo San Jorge Basin run E-W. Its tectonic development started during Jurassic time (Baldi and Nevistic, 1996) with the beginning of rifting that created a series of normal fault-controlled depocenters. Magmatism was associated with this initial rift phase. Furthermore, during the Late Jurassic and Early Cretaceous, the rate of mechanical subsidence decreased and the initial isolated depocenters were joined into a single sag basin. From early Cenozoic times onward, the basin experienced a compressional phase related to the Andean Orogeny that produced structural inversion along a series of pre-existing normal faults and a regional positive area in the western part of the basin. This structure is the N-S striking San Bernardo Range. Several changes in sedimentary facies development and distribution also occurred during this phase of structural inversion.

The Golfo San Jorge Basin is filled with siliciclastic, volcanic and volcaniclastic deposits (Barcat *et al.*, 1989). During the Jurassic early syn-rift phase mainly igneous rocks accumulated there, included in the Lonco Trapial Group, in the northern region, and Bahía Laura Group in the southern area (Fig. 25). The 800 m thick lowermost sedimentary sequence, related to the late syn-rift stage, consists of lacustrine shale, rich in organic matter, and fluvial-deltaic sandstone beds named the Pozo D-129, Anticlinal Aguada Bandera, and Cerro Guadal formations (Lesta *et al.*, 1980). The early sag phase accommodated lacustrine black shale beds in the centre of the basin, the Pozo D-129 Formation, that grades laterally into sandstone and volcaniclastic units of the Matasiete Formation.

From Albian times onward, the Golfo San Jorge Basin experienced its late sag phase of evolution, dominated by thermal subsidence (Fig. 27). The Chubut Group represents this stage, and consists of a basal sequence of shale beds grading upward to sandstone deposits, the Mina El Carmen Formation. These beds are followed by a dominantly sandy, fluvial to deltaic section included in the Comodoro Rivadávia and Cañadón Seco formations that developed from N to S (Baldi and Nevistic, 1996). During the final stages of the Cretaceous, the remnants of the sag depositional space were filled up by a progradational succession of sedimentary rocks known as Yacimiento El Treból and Meseta Espinosa formations.

Due to particularly high freeboard conditions towards the domain of the Atlantic Ocean the Golfo San Jorge Basin received its first marine sediments only during the early Cenozoic, and such a marine record is represented by the conglomerate, argillite and sandstone beds of the Salamanca Formation. The rest of the stratigraphic record of the Golfo San Jorge Basin, dominantly of Oligocene to Miocene age, consists of alternating continental sedimentary rocks and transgressive marine beds.

San Julian/Malvinas Norte basins

The San Julian Basin is an offshore trough situated to the SE of the Golfo San Jorge Basin (Fig. 14), on the Atlantic margin of Argentina. It has an area of about 5 000 km^2, and trends in E-W. The maximum thickness of its volcaniclastic-sedimentary sequence is about 6000 m. The basin developed over the northern border of the Deseado Massif, that is considered an exotic terrane docked against the border of Gondwana during the Late Paleozoic (Figueiredo *et al.*, 1996).

The tectono-sedimentary evolution of the San Julian Basin started with a rift phase of Middle Jurassic age (Fig. 25), when an up to 1400 m of massive tuff beds, brecciated argillite and sandstone with tuffaceous matrix were accumulated. These beds are known as the Bahia Laura Group, and they are exposed in the inland Patagonia area. They have been dated at 168±2 Ma to 170±4 Ma by Rb/Sr method (Rapela and Pankhurst, 1994).

The Bajo Grande Formation that follows is still dominated by volcaniclastic rocks and is of Late Jurassic-Neocomian age. According to Figueiredo *et al.* (1996), it corresponds to a phase of sagging with some remnant activity of the major boundary faults. Tuff beds derived from subaerial ash flows, including glass shards and exhibiting fine vitric textures mainly represent sedimentation during this phase. This unit and the previous one were involved in post-depositional compressional deformation of Cretaceous age, caused by stresses developed along the Malvinas (Falklands) Fault Zone during the early stages of drifting of South America to the W. This resulted in an angular unconformity separating the Neocomian sequence from the flat-lying Maastrichtian sediments. The Maastrichtian sediments, named Laguna Palacios Formation, are likewise tuffaceous and about 100 m thick.

Prograding, siliciclastic sedimentation started in the Paleocene with the first marine incursion, and covered almost completely the San Julian Basin during the Cenozoic. The tectonic environment became very quiet, but some minor reactivation of border faults occurred. These were probably related to the Miocene Andean movements along the western border of the continent.

The Malvinas Norte Basin occupies an area of about 30 000 km^2 and is situated to the N of the Malvinas (Falklands) Islands, in the off-shore region of southern Argentina (Fig. 14). It is a Y-shaped, NNW-SSE-trending system of grabens accommodating a still undrilled sequence having a maximum thickness, estimated by seismic data, of about 9000 m. Based on lithological information from the San Julian Basin, located some 350 km to the NW, the sediments of the Malvinas Norte Basin are interpreted to be of siliciclastic-volcaniclastic nature, and the main aspects of its tectono-sedimentary-magmatic evolution seem to closely resemble those of the adjacent trough.

Andean, Convergent Margin

Back-arc flexural basins

Phanerozoic foreland basin evolution in South America has been closely related to the development of its western, active margin. A convergent relationship between the former western Gondwana continent and the oceanic floor of Panthalassa is recognized back in time as far as the Late Proterozoic (Ramos, 1988; Bahlburgh and Breitkreuz, 1991; Gohrbandt, 1993). Flexural subsidence in different sectors of this compressional domain remained active during almost all the time, providing the space for accumulation of sediments at various time-intervals during its Phanerozoic geological history.

The sub-Andean foreland basins, related to the climax of Andean orogenic deformation and adjacent flexural subsidence, developed in the late Cenozoic. The sub-Andean setting is a continent-size geological domain spanning from Venezuela in the N, to Tierra del Fuego, in the southernmost tip of Argentina. Regional arches, trending perpendicular to the Andean Belt and defining individual basins locally interrupt the continuity of this foreland trough. However, specific and local names of basins and sub-basins do not always represent naturally individualized provinces; in some cases, they correspond to particular denominations adopted by only one of the various Andean countries that share this large province.

Considering the existing main tectonic limits, the sub-Andean domain is divided into four large provinces (Fig. 28). The northernmost extends from Venezuela to Colombia and corresponds to the Llanos-Barinas-Apure Basin. It is limited to the S by the Macarena Arch. The second province includes regions from southern Colombia, Ecuador, Peru and westernmost Brazil, and is known by the local names including the Putumayo, Oriente, Marañon, Ucayali and Acre basins. This province terminates to the S at the Fitzcarrald High. The third province is the Central Andean domain, and corresponds to southernmost Peru, Bolivia, northwestern Argentina and a small fraction of Brazil. The Madre de Dios-Beni-Chaco Basin dominates this region. The fourth province developed mostly in the Argentinean-Chilean Patagonian region includes the Neuquén and Austral basins, as well as smaller depocentres such as the Ñirihuau Basin (Cazau *et al.*, 1989).

The Phanerozoic, pre-Andean history of the Panthalassan margin of Gondwana also recorded several episodes of basin formation, occupying certain periods of time. Permian-Carboniferous foreland basins including Tarija, Calingasta-Uspallata, Paganzo, San Rafael and Claromecó (López-Gamundí *et al.*, 1994), and older basins such as the Silurian-Devonian depocenter in the Precordillera of Argentina (Ramos, 1993), had their evolution linked to tectonic loading and lithosphere flexure of specific sectors of the southwestern margin of Gondwana (Milani and Ramos, 1998).

Pre-Andean Foreland basins

Tarija-Noroeste Basin

The Tarija-Noroeste Basin is a large sedimentary basin containing a thick sequence of Paleozoic sediments, and covering an area exceeding 200 000 km^2. It covers parts of southeastern Bolivia and northwestern Argentina, and the Paraguayan Chaco. (Wiens, 1995). The stratigraphic succession of this region of South America represents an almost complete record of the Phanerozoic history of the western margin of Gondwana (Sempere, 1995). During most of this time, and in spite of being positioned adjacent to the active margin of the plate, the region behaved as a relatively stable platform having gentle dips towards Panthalassa, with minor compressional deformation associated. The existence of a very rigid basement, of cratonic nature, might be a reasonable explanation for the resistance of this particular realm.

Some of the Paleozoic tectonic episodes that affected the borders of the region and determined the main characteristics for each tectono-sedimentary cycle of deposition were remarkable. During Cambrian and Ordovician times, this domain behaved as a large back-arc margin open to the W (Starck, 1995), but this phase was interrupted by the Ocloyic Orogeny, of Late Ordovician age (Ramos, 1988). There followed a Silurian-Devonian foreland phase of sedimentation that ended by an episode of folding and basin inversion during the Late Devonian-Early Carboniferous that is known as the Chanic Orogeny.

The Permian phase of deposition accompanied the evolution of the Aymara Tectonic Cycle (França *et al.*, 1995), when the growth of the Choiyoi Magmatic Arc to the W progressively constrained the marine incursions upon the foreland domain. The Mesozoic was a time of widespread development of extensional basins (Starck, 1995), up to the early Cenozoic, when the transition to a renewed compressional regime and inversion of previous rift structures marked the beginning of the Andean Foreland Cycle in this domain. The remnants of the still undeformed Tarija-Noroeste Basin lie today in the sub-Andean flats of Bolívia and north western Argentina.

The Phanerozoic stratigraphic record of this region (Fig. 29) was summarized by Sempere (1995), for the Bolivian sector; by Wiens (1995), for the Paraguayan Chaco; and by Starck (1995), for the Argentinean Noroeste region. The sedimentary sequence in most of the Bolivian Tarija domain begins with a marine transgression, represented by the Upper Cambrian to Lower Ordovician strata of the Tacsara Supersequence that accumulated in a back-arc, NW-SE trending extensional trough (Sempere, 1995). These beds are followed by the Chuquisaca Supersequence, spanning the Late Ordovician to Middle Famennian. The lower part of this supersequence contains marine black shale in the Tokochi Formation, diamictite beds in the Cancañiri Formation, and a turbidite sequence about 1500 m thick, included in the Llallagua Formation. The upper part of the Chuquisaca Supersequence, of Silurian-Devonian age, consists of a blanket-like body of sediments and is represented by three major coarsening upwards cycles. Maximum flooding levels are represented by the shaly intervals of the Kirusillas, Chululuyoj, Icla and Los Monos formations. Remarkable regressive sequences are those of

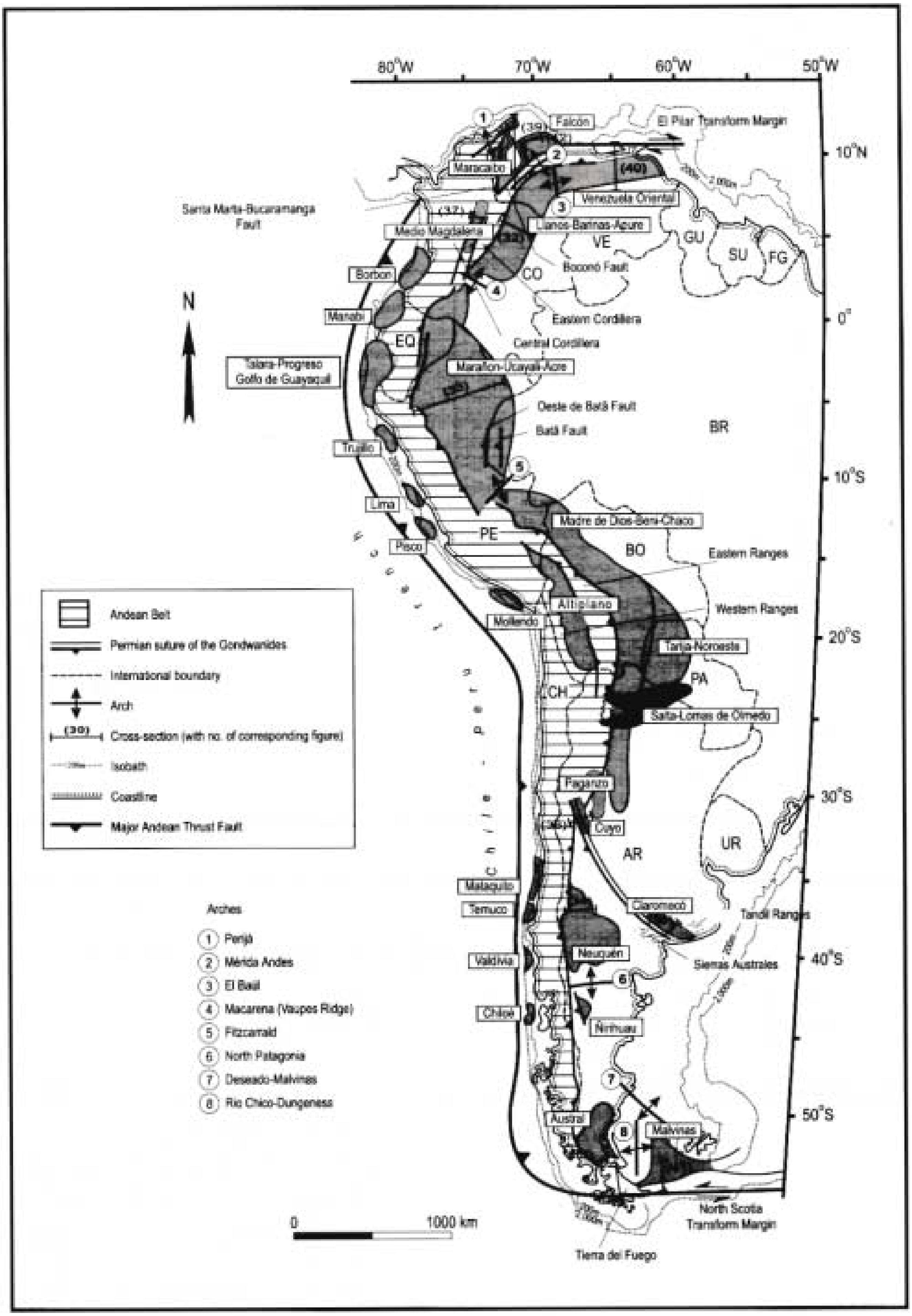
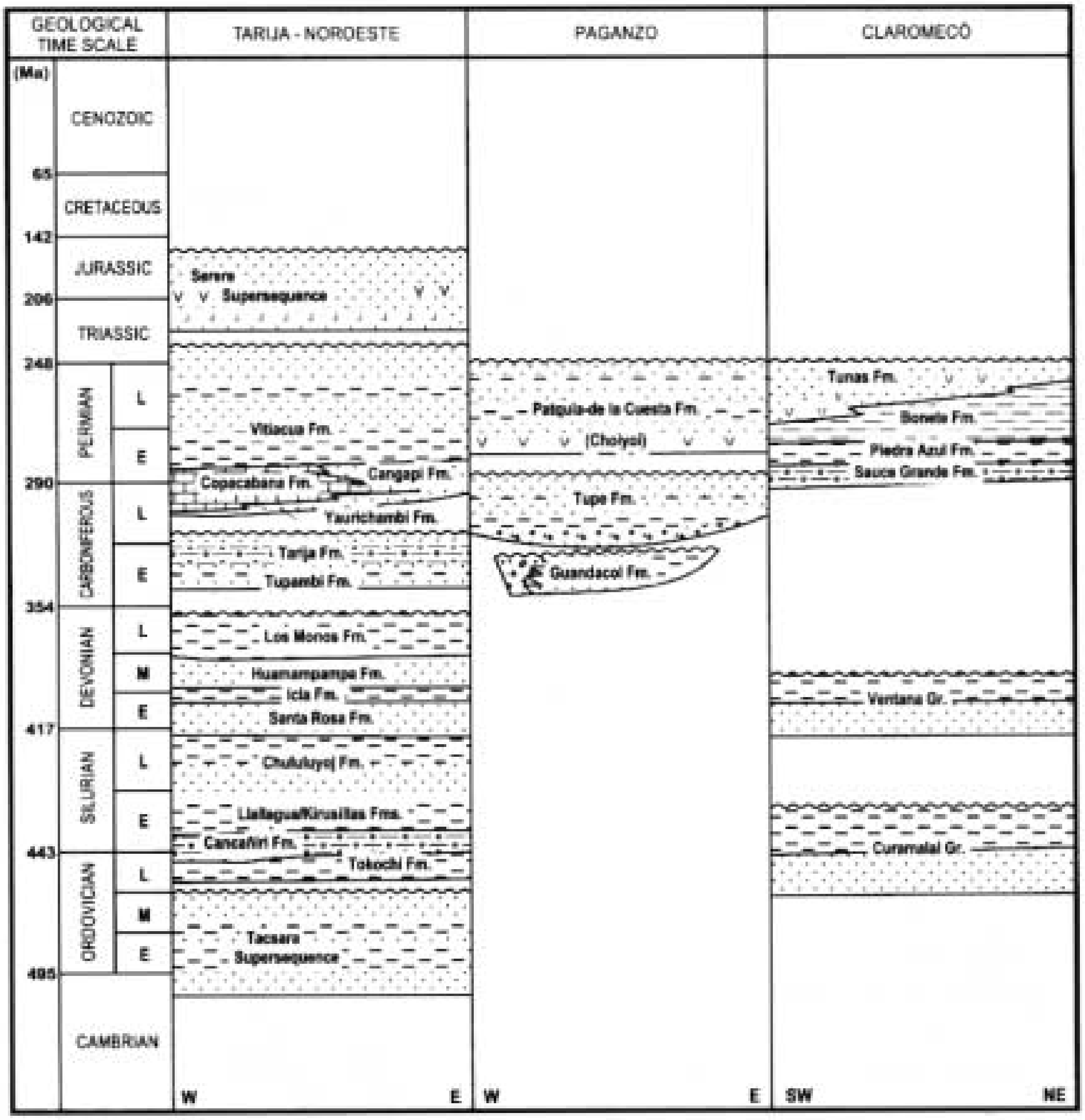


FIGURE 28 - The Andean Convergent Margin, and the Northern and Southern transform margins of South America (compiled from various sources). Countries: see Fig. 2.



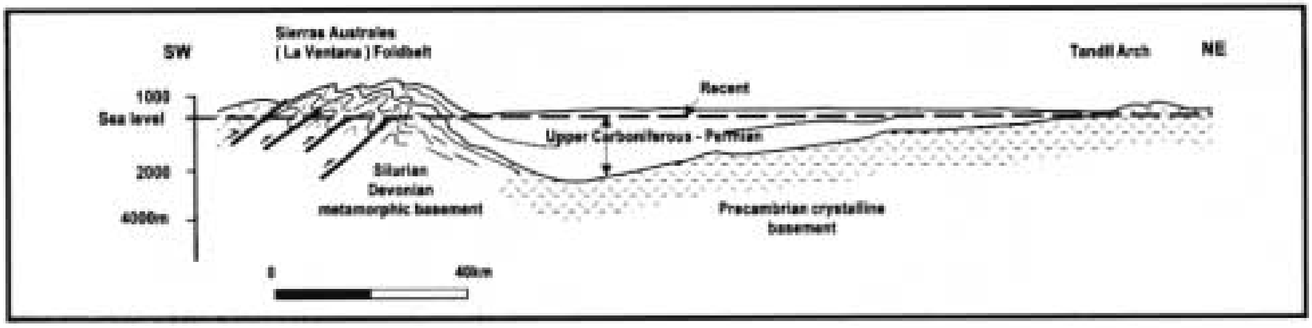


FIGURE 29 - Stratigraphic summary of some of the pre-Andean foreland basins in South America. On each chart, the folded belt margin is to the left. Main sources of information were Sempere (1995), Wiens (1995), Starck (1995), Tarija-Noroeste Basin; Fernández-Seveso and Tankard (1995), Paganzo Basin, López-Gamundí et al. (1994, 1995), Claromecó Basin. Geological time scale simplified from Gradstein and Ogg (1996), Lithological representations are of common use.

FIGURE 30 - Cross-section of the Claromecó Basin (after López-Gamundí et al., 1994; permission to reproduce granted by the Geological Society of America).

the Santa Rosa and Huamampampa formations.

The accumulation of the basal beds of the Gondwana succession during the Carboniferous marked the beginning of a clear tendency for development of continental environments. This continued throughout the Paleozoic and into the Mesozoic. Resedimentation and some factors related to the great Gondwanan glaciation, together with tectonic activity, were important mechanisms in the accumulation of the Carboniferous sediments. These are represented by the Tupambi and Tarija formations. In contrast, subtropical climate and tectonic quiescence led to the deposition of the carbonate platform sediments of the Copacabana Group during the Pennsylvanian. These limestones grade southeastwards into the sandy continental sediments of the Yaurichambi and Cangapi formations. Kungurian black shale of the Vitiacua Formation mark the last marine incursion in this area during the Paleozoic. The Carboniferous-Permian section is also known as the Cuevo Supersequence (Sempere, 1995).

From Triassic times onward, continental red beds, evaporites and associated basaltic flows and intrusions of the Serere Supersequence marked the regional stratigraphic scenario, a succession that culminated in the extensive Jurassic eolian deposits of the Ichoa Formation. This section corresponds, in part, to the Tacuru Supersequence of Starck (1995). In latest Jurassic time, this area was affected by large-scale rifting related to the initial stages of the Andean Orogeny. The Cenozoic development of the region includes the development of subsidence in the Chaco Basin. This subsidence is closely related to the growth of the Andean Chain as the result of progressive compressional deformation along the western side of the Tarija Basin and the consequent creation of a foredeep for the accumulation of several thousands of metres of terrestrial beds up to Recent times.

Paganzo Basin

The Paganzo Basin occupies about 30 000 km^2 of west-central Argentina (Fig. 28), and contains some 4500 m of Carboniferous to Permian sedimentary rocks. The basin developed over a pre-Carboniferous sedimentary basement that was deposited upon a mosaic of crustal blocks welded together during the Late Proterozoic-Cambrian Brasiliano Orogeny (Ramos, 1988). This framework of sutures in the basement provided the main lines of weakness along which the Paganzo Basin formed during the Mississipian.

The main characteristics of Lower Paleozoic sedimentary basins along southwestern Gondwana, including their dominantly marine nature, were modified by the Late Devonian-Early Carboniferous Chanic Orogeny. This intense tectonic episode is marked throughout the continent by an important hiatus, and was succeeded by the accumulation of dominantly continental deposits (Fernández-Seveso and Tankard, 1995). In the region of the Paganzo Basin, the Chanic movements were responsible for major folding and faulting of the pre-Carboniferous sedimentary basement (López-Gamundí *et al.*, 1994).

To a certain extent, the Paganzo Basin owed its origin to the Chanic stress field, by means of transtensional reactivation of weakness zones deeply-seated in the fabric of the basement. Towards the W, the Paganzo depocenter was linked to a persistently marine N-S trending trough known as the Calingasta-Uspallata and San Rafael basins (López-Gamundí *et al.*, 1994), a foredeep flexural domain adjacent to the Chanic Orogen.

The lowermost stratigraphic record of the Paganzo Basin consists of syntectonic deposits containing a Visean flora (Fig. 29). This unit, the Guandacol Supersequence (Fernández-Seveso and Tankard, 1995) is about 2000 m thick, and includes coarse-grained sedimentary rocks adjacent to basin-boundary faults and classical distal turbiditic systems; rhythmic lacustrine shale and siltstone beds with local dropstones also occur and are suggestive of a periglacial environment during Visean-Namurian times.

The initial geometry of isolated depocenters evolved to a widespread, uniformly subsiding flexural depression during the deposition of the second unit, the Tupe Supersequence, and this change in subsidence style is marked by an important unconformity. These Westphalian-Stephanian sediments are some 1300 m thick and are composed of four progradational sequences representing fluvial, lacustrine and marginal marine depositional environments, reflecting a pattern of transgressive-regressive cycles of sedimentation. A third pattern of subsidence in the Paganzo Basin is recorded by the Patquía-de la Cuesta Supersequence, mainly of Late Permian age, when the depositional area expanded even more than during Tupe deposition. The basin assumed a style of shallow epeiric sag, almost without relief, and this configuration persisted up to the beginning of the Triassic.

The Patquía-de la Cuesta Supersequence consists of continental red beds including fluvial and playa lake facies up to 1300 m thick deposited in a shallow marine basin, the remnants of the "Tupe Sea" (Fernández-Seveso and Tankard, 1995). Pyroclastics and magmatic flows related to the Choiyoi episode occur at the base of the unit. Some beds with bituminous black shale occur in the Late Kungurian-Kazanian section, indicating a time of rising of sea level. This sequence, as well as the record of the whole Paganzo Basin, terminates with a major unconformity that cuts deeply into the pre-existing deposits. This erosional surface is related to the extensional tectonics that created the successor, Triassic back-arc rift basins of western Argentina.

Claromecó Basin

The Claromecó Basin, also known as Sauce Grande Basin or Ventania Paleozoic Basin, is situated in the Province of Buenos Aires, Argentina. The basin covers an area of about 40 000 km^2 (Fig. 28). This onshore Carboniferous-Permian basin occurs between two prominent structural highs, the Sierras Australes Fold and Thrust Belt to the SW and the Tandilia Arch to the NE. To the SE it sinks below the younger sequences of the offshore Colorado Basin.

The pre-Carboniferous units in the Claromecó Basin area are represented by Precambrian igneous and metamorphic basement rocks and by Early Paleozoic sedimentary successions. The sedimentary rocks are included in two major units (Fig. 29). The first is the Curamalal Group of probable Ordovician-Silurian age, and the second is the Ventania Group of Devonian age (Harrington, 1947, *apud* López-Gamundí *et al.*, 1994). The sediments of both these groups have a total thickness of about 2000 m. They were deposited along the pre-Carboniferous margin of Gondwana open to Panthalassa. The top of this sequence is marked by a gentle regional angular unconformity (Massabie and Rossello, 1984).

The Latest Carboniferous to Late Permian Pillahuincó Group is up to 4500 m thick, and represents the sedimentary accumulation in the Claromecó Basin. These sediments consist of a quartzose, dominantly diamictitic unit at the base, the Sauce Grande Formation, that may reach a thickness of 1100 m, and which exhibits paleocurrents that indicate a provenance from the cratonic areas to the NE. The unit grades to a shaly sequence, the Piedra Azul Formation that continues into the bioturbated shale and sandstone beds of the Bonete Formation. The Bonete Formation is an important stratigraphic marker of regional significance due to the presence of the *Eurydesma* fauna of Early Sakmarian (Tastubian) age.

The following unit, Tunas Formation, shows some contrasting characteristics with respect to the underlying sediments. The sediments are less quartzose and the southwestern provenance indicated by paleocurrents (López-Gamundí *et al.*, 1994), together with the presence of tuffaceous beds, intercalated with the Permian red beds, indicate magmatic activity in the surrounding areas as well as uplift of Sierras Australes to the SW. These sediments provided the detritus to fill this syn-tectonic foreland trough (Fig. 30). Rapid subsidence accommodated the 1500 m of the Tunas Formation close to the orogenic front. A final compressional phase of deformation in the area occurred already in the Triassic, that also involved the sediments of the Tunas Formation.

Andean Foreland basins

Llanos-Barinas-Apure Basin

The Llanos Basin is situated in northeastern Colombia, and it lies to the E of the Andean frontal thrust (Fig. 28). In the neighbouring Venezuela it is named Barinas-Apure Basin, and has a total area of about 200 000 km². The Llanos Basin is a classical example of a complex, foreland multi-cycle basin, the development of which started in the Triassic-Jurassic. The main tectonic events that influenced its evolution are related to the convergent plate setting of the western margin of South America.

Cooper *et al.* (1995) defined the major stages in the evolution of the Llanos Basin, including a Triassic-Jurassic rift phase followed by a phase of periodic extension in a back-arc setting during Barremian to Maastrichtian times. The onset of the pre-Andean foreland subsidence occurred in the early Paleocene and continued up to the early Miocene, but was interrupted by a phase of compressional deformation in the middle Eocene. The middle Miocene marked the settlement of the Andean-related foreland subsidence, and from the late Miocene onwards, the Llanos Basin underwent compression and inversion related to the development of the frontal fold and thrust belt of the Eastern Cordillera.

Sediment accumulation in the Llanos Basin followed a progressive eastward migration of the depocenter during the Cenozoic (Fig. 31), with the strata onlapping the Guiana Shield. As a consequence, the stratigraphic record is more complete in the western Llanos region. In most of the area, Upper Cretaceous strata directly overlie Paleozoic sedimentary and metamorphic basement.

The sediments of the Llanos Basin include a marine sequence of Cenomanian-Campanian age. The succession starts with shallow marine and shoreline sandstone beds at the base that form the Une Formation. Black shale beds, rich in organic matter, of Turonian-Early Coniacian age cover these basal units. There then follow beds of black shale, chert and phosphate layers, the Gacheta Formation. Regressive sandstone and shale beds of the Guadalupe Formation represent the end of this first transgressive marine sequence.

A second stratigraphic sequence was deposited during the late Paleocene, overlying the wide unconformity surface that marks the top of the underlying sequence (Cooper *et al*, 1995). This cycle is represented by beds of transgressive sandstone of the Barco Formation that grade to muddy coastal plain deposits known as the Los Cuervos Formation. The following transgressive cycle that started in the middle-late Eocene accumulated another succession with sedimentological style similar to the previous one, known as Mirador Formation, and including coastal, highly mature quartzarenite deposits.

From Oligocene times onward, four major cycles of marine-influenced coastal plain deposits were deposited, included in the Carbonera Formation (Cooper *et al.*, 1995), each bounded by surfaces of regional marine flooding. Onlap towards the Guiana Shield, the source area to the SE, made the units progressively sandier in that direction.

During middle Miocene time, compressional deformation in the orogenic belt increased, resulting in higher subsidence rates and in a rise of relative sea level, providing conditions for deposition of mudstone beds of the León Formation, and marking the Andean foreland basin phase (Fig. 32). In the late Miocene, a second source area was developed to the W, with the emergence of the Eastern Cordillera. The final phase of sedimentation in the Llanos Basin was the accumulation of molasse deposits. The deposition of some 3000 m of coarse grained continental clastic sediments, the Guayabo Formation, accompanied the positioning of the Andean front to its present situation.

Marañon-Ucayali-Acre Basin

The Marañon-Ucayali-Acre Basin, also known by the local names of the Putumayo, Napo, Santiago, Huallaga, Ene, and the Oriente Basin of Ecuador, is a very large foreland sedimentary province, comprising an area exceeding 900 000 km² (Fig. 28) and recording a long history of basin development that spans almost all the Phanerozoic Eon. A thick sequence of Paleozoic sedimentary rocks is included in the stratigraphic record of this area. The present day asymmetric geometry of the basin reflects late Cenozoic deformation. The western margin of the basin is marked by a line of thrust faults affecting Mesozoic sedimentary rocks that defines the eastern limit of the Andean Mountain Chain. Uplift of the Andean Belt has provided about 4000 m of continental clastics sediments to the basin. To the E, the Cretaceous to Cenozoic sedimentary sequence, as well as almost all the older sequences, wedge out over the Central Brazil Shield.

The Brazilian extension of this foreland domain, the Acre Basin, shows a structural framework defined by a set of N-S reverse faults and inverted normal faults (Oliveira,

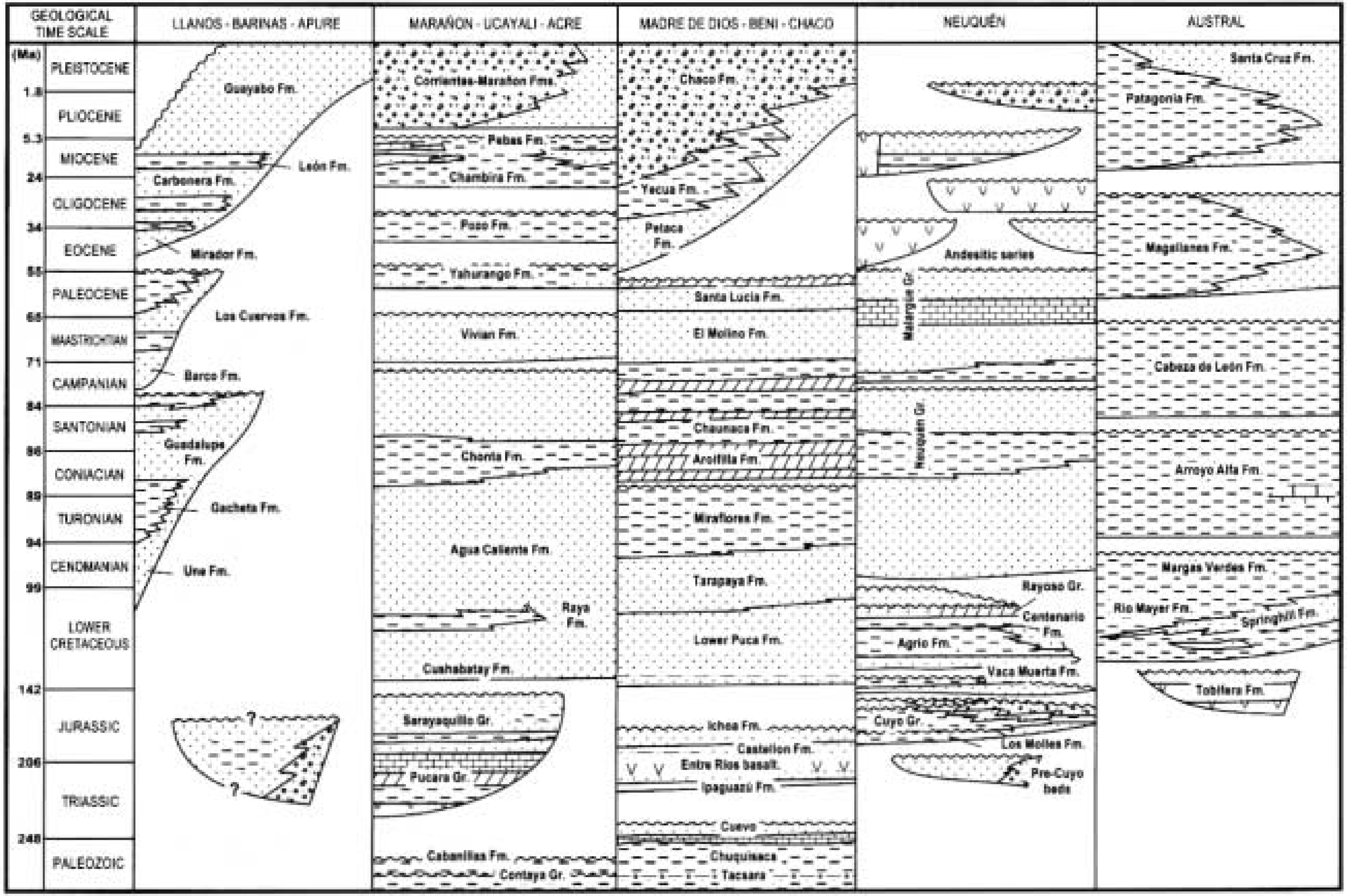


FIGURE 31 - Stratigraphic summary of the Andean foreland basins in South America. On each chart, the folded belt margin is to the left. Chart for the Marañon-Ucayali-Acre is representative of the western domain; towards the east, in the Acre Basin, most of the units pinch out over the Brazilian Shield. Main sources of information were Cooper et al. (1995), Llanos-Barinas-Apure Basin; Mathalone and Montoya (1995), Marañon-Ucayali-Acre Basin; Sempere (1995), Madre de Dios-Beni-Chaco Basin; Vergani et al. (1995), Neuquén Basin; Robbiano et al. (1995), Austral Basin. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.

1994). The most prominent of these, the Batã Fault defines a hingeline that is the eastern termination of the Paleozoic-to-Jurassic sedimentary sequence. The other important fault, the Oeste de Batã Fault, marks the western border of the depressed Acre Basin and extends to the foot of the Serra do Divisor to the W, the uplifted footwall of this west-dipping reverse fault. These hills attain significant topographical expression locally and form the international frontier between Peru and Brazil. In the depocenter of the Acre Basin the crystalline basement lies some 6000 m below the surface.

The Precambrian basement provided the major lines for Phanerozoic basin evolution in this area. Along the western margin of Gondwana, accretion of continental fragments designed a fabric of sutures that has been repeatedly reactivated through time. In the region of the Marañon Basin, a NW-SE trend of Precambrian structures defines the grain on which the basin evolved (Mathalone and Montoya, 1995). The oldest Paleozoic rocks that overlie the basement are the marine Ordovician strata up to 1000 m thick. These beds are known as the Contaya Group (Fig. 31), and include sandstone and shale beds, rich in organic matter. Silurian times were basically erosive. Sedimentation continued during the Devonian with the deposition of the dominantly shaly Cabanillas Formation, about 1600 m thick. The Chanic Orogeny of Late Devonian-Early Carboniferous age defined a regional lacuna in the stratigraphic succession of the Marañon Basin, as well as in several basins of South America (Urien *et al.*, 1995). To some extent the ancient lines of crustal weakness were reactivated during the Chanic Orogeny to cause subsidence and sediment accumulation.

The Carboniferous sequence started with the fluvial clastic deposits of the Ambo Group, including conglomerate, sandstone and shale beds with associated coal seams and volcaniclastic deposits. This is followed by the transgressive, mixed siliciclastic-carbonate sediments the Tarma Formation (Westphalian). The sediments evolved from the Westphalian trough Early Permian times into the large carbonate platform of the Copacabana Group, the sediments of which grade laterally to evaporite beds. The sea retreated during the Late Permian, and a sequence up to 600 m thick of shale, rich in organic matter, was deposited. These shale beds are assigned to the Ene Formation, and they accumulated under hypersaline conditions. By the end of the Permian, regional uplift provided abundant source areas for the continental clastic sediments of the Mitu Group, including volcaniclastic beds, that occupied extensional, fault-bounded depocenters (Mathalone and Montoya, 1995). Regional erosion and abundant volcanic activity took place in the area of the actual Andean Belt during Late Permian and Early Triassic.

Particularly in the Acre Basin of Brazil, the Paleozoic record includes a major unit, known as the Carboniferous-Permian Sequence (Feijó and Souza, 1994). The existence of a Devonian section in the Acre Basin is inferred from seismic data. The Carboniferous-Permian Sequence begins with the coarse-grained clastic wedge of the Apuí Formation, succeeded by beds of bioclastic calcarenite, white anhydrite, shale and calcilutite, known as the Cruzeiro do Sul Formation. These beds carry a palynological content of Early Permian age. A Late Permian section of grey sandstone and shale beds named the Rio do Moura Formation, closes the stratigraphical record of this sequence.

The Mesozoic history of sedimentation in the Marañon-Ucayali-Acre Domain started with the transgressive marine Pucara Group, spanning the range Late Triassic to Early Jurassic. A package of evaporite and carbonate beds with shale beds, rich in organic matter, accumulated in an asymmetric, deepening-to-the-west basin. These marine beds are covered by a section of continental red beds known as the Sarayaquillo Formation, of Jurassic age and, contrary to the geometry observed for the Pucara Basin, the Jurassic beds thin to the W indicating the uplift of Proto-Andean Mountains. In the Acre Basin, some 3000 m of sediments accumulated during the Jurassic, including fine-grained yellow sandstone units and red beds intercalated with beds of white anhydrite, halite and basalt flows, known as the Juruá-Mirim Formation. The pre-Cretaceous sedimentation ended with a regional unconformity related to the Nevadan event (Mathalone and Montoya, 1995).

The Cretaceous development in this area was recorded as a westward-thickening wedge of siliciclastic rocks including two major transgressive-regressive cycles of deposition. The Cretaceous sequence begins with the marine sandstone beds of the Cushabatay Formation. This basal unit is overlain by an Albian shaly package rich in organic matter, assigned to the Raya Formation and about 300 m thick. The Albian to Early Turonian Agua Caliente Formation consists of coarse-grained and fine-grained sandstone bed and fossiliferous black shales.

Major Cretaceous transgressive conditions were attained during Coniacian-Early Santonian times, when the shaly Chonta Formation was deposited. This unit is well developed in the northern part of the Marañon Basin of Peru and its prolongation in Ecuador, the Oriente Basin, where the black shale beds are called the Napo Formation. To the S and the E, the entire Cretaceous record consists of continental deposits. The Vivian Formation, a regressive unit of Campanian fluvial to coastal sandstone, closed the Cretaceous cycle of deposition. Maastrichtian, transgressive marine beds are of local occurrence, particularly in the Ucayali depocenter, and this is believed to have been controlled by regional erosion during the Peruvian Tectonic Cycle of Late Cretaceous age.

The Cenozoic history of this region accompanied the evolution of the Andean Fold and Thrust Belt, and received about 4000 m of continental sedimentary rocks (Fig. 33). These sediments are distributed as a westward-thickening wedge including the Yahuarango Formation (Paleocene), the Pozo Formation (upper Eocene-lower Oligocene), the Chambira Formation (upper Oligocene-lower Miocene), the Pebas Formation (upper Miocene) and the coarse-grained sediments of the Corrientes/Marañon formations (Pliocene). In the Acre Basin, Cretaceous to Cenozoic times added about 3000 m of continental clastic sediments to its stratigraphic record (Milani and Zalán, 1998).

Madre de Dios-Beni-Chaco Basin

This large part of the Andean foreland domain extends between the latitudes of 10°S and 30°S, including two segments about 1300 km long each. The northernmost segment strikes in NW-SE and the other, situated to the S, strikes N-S. They join at Santa Cruz deflection at the Bolivian

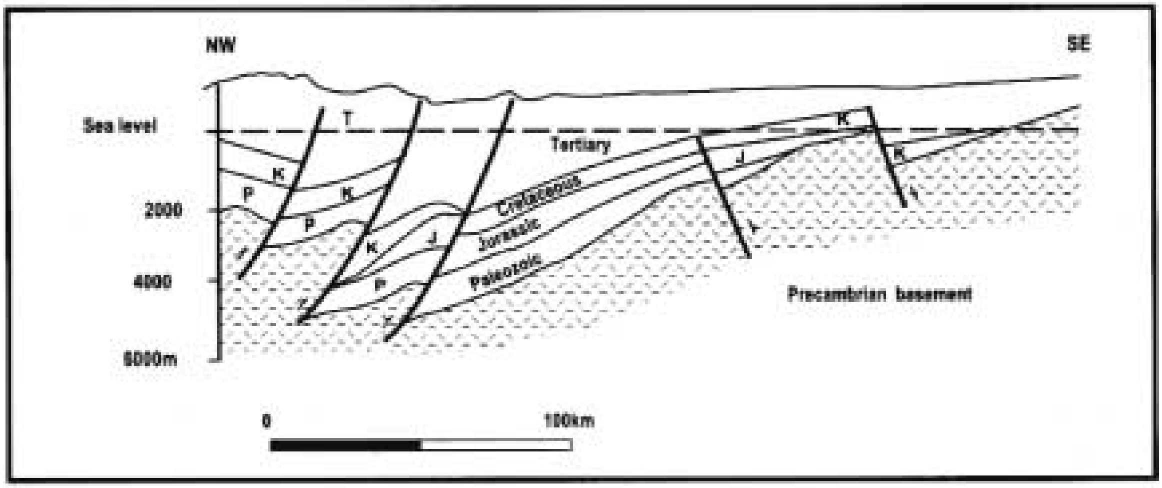


FIGURE 32 - Cross-section of the Llanos-Barinas-Apure Basin (after Alves, 1989).

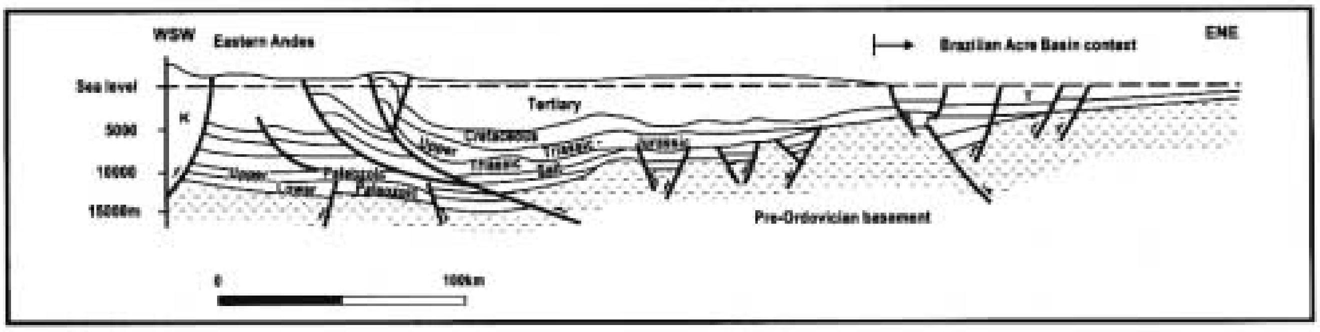


FIGURE 33 - Cross-section of the Marañon Basin (after J.M.P. Mathalone and M. Montoya R., 1995, AAPG Memoir Series No. 62, AAPG®1995, reprinted by permission of the American Association of Petroleum Geologists whose permission is required for future use).

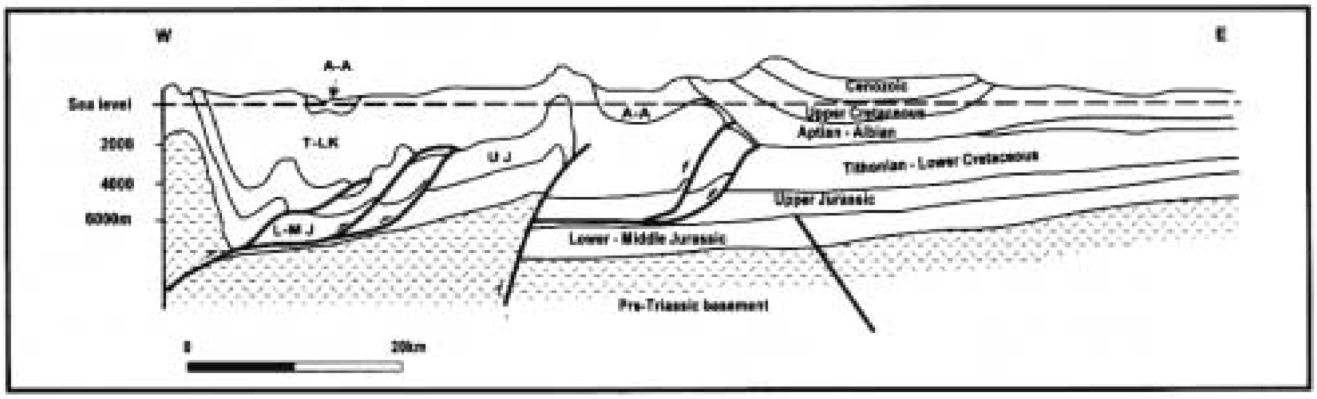


FIGURE 34 - Cross-section of the Neuquén Basin (after Uliana and Legarreta, 1993; permission to reproduce granted by the Journal of Petroleum Geology).

bend of the Andean front (Fig. 28). The total area of this province is about 700 000 km². In like manner to the Marañon Basin it includes a thick sequence of Phanerozoic sedimentary rocks showing a long history of subsidence that marked the southwestern margin of Gondwana.

The easternmost end of the Madre de Dios Basin occupies 10 000 km² of Brazilian territory, a region covered by Cenozoic continental clastic sediments and tropical forest. In that area, the maximum depth to the crystalline basement is around 1600 m, and the various sedimentary sequences, that are fully developed in the Bolivian sector, pinch out against the Central Brazil Shield (Milani and Zalán, 1998). The sedimentological record of the Madre de Dios Basin, both in lithology and in stratigraphic nomenclature, resembles that of the neighbouring Marañon-Ucayali-Acre Basin.

Towards the S, the Beni-Chaco Domain lies in a flat-lying sub-Andean region; the Jurassic to Cenozoic flexural depression developed over a basement of Paleozoic to Triassic rocks corresponding to the sequences that filled the Tarija Basin. Latest Jurassic was the time when large-scale rifting related to the beginning of the Andean system affected the region (Sempere, 1995). Continental red beds accumulated, constituting the lowermost part of the Puca Group (Fig. 31). This sedimentation was followed by a sag phase of subsidence up to the Turonian, when bioturbated shallow marine limestone units, rich in organic matter, were deposited in a flat basin during tectonic quiescence, constituting the Miraflores Formation. During the Late Cretaceous, the evaporitic-sandy-shaly sediments of the upper Puca Group accumulated, including the Aroifilla, Chaunaca, El Molino and Santa Lucia formations, the deposition of which was accompanied by relatively widespread alkaline basaltic volcanism.

The Cenozoic to Recent development of the Beni-Chaco region has been closely controlled by the evolution of the Andean Belt, the growth of which has provided the foredeep flexural subsidence to accommodate an up to 5000m of terrestrial beds up to Recent times. This section includes the sandy Petaca Formation, the shaly Yecua Formation and the thick sandy-conglomeratic molassic deposits of the Chaco Formation (Welsink *et al.*, 1995a).

Neuquén Basin

The Neuquén Basin covers an area of more than 160 000 km² of the western-central Argentina, and contains some 7000 m of sediments. It has a triangular shape (Fig. 28) and is limited to the W by the Andean Belt. This western side represented the region of linkage with adjacent Panthalassa during some of the several cycles of subsidence and evolution of the basin. The Neuquén Basin was in fact a local, cratonward reentrant in the large, N-S trending foreland domain along the active margin of Gondwana. During most of its history, the Neuquén Basin was situated between an evolving magmatic arc to the W and the cratonic regions of Gondwana to the E.

The develoment of the Neuquén Basin involved four major tectonic stages (Legarreta and Gulisano, 1989; Uliana and Legarreta, 1993). This started with a Late Triassic-Early Jurassic extensional collapse of a Late Paleozoic orogenic belt, followed by a thermal subsidence phase up to Late Cretaceous times. There then occurred a flexural subsidence stage caused by the load of the magmatic arc from Late Cretaceous to Early Cenozoic. Several episodes of basin inversion occurred, that modified to variable degrees the geometry of the depocenter and the relationship of the basin with the surrounding source areas. By Neogene times, compressional deformation resulting from the evolution of the Andes created the present day configuration of the basin.

The Neuquén Basin developed over a basement of magmatic, metamorphic and sedimentary rocks of Paleozoic to Triassic age. Early rifting occurred during Late Triassic times (Vergani *et al.*, 1995), and created a series of half-grabens that lodge the pre-Cuyo strata (Fig. 31), a thick sequence of continental coarse-grained sedimentary rocks and associated volcanic and volcaniclastic suites. Extensional conditions and normal fault controls over the sedimentation (Fig. 34) continued up to Early Jurassic and accommodated the lower beds of the Cuyo Group, represented by the marine shale beds, rich in organic matter, of the Los Molles Formation. Up to the beginning of Late Jurassic, a regressive sequence developed with provenance of fluvial-deltaic sediments from the E-SE, the post-Aalenian Cuyo section. This cycle culminated in an important tectonic event known as the Araucanian Inversion, during Late Oxfordian-Early Kimmeridgian.

During the Callovian, a N-S trending compressional belt developed in the western Neuquén Basin, known as the Dorsal de Huincul, and this was a preferential site for inversion of pre-existing rift-related structures. The Araucanian event was manifest as regional uplift and erosion, with deposition of syn-orogenic continental "molassic" beds of the lower Lotena Group in the adjacent depressions. At the end of the inversion episode, marine to evaporitic facies of the La Manga, Barda Negra and Auquilco formations accumulated in a wide sag basin formed by compressional release (Vergani *et al.*, 1995). A second event occurred during the Late Oxfordian-Earliest Kimmeridgian, reactivating older structures with intensity decreasing to the E. By the end of the Araucanian cycle of inversion, some 2000 m of strata had been removed by erosion from some parts of the Neuquén Basin.

The accumulation of the Mendoza Group followed the regional inversion phase. Sedimentation began with fluvial deposits associated with pyroclastic material, the Tordillo Formation. These fluvial clastic sediments grade to the eolian and fine-grained lacustrine rocks of the Catriel and Sierras Blancas formations. During the Tithonian, basin expansion and marine invasion provided accommodation space for a progradational unit that includes the Vaca Muerta Formation, a sequence of black shale beds, rich in organic matter (Mitchum and Uliana, 1985, *apud* Vergani *et al.*, 1995). The lower Mendoza Group terminates in Berriasian carbonate units known as Quintuco Formation.

Marine sedimentation that marked the previous cycle was partially arrested during Valanginian times, due to recurrent basin inversion. Mixed carbonate-siliciclastic sedimentation developed, comprising the Mulichinco and Chachao formations (Gulisano *et al.*, 1984). After the Valanginian, a phase of tectonic quiescence facilitated marine encroachment and the deposition of platform sediments, the Agrio Formation, that interfingers with the marginal, prograding fluvial siliciclastic sediments of the

Centenario Formation. This cycle of sedimentation continued up to Albian times, in a trend of progressive basin shallowing that allowed the accumulation of evaporite units and continental red beds of the Rayoso Group. The movements related to the Mirano inversion abruptly interrupted sedimentation.

This earliest Cenomanian phase of structural rearrangement reactivated surrounding source areas, resulting in the deposition of 1000 m of continental beds up to the end of the Cretaceous. The Neuquén Group is a Cenomanian to Campanian succession of fining and thinning upward units being mainly composed of fluvial deposits with provenance from the E. In the western part of the basin, volcaniclastic material is also present (Legarreta and Gulisano, 1989). The Late Campanian to Paleocene strata of the Malargüe Group completes the Riográndico Sedimentary Cycle. The sediments consist of marine and continental strata deposited in an epicontinental sea, with a maximum flooding event represented by the carbonate deposits of the Roca Formation (Late Maastrichtian to early Paleocene). Sedimentary provenance changed with respect to previous sequences. During the accumulation of the Malargüe Formation, the volcanic arc to the W became the major source of clastic material for the basin (Legarreta and Gulisano, 1989).

The post-Paleocene history of subsidence of the Neuquén Basin was driven by the load of the rising Andean Orogen, and was marked by the advance of the magmatic arc upon the foreland domain, with recurrent episodes of strong basement reactivation resulting in basin inversion. Thin-skinned tectonics also developed at this stage, with detachment levels within marine shale units of the Cuyo and Mendoza groups and within evaporite beds of the Rayoso Group (Vergani *et al.*, 1995).

Austral Basin

The Austral Basin, also named Magallanes Basin, corresponds to the southernmost part of the Andean foreland realm, and it covers an area of about 170 000 km^2 (Fig. 28). It is limited to the N by the Deseado High, a block of basement that separates the Austral Basin from the Golfo San Jorge Basin. To the E, the N-S trending Rio Chico High, or Dungeness Arch, defines the limit between the Austral and the Malvinas basins; towards the W, the Austral Basin faces the Andean Belt.

The basin developed from a Late Jurassic rift phase (Fig. 31), when NW-SE trending asymmetric grabens were opened and filled mainly with volcaniclastic sediments and associated lacustrine siliciclastic deposits of the Tobífera Formation. These grabens were concentrated around the Rio Chico High, the positive area of dissected Paleozoic basement that provided sediments for the basins. The rising of sea level, due to the climax of volcanic activity in the young marginal basin, allowed the deposition of transgressive sequences during Oxfordian-Kimmeridgian times, the sandy facies of which is named the Springhill Formation (Robbiano *et al.*, 1996). This succession is covered by a regressive sequence of fluvial and coastal sandstone beds of the Hidra Formation (Berriasian-Early Valanginian).

The cycle that followed showed progressively reduced magmatism. In the Austral Basin this is manifest as a period of sagging, with lower subsidence rates and the accumulation of a transgressive sequence known as Rio Mayer and Springhill formations (Late Valanginian-Early Aptian). During this cycle, a section of marine shale rich in organic matter was deposited, corresponding to the *Inoceramus* Inferior Formation of Hauterivian age.

Aptian times marked an important phase of structural modifications in the Austral Basin. The initial cycle of foreland subsidence, during the Cretaceous, was promoted by the Patagonides movements (Plozkiewicz and Ramos, 1978, *apud* Arbe, 1989). This brought about the closure of the marginal basin to the W of the depositional centre of the Austral Basin. Volcanic ash appears and this becomes more frequent in the stratigraphic record of this cycle, a sequence known as Margas Verdes Formation, that extended in time up to the Late Cenomanian.

During the rest of the Cretaceous, a complete transgressive-regressive cycle was deposited, including the dominantly shaly Arroyo Alfa and Cabeza de Léon formations. These formations include zones of marine shale beds, rich in organic matter. The top of this cycle is marked by a major unconformity that corresponds to the beginning of the subsidence of the Cenozoic foreland. From early Cenozoic times onward, the development of the North Scotia Ridge along the southern margin of the sag as part of the newly-formed active margin of the South American and Scotia plates, imposed a strong structural rearrangement on the Austral Basin. During these movements, in the early Paleocene, the southern part of the Austral Basin underwent strong E-W transpressive deformation (Robbiano *et al.*, 1996).

The Cenozoic record shows two transgressive-regressive cycles. The first form the Paleocene to Oligocene, and the second from the Miocene to Recent. A prominent pre-Miocene unconformity, related to the beginning of the Andean Orogeny, divides these two tectono-sedimentary cycles. The last stage of sedimentation corresponds mainly to the progradation of clastic sediments with provenance from the W and SW that may attain a thickness of 5000 m in the sub-Andean foredeep. On the northern flank of the basin, continental to shallow marine deposits mark the late Cenozoic progradation over the foreland ramp, constituting the Santa Cruz and Patagonia formations.

Back-arc extensional basins

This class of basins refers to the particular case of Mesozoic extensional troughs situated in the western, Andean Domain of South America. The development of back-arc rifts is attributed to periods of relaxation of compressional stresses of long duration, a conspicuous characteristic of the convergent margin of the continent. Compressional stresses along the western margin of South America have been active through most of the Phanerozoic (Urien *et al.*, 1995). Particularly important periods of relaxation were the Triassic (López-Gamundí *et al.*, 1994) and the Early Cretaceous (Ramos, 1988). Gust *et al.* (1985) included the Mesozoic events of extension observed in western South America into a single cycle that led to the opening of the South Atlantic.

Back-arc rifts were sites of local extensional subsidence that took advantage of the pre-existing compressional structural grain. This seems to have been the case for the Mesozoic basins of western Argentina. However, other authors explain the origin of some of these troughs by transtension along pre-existing structural features under persistent Mesozoic regional compression (Kokogian *et al.*, 1993).

In a regional context, the convergent setting in which these grabens developed is sharply contrasting with the fully extensional domain found on the Atlantic side of the plate, where the initial crustal rupture by rifting led to the evolution of a large passive margin and the growth of oceanic floor. The Mesozoic back-arc rifts were arrested by the Neogene compressional deformation and crustal shortening, and in fact, some of these became included in the Andean Chain.

Cuyo Basin

The Cuyo Basin is a NW-SE trending graben with an area of about 35 000 km^2 and containing about 7000 m thick of Triassic to Cenozoic sediments (Fig. 28). The Cuyo Basin is situated in the Mendoza and San Juan provinces of western Argentina, to the E of the Precordillera block, a region that holds evidences of a series of Phanerozoic orogenic episodes due to terrane accretion along the active margin of Gondwana (Ramos, 1990). Several other rift basins exist in the same domain, and some of them are the Ischigualasto, Marayes and San Luis basins, sharing with the Cuyo, the largest of these, the same general characteristics. To the W and NW, the development of the Triassic sequence of the Cuyo Basin was arrested by the Cenozoic structural compressional framework of the Andean Belt (Kokogian *et al.*, 1993).

Six sub-basins or depocenters were identified along the strike of the Cuyo Rift, each one corresponding to a segment of large downthrow in the adjacent border fault. The sub-basins are separated one from the others by transfer zones that define prominent basement highs (Ramos, 1992).

The Triassic record of the Cuyo Basin is a 3500 m thick sequence of continental sedimentary deposits with some intercalations of volcanic rocks (Fig. 35). The syn-rift sequence begins with a section of conglomerate and sandstone beds known as the Rio Mendoza Formation. These beds grade upward into the interval of fluvial-lacustrine sandstone and shale units of the Las Cabras Formation. Basaltic rocks, dated at 235 Ma (Ramos and Kay, 1991) occur intercalated with the upper sedimentary strata of the Las Cabras Formation. The syn-rift sequence terminates with a fining-upwards cycle of fluvial to lacustrine deposits, included in the Potrerillos and Cacheuta formations. The Cacheuta Formation represents a phase of severe starvation in the lacustrine system, and consists of black shale, rich in organic material of Late Carnian age (Moratello, 1993). The entire syn-rift section contains a well-developed tetrapod fauna (Báez *et al.*, 1993) that spans the range Scythian-Carnian.

With the decrease of normal fault activity, a phase of sagging took place and the post-rift sedimentation onlapped the previous deposits, expanding the depositional area of the Cuyo Basin (Fig. 36). The Norian to Rhaetian fluvial and lacustrine deposits of the Río Blanco Formation represent this stage (Kokogian *et al.*, 1993).

The accumulation of continental successions proceeded in the Cuyo area during post-Triassic times, reaching a thickness of about 3500 m of Jurassic to Cenozoic rocks. However, this uppermost sequence was no longer related to the former rifting, but rather to regional flexural subsidence in the Andean foreland domain. This last cycle of evolution includes basalt extrusions dated as Early Cretaceous, known as the Punta de las Bardas Formation (124 to 131 Ma, Legarreta *et al.*, 1993). These flows occur above the reddish conglomerate and sandstone beds of the Barrancas Formation, and the lavas are overlain by a conglomeratic-sandy-shaly sequence of Eocene age, corresponding to the Papagayos and Divisadero Largo formations (Kokogian and Mancilla, 1989).

Salta-Lomas de Olmedo Basin

The Salta-Lomas de Olmedo Basin is situated in the sub-Andean region of northwestern Argentina (Fig. 28), and covers an area exceeding 100 000 km^2. This rift system continues into Paraguay, where it is known as the Pirity Basin (Wiens, 1995). The extensional trough developed during Early Cretaceous to early Cenozoic times, and accommodated a package surpassing 5000 m of sedimentary and associated igneous rocks; being afterwards partially arrested by the eastward-advancing, thrusted front of the Andes.

Pre-existing, Late Proterozoic-Early Cambrian sutures along which the ancient plates were welded together influenced the positioning of the faulted basin. According to Ramos and Vujovich (1994), *apud* Comínguez and Ramos (1995), in the region of the Salta-Lomas de Olmedo Basin there exists a ENE-WSW trending, northward-dipping suture zone that controlled the dip of the main detachment fault of the Mesozoic rift. Thus the Salta-Lomas de Olmedo rift is an asymmetric trough displaying a similar dip of its basement towards the N.

The Cretaceous-Tertiary basin overlies a section of older sedimentary rocks (Fig. 35), including the Upper Proterozoic-Lower Cambrian turbidite deposits of the Puncoviscana Formation; the Cambrian-Ordovician siliciclastic platform strata of the Las Breñas Formation; and the Silurian-Devonian marine sequence of the Kirusillas, Copo, Santa Rosa and Icla formations, a section having a widespread occurrence along the western region of South America.

The Salta Group fills the Salta-Lomas de Olmedo Basin, and the syn-rift sequence is named the Pirgua Subgroup (Gómez Omil *et al.*, 1989), with ages ranging between the Valanginian and the Santonian. This sequence consists of fluvial deposits conglomerate, red sandstone and shale associated with alkaline basaltic rocks. Three major cycles of volcanism concurrent with the rift phase are described (Comínguez and Ramos, 1995). These are the Lower Cretaceous Cachipunco basalt (128-112 Ma); the Upper Cretaceous Las Conchas basalt (78-76 Ma), volumetrically the most important; and a third event, of Paleocene age (63-55 Ma). The volcanic centre migrated eastwards through time.

The lower part of the Pirgua Subgroup consists of alluvial fan facies of the La Yesera Formation, accumulated along the faulted margin of the rift system and interfingering with the fluvial, eolian, and playa lake facies of the Las Curtiembres Formation. Fault activity decreased with time,

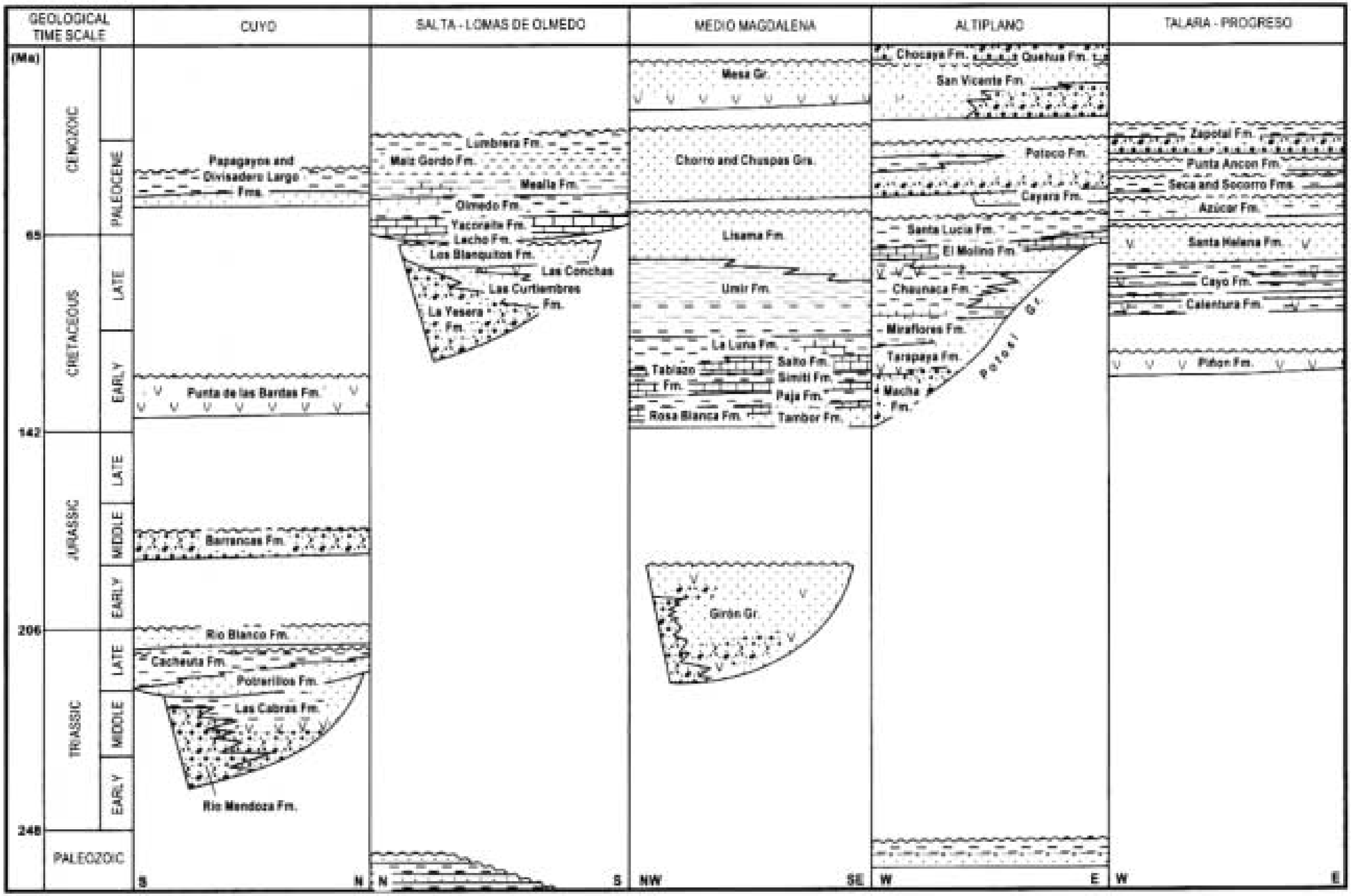


FIGURE 35 - Stratigraphic summary of the back-arc extensional basins, intermontane and fore-arc basins of South America. Main sources of information were Kokogian and Mancilla (1989), Cuyo Basin; Cominguez and Ramos (1995) Gómez Omil et al. *(1989), Salta-Lomas de Olmedo Basin; Schamel (1991), Medio Magdalena Basin; Welsink* et al. *(1995b), Altiplano Basin; Jaillard* et al. *(1995), Talara-Progreso Basin. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.*

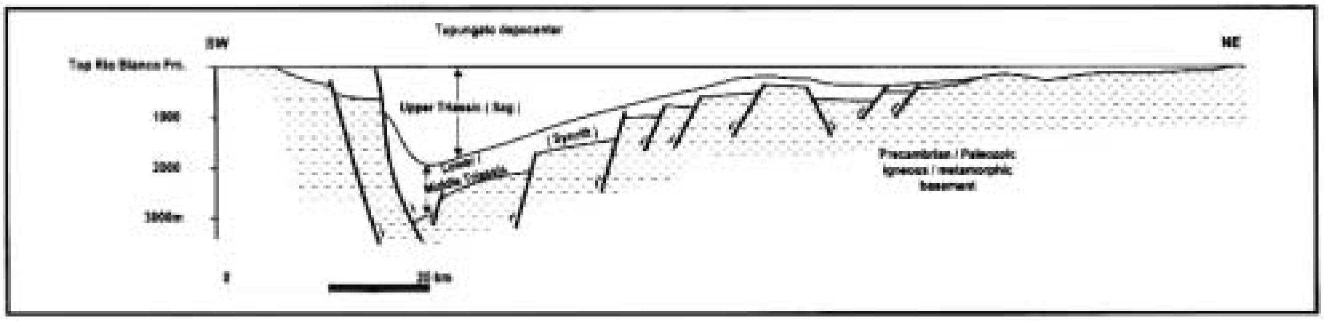


FIGURE 36 - Stratigraphic cross-section of the Cuyo Basin (after Kokogian et al., 1993; permission to reproduce granted by Asociación Geológica Argentina).

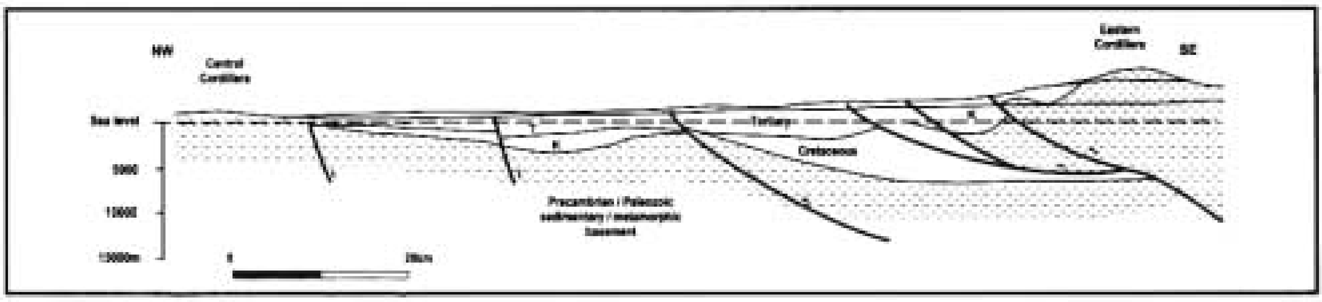


FIGURE 37 - Cross-section of the Medio Magdalena Basin (after S. Schamel, 1991, AAPG Memoir Series No. 52, AAPG® 1991, reprinted by permission of the American Association of Petroleum Geologists whose permission is required for future use).

and fine-grained deposits developed in the upper part of the rift fill, corresponding to the Los Blanquitos Formation.

The sag deposits of the Balbuena and Santa Barbara subgroups of Late Cretaceous to Eocene age succeeded the syn-rift sequence. These deposits have a total maximum thickness of about 2200 m. The post-rift section begins with the eolian beds of the Lecho Formation, covered by the mixed carbonate-clastic sediments of the Yacoraite Formation, of Campanian-Maastrichtian age. The reactivation of normal faults led to the emplacement of volcanic rocks associated with the Yacoraite beds, a magmatic cycle known as Palmar Largo. The uppermost unit of the Balbuena Subgroup is the Olmedo Formation, consisting of black shale and evaporite beds deposited in a hypersaline lacustrine environment.

The uppermost sequence of the sag phase is that of the Santa Barbara Subgroup, including, from base to top, the Mealla, Maiz Gordo and Lumbrera formations. These formations consist of shale, marlstone and subordinated carbonate beds, representing depositional systems related to ephemeral lakes, with fluvial sandy beds at their margins (Gómez Omil *et al.*, 1989).

Intermontane basins

De development of intermontane basins is a conspicuous characteristic of the Neogene history of the Andes Chain. They occur at several places along the mountain belt, from Venezuela to southern Argentina, and they formed according to two main types of mechanism, depending on the orientation of their bordering structures in relation to the regional stress field (Marocco *et al.*, 1995). The first was as strike-slip basins, undergoing cycles of syn-sedimentary deformation compatible with regional transcurrent motions. The second was by reverse faulting in response to compression.

The presence of sets of normal faults led to the interpretation of some of these basins as extensional half-grabens (Taborda, 1965, *apud* Schamel, 1991). However, according to Schamel (1991), extension was not the tectonic mechanism responsible for the development of these basins. The normal faults, as observed in some areas of the Magdalena Basin, are in fact the subordinate product of complex interactions between crustal loading by rising thrust sheets and transcurrent motions between adjacent blocks. In any case, the Andean intermontane basins underwent a continuous history of compressional deformation by progressive mountain building.

Medio Magdalena Basin

The Medio Magdalena Basin occupies an area of about 30 000 km² in northern Colombia (Fig. 28). In that area the Andean Belt is more than 600 km wide, or about twice of its mean size to the S, and inside this widened area there developed a Neogene successor basin along the valley that runs between the Eastern and the Central Cordillera. However, the previous history of the area witnessed much more extensive sedimentary domains, such as an extensional setting during the Triassic-Jurassic, a pericratonic trough during the Cretaceous and earliest Cenozoic, and a foreland basin during the middle Cenozoic (Schamel, 1991). Some of these pre-orogenic strata lie today in the highlands of the Eastern Cordillera, flanking the Neogene sequence that fills the NNE-SSW-trending intermontane basin.

The Mesozoic-Cenozoic sequence in the region of the

Medio Magdalena Basin (Fig. 35) was deposited over a complex of Paleozoic and Late Proterozoic metamorphic and sedimentary rocks, some of them pertaining to the domain of the Guiana Shield. A pronounced angular unconformity separates these older strata from the Mesozoic-Cenozoic successions. The basin exhibits an asymmetric transversal profile. Its eastern border is defined by a set of post-Miocene, west-vergent thrust faults that form the western border of the Eastern Cordillera (Schamel, 1991). The Central Cordillera block dips to the E and not only defines the western margin of the trough, but also underlies the Medio Magdalena Basin.

Triassic-Jurassic continental red beds with associated acid to intermediate volcanic rocks constitute the Girón Group, the basal unit of the Mesozoic section. These strata are widespread in the area. They occur filling NE-SW trending grabens, and may attain a thickness of 5 000 m. They record the phase of rift-drift in north-western South America (Pindell, 1985, *apud* Schamel, 1991) and the subsequent transition of the region to a subduction-related marginal basin (Maze, 1984, *apud* Schamel, 1991).

The Cretaceous sequence that represents a marine depositional megacycle, begins with Neocomian transgressive sandstone and conglomerate beds that rest unconformably over the Triassic-Jurassic sequence. These beds are known as the Tambor Formation. They are succeeded by a sequence of open marine limestone units and black shale beds up to 4000 m thick, included in the Rosa Blanca, Paja, Tablazo, Simiti and Salto formations. Maximum flooding was reached during Turonian-Coniacian times, and is recorded by the La Luna Formation, consisting of black shale, rich in organic material and limestone members. The drop of sea level during the Late Cretaceous and early Paleocene provided a renewed regressive cycle of sedimentation, with the deposition of the Umir and Lisama formations.

The marine Cretaceous beds were covered by a succession of three molasse sequences mainly consisting of alluvial, fluvial, and lacustrine facies that were accumulated in continental environments bounded by rising mountain chains. The basal sequence, corresponding to the Chorro and Chuspas groups, is more than 4 000 m thick (Schamel, 1991) and was fed by the rise of the Central Cordillera. The Real Group, of Miocene age succeeds it. Volcaniclastic deposits derived from the same, western uplifted area dominate the Real Group. The third molasse sequence, known as Mesa Group, was supplied with sediments from both sides of the valley, because the Eastern Cordillera was uplifted during Pliocene time (Fig. 37).

Altiplano Basin

The Altiplano Basin of Bolivia constitutes an endorheic fluvial, fluvioglacial and lacustrine depositional area of Cenozoic age situated about 4 000 m above the mean sea level, between the Western and Eastern ranges of the Chilean-Bolivian Andes (Fig. 28). It comprises an area of about 100 000 km^2, in part occupied by the Lake Titicaca, the largest body of water in the world at such a high altitude. The Altiplano Basin trends in a NW-SE orientation, and includes thick sequences of pre-Andean, Paleozoic and Mesozoic sedimentary rocks. Indeed, these sediments represent the remnants of a much larger preexisting sedimentary domain. In a tectonic perspective, the Altiplano is a contractional province originated during the Andean Orogeny (Welsink *et al.*, 1995b).

The western limit of the Cordillera Occidental, is a magmatic arc of Neogene to Recent age that developed on continental crust, the geology of which is poorly known (Roeder, 1988), but probably represents an eastward extension of the coastal Arequipa Massif (Ramos *et al.*, 1986). On the opposite side, the Cordillera Oriental is a fold-and-thrust belt involving Paleozoic sedimentary rocks that was affected by widespread granitic plutonism, with ages concentrated around the Late Triassic-Early Jurassic and the Miocene (Kontak *et al.*, 1984, *apud* Díaz-Martínez, 1995).

Precambrian rocks of the basement do not crop out in the Altiplano area, but they are known to exist in the subsurface because a reddish Late Proterozoic metagranite was sampled at a depth of 2800 m (Pareja *et al.*, 1978). The older known Paleozoic sedimentary rocks in the Altiplano Basin are Ordovician sandstone units of local occurrence (Fig. 35). Otherwise, Silurian rocks are widespread in the area and correspond to the diamictite and sandstone beds of the Cancañiri Formation. These beds are followed by a sequence of micaceous sandstone beds known as the Huanuni Formation; quartzose sandstone of the Llallagua Formation; shale in the Uncía Formation and terminating in a sandy-shaly sequence named the Catavi Formation.

The Devonian sequence includes two major transgressive-regressive cycles, starting with the basal Vila Vila Formation, consisting of quartzose sandstone, followed by a dominantly shaly section of the Belén Formation. The second cycle has a lower sandy unit found in the Huamampampa Formation, and an upper silty-shaly unit, the Colpacucho Formation. The Carboniferous system in the Altiplano area is represented by the Cumaná Formation, consisting of white sandstone and grey argillite beds related to a periglacial marine environment known as the Kasa Formation. These beds are followed by the Siripaca and Yaurichambi formations, of transitional environment, consisting of a sand and shale sequence up to 1000 m thick. The Copacabana Formation, that spans up to the Early Permian, consists of mixed carbonate-siliciclastic marine sediments (Díaz-Martínez, 1995).

The Cretaceous sequence is more than 3300 m thick in the southern part of the Altiplano Basin (Pareja *et al.*, 1978), and it unconformably overlies different Paleozoic units. It is represented by the Potosí Group, a unit characterized by rapid facies changes due to an uneven morphology of the depositional site. The Cretaceous sequence is interpreted as representing the in-fill of a rift trough and succeeding thermal sag (Welsink *et al.*, 1995b). It consists of sandstone, shale, limestone and evaporite beds that, in some places, exhibit diapiric structures. Magmatic events interrupted the Cretaceous sedimentation during the Barremian and Early Campanian.

During the Cenozoic, true intermontane conditions were attained in the Altiplano Basin following the uplift of the adjacent ranges, and the deposition of a sequence exceeding 6000 m of syn-orogenic sandstone, shale, limestone, and evaporite beds, intercalated with volcaniclastic and volcanic rocks. These rocks are included in the Cayara, Potoco, San Vicente, Quehua and Chocaya formations. Quaternary deposition in the Altiplano Basin filled the basin to its present limits. The sedimentation is represented by some 600 m of volcaniclastic beds, lava flows, conglomerate and sandstone beds, lying sub-horizontally. Cenozoic magmatism is abundant and expressed as intrusions, lava flows and ignimbrite.

Fore-arc basins

Fore-arc basins formation and sediment accumulation occur in several places along the convergent margin of South America. Such basins developed by rapid extensional subsidence between the trench slope break of the accretionary wedge and the magmatic front of the arc, being generally overlying trapped fragments of oceanic crust older than both the magmatic arc and the accretionary complex. To some extent, such rigid basement also prevented these basins from undergoing further intense compressional deformation as that observed in the concurrent sedimentary accretionary prism. Adjacent volcanic arches provided important amounts of detritus to these basins.

Among the main fore-arc sedimentary basins along the Andean margin of South America there are the Chiloé, Valdívia, Temuco, and Mataquito basins in Chile; Mollendo, Pisco, Lima, Trujillo, and Talara in Peru; Progreso, Golfo de Guayaquil, and Manabi in Equator; and Borbon in Colombia. The region of Talara-Progreso was selected here to represent this class of basin.

Talara-Progreso Basin

The Talara-Progreso Basin is situated in the coastal region of southern Ecuador and northern Peru (Fig. 28), covering an area of about 35 000 km^2 around the Gulf of Guayaquil. It is a Miocene fore-arc trough developed on a complex Cenozoic collisional tectonic setting. The region is underlain by a piece of oceanic basement (Mégard, 1987, *apud* Jaillard *et al.*, 1995), consisting of massive tholeiitic basalt and basalt-andesite lava of the Piñon Formation. These rocks have radiometric ages around 110 Ma (Goossens and Rose, 1973), and were accreted to the South American Plate in late Paleocene-early Eocene times.

The initial tectono-sedimentary history of the region includes an accumulation of the Cenomanian to Early Coniacian pelagic marine deposits of the Calentura Formation (Fig. 35). These deposits are mainly volcaniclastic and accumulated in a marginal basin overlying the oceanic crust of Aptian-Albian age. The following tectono-sedimentary cycle was marked by the presence of an adjacent island arc providing coarse-grained detritus to the marginal depression, a sequence of Santonian-Campanian age known as Cayo Formation. The activity of the Cayo Volcanic Arc diminished with time, and the sedimentary sequences become finer towards the top.

The subsequent development of the Talara-Progreso area is marked by recurrent episodes of collision and uplift followed by extensional subsidence in the fore-arc domain, creating a pattern of short-lived fore-arc basins interrupted by periods of erosion (Jaillard *et al.*, 1995). This genetic relationship between compressional phases succeeded by extensional subsidence was explained by tectonic erosion along the plane of subduction, responsible for a loss of mass along the continental margin that is believed to be enough to trigger the subsidence of the fore-arc domain when the compression stresses decreased (Von Huene and Lallemand, 1990; Von Huene and Scholl, 1991).

It was probably that, in the late Paleocene-early Eocene times, the remnants of the Cayo Arc collided with the margin of the continent, resulting in the emergence of the region and a large influx of coarse-grained, quartz-rich sediments to the adjacent slope basin, a sequence named the Azúcar Group. By the end of this compressional phase in the middle Eocene, fore-arc extensional subsidence took place, accommodating the marine strata of the Socorro and Seca formations, deposited in an outer shelf environment. The Socorro Formation consists of laminated shale beds, siltstone, and fine-grained sandstone. Near the base of the formation, slumped beds are common, and these define the Clay Pebble facies. Upwards, the Socorro Formation grades into a sequence of laminated shale beds, siltstone, marlstone, and thin-bedded sandstone units known as the Seca Formation. The upper part of the Eocene sequence includes a section of continental to shallow marine greywacke and lithic sandstone units known as Punta Ancón Formation. Alluvial coarse-grained sandstone and conglomerate assigned to the lower part of the Zapotal Formation succeed these beds.

From Oligocene time onwards, several fore-arc basins developed in the area, including the individual troughs of Talara-Progreso and the Golfo de Guayaquil basins. These originated from the recurrent compression-to-extension changes in the regional stress field. These basins were filled with fine-grained shallow marine sandstone and shale that in part correspond to the upper part of the Zapotal Formation, dated by its planktonic foraminifera content as Oligocene-early Miocene (Jaillard *et al.*, 1995). In Quaternary times most of coastal Ecuador-Peru area emerged.

Northern and Southern, Transform Margins

Along the Andean margin of South America, persistent convergence between the continental plate and the Pacific oceanic lithosphere originated a tectonic environment that has been under net compression; being a mountain belt framed cratonward by an almost continuous foreland depression one of the most obvious result of this convergent zone of widespread crustal shortening.

The tectonic behaviour along the Caribbean and Scotian sides of South America (Fig. 1) is different. Towards the N and the S, the subduction zone on the western border of the continent changes into transpressional systems. The consequence is that oblique convergence has been the dominant way of interaction between the South American and adjacent Caribbean and Scotia plates from Paleocene time onwards, defining large-scale transform boundaries, and, in this way, allowing a complex changing pattern of strike-slip motion, local crustal shortening, and lithosphere

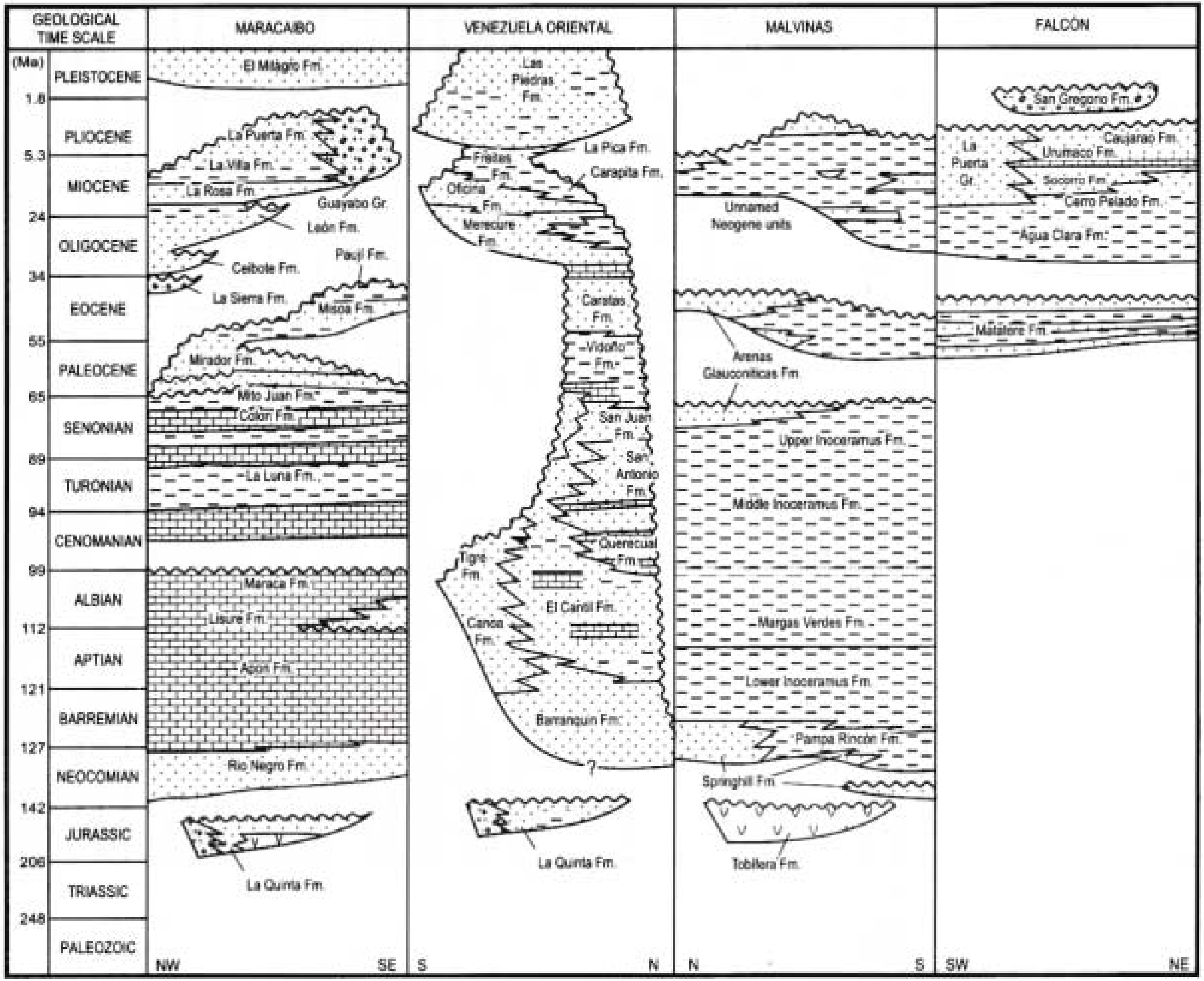


FIGURE 38 - Stratigraphic summary of the basins situated along the northern and southern transform margins of South America. Main sources of information were Parnaud et al. (1995b), Maracaibo Basin; Parnaud et al. (1995a), Venezuela Oriental Basin; Galeazzi (1996), Malvinas Basin; Macellari (1995), Falcón Basin. Geological time scale simplified from Gradstein and Ogg (1996). Lithological representations are of common use.

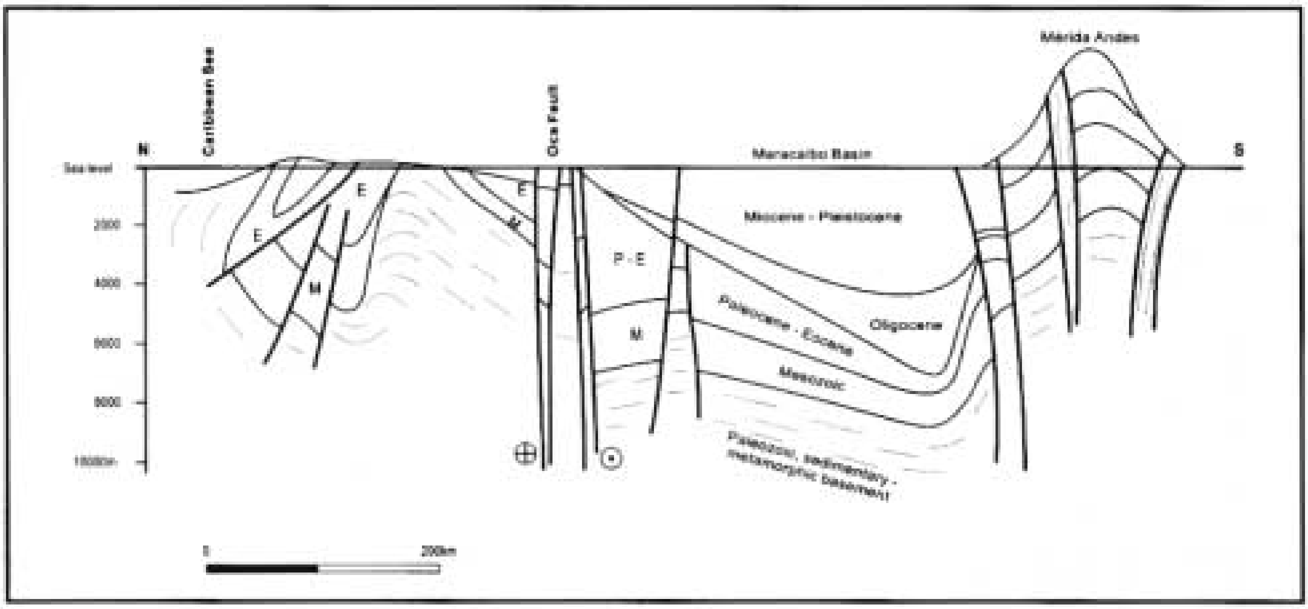


FIGURE 39 - Cross-section of the Maracaibo Basin (after Perez de Mejía et al., 1980; reproduced by courtesy of Schlumberger).

flexure by tectonic loading as the basic mechanisms of basin formation and inversion. Also classic pull-apart grabens developed at some places, related to releasing bends of these transform margins, Oca-San Sebastian-El Pilar to the N and North Scotia to the S.

Oblique flexural basins

Maracaibo Basin

The Maracaibo Basin covers an area of about 50 000 km^2 in the northwestern region of Venezuela (Fig. 28). In its northernmost part, the Andean Belt deflects to the E to become about three times wider than its mean width in the southern region. This is a very complex region that displays thin-skinned tectonics (Kellogg, 1984) and microplate movement that depicts an intricate pattern of individual movements and interaction of blocks.

The Maracaibo Basin developed in the center of a triangular lithosphere wedge, the Maracaibo-Santa Marta Block that is being pushed away from South America by compressional stresses. The movement of the Maracaibo Block towards the N is bounded by two major strike-slip zones, the right lateral Boconó Fault and the left lateral Santa Marta-Bucaramanga Fault (Lugo and Mann, 1995). To the SE the Mérida Andes mountain range separates the Maracaibo region from the sub-Andean foreland domain of the Barinas-Apure Basin. Towards the NW the Maracaibo Basin is limited by the Perijá Mountains.

Three major tectono-sedimentary stages are recognized in the geological history of the Maracaibo Basin (Fig. 38): Jurassic rifting, related to the separation between the North and South American plates, followed by a Cretaceous phase of passive margin opening to the N-NE. This stage was interrupted by the collision between the Caribbean and the South American plates during Paleocene/Eocene times that started a cycle of foreland basin development. Flexural subsidence continued from Eocene time onwards, driven by the oblique collision and development of the transpressional northern boundary of the South American Plate (Lugo and Mann, 1995). About 7000 m of Jurassic to Recent sedimentary rocks have been deposited in the basin, and of these the uppermost 3000 m correspond to Neogene coarse-grained rocks derived from surrounding Andean relief.

The Maracaibo Basin overlies Paleozoic sedimentary and metamorphic rocks (Fig. 39). Rifting took place during the Jurassic, and accommodated continental, alluvial to lacustrine red beds, the La Quinta Formation, and associated alkaline volcanic rocks within SW-NE to N-S trending troughs. Dominantly carbonate sedimentation developed during the Cretaceous passive margin stage (Eva *et al.*, 1989), over the basal transgressive clastic deposits of the Rio Negro Formation of Neocomian-Aptian age. Carbonate accumulation started with the Aptian shallow marine shelf sedimentation of the Apon Formation, and followed with the Lisure and Maraca formations, a section of mixed siliciclastic-carbonate rocks, and continued with the deposition of Cenomanian fossiliferous limestone members. The Mérida Arch, probably originated as an uplifted faulted block during the rift stage, and persisted as a positive feature controlling, to some extent, the nature of facies that filled the passive margin basin (Macellari, 1988).

During the Turonian and Santonian, at the time of maximum eustatic sea level of the Cretaceous (Vail *et al.*, 1977), a sequence of black shale beds, rich in organic matter, marlstone and limestone units accumulated in the Maracaibo Basin and surrounding areas. The La Luna Formation, together with their stratigraphically equivalent units in adjacent basins, constitute the most important known source rock on the continent, accounting for about 80% of the already discovered hydrocarbons (Mathalone, 1996). Volcanic ash beds occur at the base of the La Luna Formation, indicating the existence of an active volcanic arc to the W (Parnaud *et al.*, 1995b), along the subduction-related Pacific margin of the continent. The passive margin megasequence of the Maracaibo Basin ended with the deposition of Campanian to Maastrichtian micritic limestone beds and shale known as Colon Formation.

During the Late Cretaceous to early Paleocene, renewed tectonic activity due to the collision of the Pacific Volcanic Arc against the South American Plate gave origin to the foreland basin phase in the Maracaibo area. This change occurred in a diachronous mode over the former passive margin, from W to E, accompanied by a regressive phase of sedimentation. The Mito Juan Formation, reflecting highstand progradational depositional systems had its provenance in the W, and accumulated during this passive margin to foreland basin transitional phase. The progressive emplacement of nappes to the N of the Maracaibo Basin accelerated the flexural subsidence in the area, from late Paleocene time onwards. Marine sequences continued to accumulate in the Maracaibo Basin up to middle Eocene, when the basin underwent a phase of regional erosion. The sea regressed by late Oligocene times, and inundated widely up to the Miocene. The resulting deposits are known as the León Formation.

The Plio-Pleistocene marked the uplift of the Mérida Andes ranges, and the isolation of the Maracaibo Basin from the southeastern sedimentary domain, the Barinas-Apure Basin. The sea progressively retreated after the transgressive phase of the middle Miocene that resulted in the accumulation of the La Rosa Formation. Molasse sedimentation of the Guayabo Group invaded the basin from the Mérida High, indicating rapid subsidence along the orogenic front (Parnaud *et al.*, 1995b), accompanying the Neogene development of the Maracaibo Basin.

Venezuela Oriental Basin

The Venezuela Oriental Basin is situated in northernmost South America, covering an area of around 280 000 km^2, including the Maturín and Guárico sub-basins. Further subsidence developed during Neogene times as a foreland depression developed. This foreland depression is related to the evolution of the E-W trending Serrania del Interior Fold and Thrust Belt, and it is limited to the N by the orogenic belt itself, and by the right-lateral transcurrent El Pilar Fault. The El Pilar Fault marks the active limit between South America and the Caribbean plates (Fig. 28).

Four major stages are recognized in the Phanerozoic tectonic evolution of the area of the Venezuela Oriental Basin (Eva *et al.*, 1989). These are a Paleozoic pre-rift phase; a cycle of rifting-drifting during Jurassic and earliest Cretaceous; a passive margin phase spanning the Cretaceous-Paleogene interval; and the Neogene foreland basin stage. Pre-existing sedimentary sequences, up to 5000 m thick, form the so-called pre-rift phase. These are coastal to neritic marine Devonian and Carboniferous sediments (Stover, 1967, *apud* Parnaud *et al.*, 1995a), probably representing remnants of the widespread cratonic sequences that covered large areas of South America and constitute, in fact, the sedimentary basement for the rift sequences in this area (Fig. 38).

Rifting in the Venezuela Oriental Basin took place during Late Jurassic to Early Cretaceous times, and is represented mainly by the continental red beds and diabase sills of the La Quinta Formation. This package, an onlapping megasequence that fills a series of half-grabens, may reach a thickness of about 3600 m (Parnaud *et al.*, 1995a). Passive margin conditions were established in the Barremian, and prevailed up to the Paleogene. The sediments show three major transgressive phases. The passive margin megasequence that started with the deposition of coastal sandstone beds known as the Barranquin Formation was followed by the shallow marine beds of the El Cantil Formation and ended in a major transgression during the Turonian. Two other marine transgressions peaked in the Paleocene-early Eocene times. The first is represented by the Vidoño Formation, and the second by Areo Formation of Oligocene age. Marine shale beds, rich in organic material, were deposited during the Cenomanian to Campanian incursion of the sea, and correspond to the Querecual and San Antonio formations (Alberdi and Lafargue, 1993). To the S, the marine basin was delimited the continental sandy successions of the Canoa and Tigre formations.

The convergence of the Caribbean Plate against South America during the Oligocene terminated the passive margin phase in the Venezuela Oriental Basin, and started its foreland cycle (Fig. 40). The oblique collision megasequence (Parnaud *et al.*, 1995a) is a syn-orogenic wedge up to 6000 m thick. Deposition accompanied the eastward migration of the foredeep and the diachronous evolution of the transpressional orogenic belt related to the El Pilar transform margin and adjacent Serrania del Interior Thrust and Fold Belt, the major source of sediments during the foreland cycle. Thin-skinned tectonics developed using as detachment surfaces the Jurassic-Cretaceous interface and the boundary between the Mesozoic-Paleogene and the Neogene.

The foreland megasequence consist of Oligocene-Miocene sandy sediments that prograded from the S, included in the Merecure and Oficina formations. In the deep basin, a thick sequence of shales with turbiditic sandstone beds at the base was deposited. This sequence is knows as the Carapita Formation (Parnaud *et al.*, 1995a). Shallow marine and continental deposits filled the foredeep trough during the Pliocene and Pleistocene, defining the La Pica and Las Piedras fomations.

Malvinas Basin

The Malvinas Basin is a tectonic province developed in the southernmost region of the continent, adjacent to the E-W trending, left-lateral transpressional North Scotia Ridge, the active element of contact between South American and Scotia plates. The basin has a triangular shape with an area of more than 100 000 km^2 entirely in the offshore domain of southern Argentina. It contains a sequence some 7000 m thick of Mesozoic-Cenozoic rocks. It is separated from the Austral Basin, to the W by the N-S trending Dungeness Arch, and its strata onlap the Malvinas (Falklands) Platform to the E-NE (Fig. 28). The regional tectonic pattern of the basin is given by a relatively simple northern domain, where normal faults dominate, and a complex, transpressional southern border, related to the transcurrent limit of plates.

The Malvinas Basin evolved as a non-conventional foreland basin (Galeazzi, 1996) in a particularly complex tectonic domain, right where the thrust and fold belt of the Andes meets the southern transform margin of South

America. This condition provided a changing pattern of flexural subsidence during time that alternated with episodes of transpressional deformation and basin inversion.

The pre-Jurassic basement for the sedimentary successions in the Malvinas Basin is of continental nature and mainly represented by low grade metamorphic rocks and granite, the lithological context of the Panthalassan margin of Gondwana deformed by compression during the Permian to Early Triassic Cape-La Ventana Orogeny (Veevers *et al.*, 1994). Rifting in the Malvinas Basin occurred during the Jurassic (Fig. 38), in a back-arc tectonic setting, followed by thermal subsidence that spanned the range Late Jurassic-Maastrichtian. The presence of the Dungeness Arch as a positive feature, associated with depocenters adjacent to normal faults, exerted important control upon the distribution of strata during these initial phases of evolution. The rift section consists of tuffs intercalated with lavas (Yrigoyen, 1989), a package that may reach a thickness of 800 m and is known as Tobífera Formation (Thomas, 1949, *apud* Galeazzi, 1996).

During Paleocene and Eocene times, transtensional deformation took place in the southwestern sector of the basin and, from late Eocene time onwards, the basin underwent foreland flexural subsidence driven by the development of the transpressional North Scotia Ridge to the S and the Patagonic Orocline to the SW (Fig. 41). A thrust and fold belt appeared right to the S of the basin, imposing extreme changes on its shape, subsidence history and sedimentary patterns.

The post-Tobífera stratigraphic record of the Malvinas Basin begins with a Late Jurassic-Neocomian transgressive sequence related to the northwards flooding of the basin by the waters of the Weddell Sea. This retrogradational to aggradational section, included in the Springhill and Pampa Rincón formations, consists of deltaic to coastal sandstone beds, argillite and coal seams of coastal plain environments; and middle to distal platform carbonate units and shale. These units filled a wide sag related to the thermal subsidence phase that continued up to the end of the Cretaceous (Galeazzi, 1996). Drowning of the depositional platform continued, and an open marine sedimentation represents the Aptian to Maastrichtian section, a dominantly shaly succession named the *Inoceramus* and Margas Verdes formations.

Glauconitic sandstone and shale of the Arenas Glauconíticas Formation, accumulated in a distal shelf environment marks the uppermost Cretaceous to lower Eocene interval. By the end of the Eocene, the Malvinas Basin underwent a regional transpressional tectonic regime and assumed definitively its oblique foreland character. The post-Eocene stratigraphic record of the Malvinas Basin includes a southward thickening wedge of Oligocene deep marine deposits followed by a prograding, fluvio-deltaic siliciclastic sequence of Neogene age.

Pull-apart basins

Falcón Basin

The Falcón Basin is situated in northwestern Venezuela, where it covers an area of 15 000 km². This trough originated and developed as a pull-apart basin related to the dextral strike-slip motion between the Caribbean and South American plates along the large E-W trending Oca-San Sebastian-El Pilar Transform Zone (Fig. 28). The basin started to develop as a pull-apart trough in the early Cenozoic (Muessig, 1984, *apud* Boesi and Goddard, 1991), over a basement of igneous, metamorphic rocks and, locally, remnants of Mesozoic and Paleozoic sedimentary successions.

Late Paleocene to middle Eocene times were marked by the accumulation of Caribbean Plate-derived turbidite beds known as the Matatere Formation (Fig. 38) in a large depression that soon would turn into the pull-apart Falcón Basin. During the late Eocene, right-lateral transcurrent movement along the northern, transform margin of South America probably activated a transtensional depocenter along a releasing bend in northwestern region of Venezuela. As a consequence, a series of horsts and grabens, bounded by NW-SE trending normal faults appeared. These faults were arranged in a right-stepping *en échelon* pattern and defined the depocenters of the nascent pull-apart Falcón Trough.

In the Miocene-Pliocene interval, the basin underwent a strong transpressive phase of inversion (Fig. 42), with the creation of a belt of SW-NE trending anticlines along the southeastern border of the basin (Boesi and Goddard, 1991). The continued tectonic evolution of the region included diffuse right-lateral transcurrent displacement along an E-W striking fault system. Such movement progressively concentrated in the Oca Fault, defining a segment of the present-day active northern margin of the South American Plate. Due to mountain building, large volumes of sediments previously deposited in the Falcón Basin were removed during this late phase of structural inversion, and transported to the Caribbean Sea.

Sedimentation in the Falcón Basin reflects two main stages (Díaz de Gamero, 1977, *apud* Boesi and Goddard, 1991): an initial transgressive phase from late Eocene to middle Miocene, in which the sea invaded from the NE with the resulting accumulation of shale beds, rich in organic material, included in the Água Clara and Cerro Pelado formations; and a regressive phase, from late Miocene onward, as a consequence of basin inversion that caused a marine regression toward the northeastern border of the basin. Sedimentation during the regressive phase was dominantly sandy with the deposition of carbonate units in the western and southwestern parts of the basin, where the La Puerta Group developed, and sandy in the central domain, the realm of the Urumaco and Codore formations. The marginal deposits grade to finer facies towards the central and eastern parts of the basin, where developed a calmer depositional site. These distal deposits are known as the Socorro and Caujarao formations. Pliocene molasse known as San Gregorio Formation, covered large sectors of the Falcón Basin.

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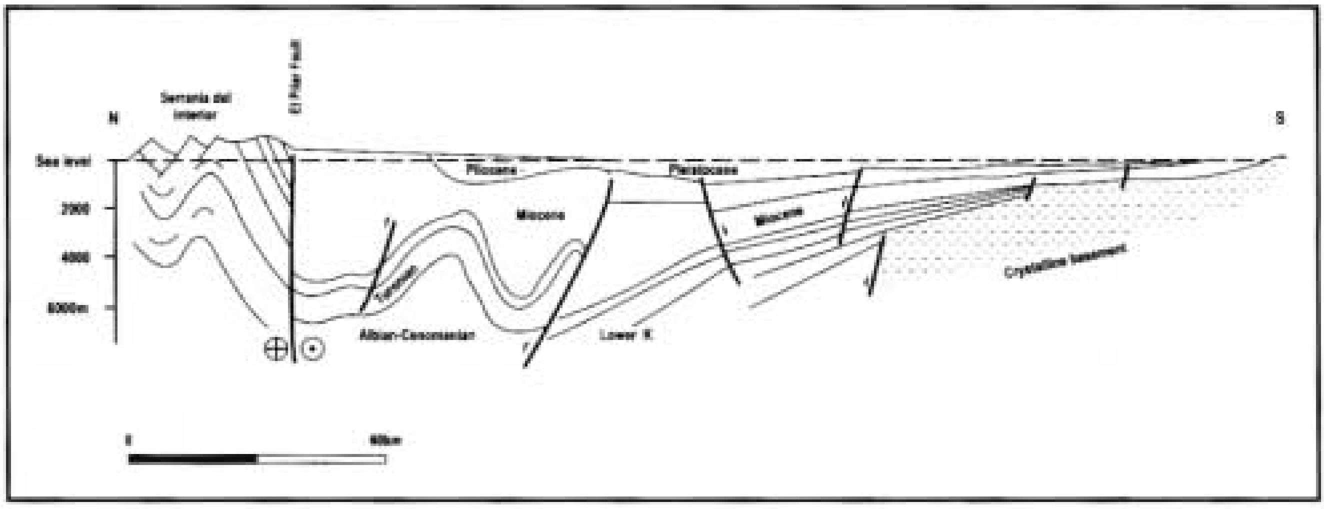
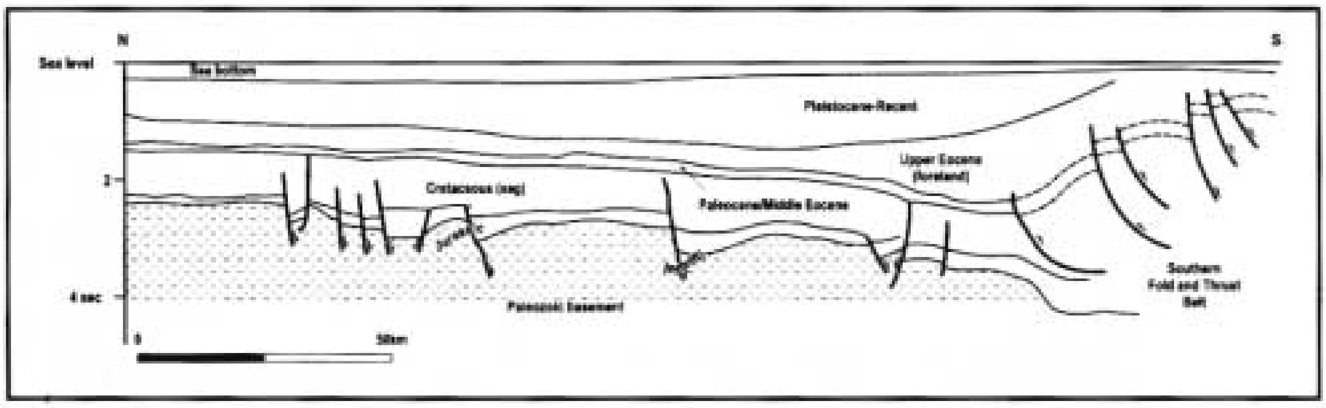




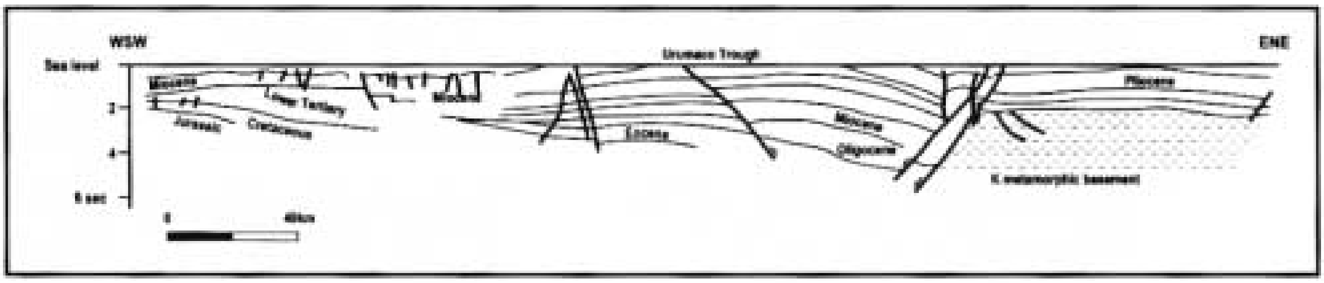


FIGURE 40 - Cross-section of the Venezuela Oriental Basin (after Perez de Mejía et al., 1980; reproduced by courtesy of Schlumberger).

FIGURE 41 - Geoseismic cross-section of the Malvinas Basin (after Galeazzi, 1996; permission to reproduce granted by Asociación Geológica Argentina).

FIGURE 42 - Geoseismic cross-section of the northwestern part of the Falcón Basin (after C.E. Macellari, 1995, AAPG Memoir Series No. 62, AAPG® 1995, reprinted by permission of the American Association of Petroleum Geologists whose permission is required for future use).

REFERENCES

Adighije, C.I. (1981). A gravity interpretation of the Benué trough, Nigeria. *Tectonophysics*, 79, 109-128.

Alberdi, M. and Lafargue, E. (1993). Vertical variations of organic matter content in Guayuta Group (Upper Cretaceous), interior mountain belt, eastern Venezuela. *Organic Geochemistry*, **20**, 425-436.

Albertão, G.A. (1995). Abordagem interdisciplinar e epistemológica sobre as evidências do limite Cretáceo-Terciário, com base em leituras efetuadas no registro sedimentar das bacias da costa leste brasileira. *Boletim de Geociências da Petrobras*, 9(2/4), 321-322.

Almeida, F.F.M. and Carneiro, C.D.R. (1998). Origem e evolução da Serra do Mar. *Revista Brasileira de Geociências*, **28**(2), 135-150.

Almeida, F.F.M and Hasui, Y. (1984) *O Pré-Cambriano do Brasil*. Edgard Blücher, São Paulo, Brasil.

Almeida, F.F.M., Hasui, Y., Neves, B.B.B., and Fuck, R.A. (1977) Províncias estruturais brasileiras. *Anais do I Simpósio de Geologia do Nordeste*, Campina Grande, Brazil, 363-391.

Alves, R.J. (1989). Bacias sedimentares: potencial petrolífero da América do Sul. *Boletim de Geociências da Petrobras*, 3(4), 321-336.

Aranha, L.G.F., Lima, H.P., Souza, J.M.P. and Makino, R.K. (1990). Origem e evolução das bacias de Bragança-Viseu, São Luís e Ilha Nova. In: *Origem e evolução de bacias sedimentares*, eds. Raja Gabaglia, G.P and Milani, E.J., pp. 221-233. Petrobras, Rio de Janeiro, Brasil.

Araripe, P.T. and Feijó, F.J. (1994). Bacia Potiguar. *Boletim de Geociências da Petrobras*, 8(1), 127-141.

Arbe, H.A. (1989). Estratigrafia, descontinuidades y evolución sedimentaria del Cretaceo en la Cuenca Austral, Provincia de Santa Cruz. In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 419-442. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Asmus, H.E. (1984). Geologia da margem continental brasileira. In: *Geologia do Brasil*, eds. Schobbenhaus, C., Campos, D.A., Derze, G.R., and Asmus, H.E., pp. 443-472. Departamento Nacional da Produção Mineral, Brasília, Brasil.

Asmus, H.E. and Ponte, F.C. (1973). The Brazilian marginal basins. In: *The ocean basins* and *margins*, eds. Nairn, A.E.M. and Stehli, F.G., pp. 87-133. Plenum Press, New York, USA.

Báez, A.M., Marsicano, C. and Cione, A. (1993). Vertebrados mesozóicos. In: *Geologia y recursos naturales de Mendoza*, ed. Ramos, V.A., pp. 341-348. Asociación Geológica Argentina/ Instituto Argentino de Petróleo, Buenos Aires, Argentina.

Bahlburg, H. and Breitkreuz, C. (1991). Paleozoic evolution of active margin basins in: the Southern Central Andes (Northwestern Argentina and Northern Chile). *Journal of South American Earth Sciences*, 4(3), 171-188.

Baldi, J.E. and Nevistic, V.A. (1996). Cuenca costa afuera del Golfo San Jorge. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 171-192. Asociación Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Barberena, M.C., Araújo, D.C., Lavina, E.L., and Azevedo, S.A.K. (1985) O estado atual do conhecimento sobre os tetrápodos permianos e triássicos do Brasil Meridional. *Coletânea de Trabalhos Paleontológicos*, **27**(2), 21-28.

Barcat, C., Cortiñas, J.S., Nevistic, V.A. and Zucchi, H.E. (1989). Cuenca Golfo San Jorge. In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 319-345. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Beltrami, C.V., Alves, L.E.M., and Feijó, F.J. (1994). Bacia do Ceará. *Boletim de Geociências da Petrobras*, **8**(1), 117-125.

Berrangé, J.P. (1977) The geology of southern Guyana, South America. Institute of geological sciences, London, Great Britain, *Overseas memoir*, **14**, 75-78.

Bertani, R.T., Costa, I.G. and Matos, R.M.D. (1990). Evolução tectono-sedimentar, estilo estrutural e hábitat do petróleo da Bacia Potiguar. In: *Origem e evolução de bacias sedimentares*, eds. Raja Gabaglia, G.P and Milani, E.J., pp. 291-310. Petrobras, Rio de Janeiro, Brasil.

Biddle, K.T., Snavely III, P.D., and Uliana, M.A. (1996). Plateau de las Malvinas. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 225-252. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Boesi, T. and Goddard, D. (1991). A new geologic model related to the distribution of hydrocarbon source rocks in the Falcón Basin, northwestern Venezuela. In: *Active margin basins*, ed. Biddle, K.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **52**, 303-319

Brandão, J.A.S.L. and Feijó, F.J. (1994a). Bacia da Foz do Amazonas. *Boletim de Geociências da Petrobras*, **8**(1), 91-99.

Brandão, J.A.S.L. and Feijó, F.J. (1994b) Bacia do Pará-Maranhão. *Boletim de Geociências da Petrobras*, **8**(1), 101-102.

Cainelli, C. and Mohriak, W.U. (1998) *Brazilian Geology part II: Geology of Atlantic Eastern Brazilian basins*. Rio'98 AAPG International Conference and Exhibition, American Association of Petroleum Geologists/Associação Brasileira de Geólogos de Petróleo, Short Course Notes, Rio de Janeiro, Brasil.

Caixeta, J.M., Bueno, G.V., Magnavita, L.P. and Feijó, F.J. (1994). Bacias do Recôncavo, Tucano e Jatobá. *Boletim de Geociências da Petrobras*, **8**(1), 163-172.

Caputo, M.V. (1984). *Stratigraphy, tectonics, paleoclimatology and paleogeography of Northern basins of Brazil*. University of California, PhD Thesis, Santa Barbara, USA (unpublished).

Caputo, M.V and Crowell, J.C. (1985). Migration of glacial centers across Gondwana during Paleozoic Era. *Geological Society of America Bulletin*, **96**, 1020-1036.

Caputo, M.V. and Silva, O.B. (1990). Sedimentação e tectônica da Bacia do Solimões. In: *Origem e evolução de bacias sedimentares*, eds. Raja Gabaglia, G.P and Milani, E.J., pp. 169-193. Petrobras, Rio de Janeiro, Brasil.

Cazau, L., Mancini, D., Cangini, J., and Spalletti, L. (1989). Cuenca de Ñirihuau. In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 299-318. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Chang, H.K., Kowsmann, R.O., Figueiredo, A.M.F., Bender, A.A. (1992). Tectonics and stratigraphy of the East Brazil rift system (EBRIS): an overview. *Tectonophysics*, **213**, 97-138.

Comínguez, A.H. and Ramos, V.A. (1995). Geometry and seismic expression of the Cretaceous Salta rift system, northwestern Argentina. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 325-340.

Conceição, J.C.J., Zalán, P.V. and Wolff, S. (1988). Mecanismo, evolução e cronologia do rifte Sul Atlântico. *Boletim de Geociências da Petrobras*, 2(2/4), 255-265.

Cooper, M.A., Addison, F.T., Alvarez, R., Hayward, A.B., Howe, S., Pulham, A.J. and Taborda, A. (1995). Basin development and tectonic history of the Llanos Basin, Colombia. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 659-666.

Cordani, U.G., Neves, B.B.B., Fuck, R.A., Porto, R., Thomaz F., A. and Cunha, F.M.B. (1984). *Estudo preliminar de integração do Pré-Cambriano com os eventos tectônicos das bacias sedimentares brasileiras*. Petrobras, Série Ciência-Técnica-Petróleo 15, Rio de Janeiro, Brasil.

Cremonini, O.A., Goulart, J.P.M., and Soares, U.M. (1998). Tectonic-stratigraphic evolution of offshore Potiguar Basin, Brazil. *Proceedings of the Rio'98 AAPG International Conference* and *Exhibition*, Rio de Janeiro, Brazil, 44-45.

Cruz, N.M.C. (1987) *Quitinozodrios da Bacia do Solimões, Brasil.* Petrobras/ Companhia de Pesquisas de Recursos Minerais, internal report, Belém, Brasil.

Cunha, P.R.C., Gonzaga, F.G., Coutinho, L.F.C. and Feijó, F.J. (1994). Bacia do Amazonas. *Boletim de Geociências da Petrobras*, 8(1), 47-55.

Daemon, R.F. and Contreiras, C.J.A. (1971). *Zoneamento palinológico da Bacia do Amazonas.* In: *Anais do XXV Congresso Brasileiro de Geologia* 3, São Paulo, Brasil, 79-88.

Damuth, J.E. (1976) Sedimentation on the north Brazilian continental margin. *Anais da Academia Brasileira de Ciências*, 48, 43-50.

Demercian, S., Szatmari, P., and Cobbold, P.R. (1993). Style and pattern of salt diapirs due to thin-skinned. gravitational gliding, Campos and Santos basins, offshore Brazil. *Tectonophysics*, 228, 393-433.

De Wit, M.J., Jeffery, M., Bergh, H., and Nicolaysen, L. (1988). *Geological map of sectors of Gondwana, reconstructed. to their disposition ~ 150 Ma.* American Association of Petroleum Geologists/University of Witwatersrand, wall map, Tulsa, USA.

Dias, J.L. (1998). *Análise sedimentológica e estratigráfica do* Andar *Aptiano em parte da margem leste do Brasil e no platô das Malvinas - considerações sobre as primeiras incursões marinhas do oceano Atlântico Sul meri*dional. Universidade Federal do Rio Grande do Sul, PhD thesis, Porto Alegre, Brasil (unpublished).

Dias, J.L., Sad, A.R.E., Fontana, R.L., and Feijó, F.J. (1994). Bacia de Pelotas. *Boletim de Geociências da Petrobras*, 8(1), 235-245.

Dias, J.L., Scarton, J.C., Esteves, F.R., Carminatti, M., and Guardado, L.R. (1990). Aspectos da evolução tectono-sedimentar e a ocorrência de hidrocarbonetos na Bacia de Campos. In: *Origem e evolução de bacias sedimentares*, eds. Raja Gabaglia, G.P and Milani, E.J., pp. 333-360. Petrobras, Rio de Janeiro, Brasil.

Díaz-Martínez, E. (1995). *Devonico Superior y Carbonifero del Altiplano de Bolivia: estratigrafia, sedimentologia y evolución paleogeográfica.* ORSTOM Bolívia, Informe 46, La Paz, Bolívia.

Du Toit, A.L. (1927). *A geological comparison of South America with South Africa.* The Carnegie Institution, publ. 381, Washington, USA, 157 p.

Eiras, J. F. (1996). Influência tectônica do arco de Carauari na sedimentação fanerozóica da bacia do Solimões, Norte do Brasil. In: *Anais do XXXIX Congresso Brasileiro de Geologia* 1. Sociedade Brasileira de Geologia, Salvador, Brasil, 50-53.

Eiras, J.F., Becker, C.R., Souza, E.M., Gonzaga, F.G., Silva, J.G.F., Daniel, L.M.F., Matsuda, N.S. and Feijó, F.J. (1994a). Bacia do Solimões. *Boletim de Geociências da Petrobras*, 8(1), 17-45.

Eiras, J.F. and Kinoshita, E.M. (1988). Evidências de movimentos transcorrentes na Bacia do Tacutu. *Boletim de Geociências da Petrobras*, 2(2/4), 193-208.

Eiras, J.F. and Kinoshita, E.M. (1990). Geologia e perspectivas petrolíferas da Bacia de Tacutu. In: *Origem e evolução de bacias sedimentares*, eds. Raja Gabaglia, G.P and Milani, E.J., pp. 197-220. Petrobras, Rio de Janeiro, Brasil.

Eiras, J.F., Kinoshita, E.M. and Feijó, F.J. (1994b). Bacia do Tacutu. *Boletim de Geociências da* Petrobras, 8(1), 83-89.

Eva, A., Burke, K., Mann, P. and Wadge, G. (1989). Four-phase tectonostratigraphic development of the southern Caribbean. *Marine and Petroleum Geology*, 6, 9-21.

Feijó, F.J. (1994a), Bacia de Barreirinhas. *Boletim de Geociências da Petrobras*, 8(1), 103-109.

Feijó, F.J. (1994b). Bacia de Pernambuco-Paraíba. *Boletim de Geociências da Petrobras*, 8(1), 143-147.

Feijó, F.J. (1994c). Bacias de Sergipe e Alagoas. *Boletim de Geociências da Petrobras*, 8(1), 149-161.

Feijó, F.J. and Souza, R.G. (1994). Bacia do Acre. *Boletim de Geociências da Petrobras* 8(1), 9-16.

Fernandes, L.A. and Coimbra, A.M. (1996). A Bacia Bauru (Cretáceo Superior, Brasil). *Anais da Academia Brasileira de Ciências*, 68(2), 195-205.

Fernández-Seveso, F. and Tankard, A. (1995), Tectonics and stratigraphy of the Late Paleozoic Paganzo Basin of western Argentina and its regional implications. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, 62, 285-301.

Figueiredo, A.M.F., Braga, J.A.E., Zabalaga, J.C., Oliveira, J.J., Aguiar, G.A., Silva, O.B., Mato, L.F., Daniel, L.M.F., Magnavita, L.P. and Bruhn, C.H. (1994). Recôncavo Basin, Brazil: a prolific intracontinental rift basin. In: Interior rift basins, ed. Landon, S.M. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, 59, 157-204.

Figueiredo, A.M.F., Miranda, A.P., Ferreira, R.F. and Zalán, P.V. (1996). Cuenca de San Julian. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 193-212. Asociacion Geologica Argentina/ Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Fontana, R.L. (1989). Evidências geofísicas da presença de hidratos de gás na Bacia de Pelotas. In: *Anais do Primeiro Congresso Brasileiro de Geofísica* 1. Sociedade Brasileira de Geofísica, Rio de Janeiro, Brasil, 234-248.

Fontana, R.L. (1990). Desenvolvimento termomecânico da Bacia de Pelotas e parte sul da Plataforma de Florianópolis. In: *Origem e evolução de bacias sedimentares*, eds. Raja Gabaglia, G.P and Milani, E.J., pp. 377-400. Petrobras, Rio de Janeiro, Brasil.

França, A.B. and Potter, P.E. (1988). Estratigrafia, ambiente deposicional e análise de reservatório do Grupo Itararé (Permocarbonífero), Bacia do Paraná (Parte 1). *Boletim de Geociências da Petrobras*, 2(2-4), 147-191.

França, A.B., Milani, E.J., Schneider, R.L., López-Paulsen, O., López-Pugliessi, J.M., Suárez S., R., Santa Ana, H., Wiens, F., Ferreiro, O., Rossello, E.A., Bianucci, H.A., Flores, R.F.A., Vistalli, M.C., Fernández-Seveso, F., Fuenzalida, R.P., and Muñoz, N. (1995). Phanerozoic correlation In: Southern South America. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J.. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, 62, 129-161.

Fryklund, B., Marshall, A. and Stevens, J. (1996). Cuenca del Colorado. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 135-158. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Galeazzi, J.S. (1996). Cuenca de Malvinas. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 273-309. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Góes, A.M.O., Souza, J.M.P. and Teixeira, L.B. (1990). Estágio exploratório e perspectivas petrolíferas da Bacia do Parnaíba. *Boletim de Geociências da Petrobras*, 4(1), 55-64.

Góes, A.M.O. and Feijó, F.J. (1994). Bacia do Parnaíba. *Boletim de Geociências da Petrobras*, 8(1), 57-67.

Gohrbandt, K.H.A. (1993). Paleozoic paleogeographic and depositional developments on the central proto-Pacific margin of Gondwana: their importance to hydrocarbon accumulation. *Journal of South American Earth Sciences*, 6(4), 267-287.

Gómez Omil, R.J., Boll, A., and Hernández, R.M. (1989), Cuenca cretácico-terciária del Noroeste Argentino (Grupo Salta). In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 43-

64. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Gonzaga, F.G., Gonçalves, F.T.T., and Coutinho, L.F.C. (1997), Modelling of generation and migration in the Amazonas Basin, Brazil. In: *AAPG Hedberg Research Symposium on Petroleum Systems of the South Atlantic Margin, extended. abstracts volume*, Rio de Janeiro, Brasil, 68-71.

Goossens, P.J. and Rose, W.I. (1973), Chemical composition and age determination of tholeiitic rocks in the Basic Cretaceous Complex, Ecuador. *Geological Society of America Bulletin*, **84**, 1043-1052.

Gorini, M.A. (1977) *The tectonic fabric of the Equatorial Atlantic and adjoining continental margins: Gulf of Guinea to northeastern Brazil*. Columbia University, PhD thesis, Columbia, USA (unpublished).

Gradstein, F. and Ogg, J. (1996). A Phanerozoic time scale. Supplement to *Episodes* **19**(1-2).

Gulisano, C.A., Gutierrez-Pleimling, A.R. and Digregorio, R.E. (1984). Análisis estratigráfico del intervalo Tithoniano-Valanginiano (formaciones Vaca Muerta, Quintuco y Mulichinco) en el suroeste de la Província de Neuquén. In: *Actas de lo Noveno Congreso Geológico Argentino* 1,. Asociación Geológica Argentina, San Carlos de Bariloche, Argentina, 221-235.

Gust, D.A., Biddle, K.T., Phelps, D.W., and Uliana, M.A. (1985). Associated. middle to late Jurassic volcanism and extension in Southern South America. *Tectonophysics*, **116**, 223-253.

Jager, J. (1997). *Shell en la Cuenca Chaco-Paranaense*. Shell internal report, distributed. by the Instituto Argentino del Petróleo y del Gás, Mesa Redonda - Cuenca Chaco-Paranaense, Buenos Aires, Argentina.

Jaillard, E., Ordoñez, M., Benitez, S., Berrones, G., Jiménez, N., Montenegro, G. and Zambrano, I. (1995). Basin development in an accretionary, oceanic-floored. fore-arc setting: southern coastal Ecuador during Late Cretaceous-Late Eocene time. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 615-631.

Johnson, J.G. (1971). Timing and coordination of orogenic, epeirogenic and eustatic events. *Geological Society of America Bulletin*, **82**, 3263-3298.

Juan, R.C., Jager, J., Russell, J., and Gebhard, I. (1996). Flanco norte de la Cuenca del Colorado. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 117-133. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Keeley, M.L. and Light, M.P.R. (1993). Basin evolution and prospectivity of the Argentine continental margin. *Journal of Petroleum Geology*, **16**(4), 451-464.

Keidel, J. (1916), La geología de las sierras de la Província de Buenos Aires y sus relaciones con las montañas de Sud Africa y los Andes. *Anales del Ministerio de Agricultura de la Nación, Sección Geología, Mineralogía y Minería*, **9**(3), 1-78, Buenos Aires, Argentina.

Kellogg, J.N. (1984), Cenozoic tectonic history of the Sierra de Perijá, Venezuela-Colombia and adjacent basins. In: The Caribbean-South American plate boundary and regional tectonics, eds. Bonini, W.E., Hargraves, R.B. and Shagan, R. *Geological Society of America, Memoir*, Boulder, USA, **162**, 239-261.

Kokogian, D.A., Fernández-Seveso, F., and Mosquera, A. (1993), Las secuencias sedimentarias triasicas. In: *Geologia y recursos naturales de Mendoza*, ed. Ramos, V.A., pp. 65-78. Asociación Geológica Argentina/Instituto Argentino de Petróleo, Buenos Aires, Argentina.

Kokogian, D.A. and Mancilla, O. (1989), Analisis estratigrafico secuencial de la Cuenca Cuyana. In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 169-201. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Kress, P.R., Franzin, H.J., and Marinelli, R.V. (1996), Cuenca de Malvinas Oriental. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 311-321. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Krook, L. (1975), Surinam. In: *The encyclopedia of world regional geology, part 1: Western Hemisphere (including Antarctica* and *Australia)*, ed. Fairbridge, R.W., pp. 480-492. Dowden, Hutchinson and Ross, Stroudsburg, USA.

Lana, M. C. and Milani, E. J. (1986), A microplaca do Nordeste brasileiro - um elemento dinâmico no rifteamento cretácico inferior. In: *Anais do XXXIV Congresso Brasileiro de Geologia* 3. Sociedade Brasileira de Geologia, Goiânia, Brasil, 1131-1144.

Legarreta, L. and Gulisano, C.A. (1989), Analisis estratigrafico secuencial de la Cuenca Neuquina (Triasico superior-Terciario inferior). In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 221-243. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Legarreta, L., Gulisano, C.A. and Uliana, M.A. (1993), Las secuencias sedimentarias jurasico-cretacicas. In: *Geologia y recursos naturales de Mendoza*, ed. Ramos, V.A., pp. 87-114. Asociación Geológica Argentina/Instituto Argentino de Petróleo, Buenos Aires, Argentina.

Lelarge, M.L.M.V. (1993), *Thermochronologie par la methode des traces de fission d'une marge passive (dome de Ponta Grossa, SE Brésil) et au sein d'une chaine de collision (zone externe de lárc Alpin, France)*. Universite Joseph Fourier, PhD thesis, Grenoble, France (unpublished).

Lesta, P., Ferello, R. and Chebli, G. (1980). Chubut Extraandino. In: *Actas de lo Segundo Simposio de Geologia Regional Argentina*, ed. Turner, J.C.M. Academia Nacional de Ciencias, Córdoba, Argentina, *1307-1387*.

Lesta, P., Mainardi, E. and Stubelj, R. (1980). Plataforma continental argentina. In: *Actas de lo Segundo Simposio de Geologia Regional Argentina*, ed. Turner, J.C.M., pp. 1577-1601. Academia Nacional de Ciencias, Córdoba, Argentina.

Light, M.P.R., Keeley, M.L., Maslanyj, M.P., and Urien, C.M. (1993). The tectono-stratigraphic development of Patagonia, and its relevance to hydrocarbon exploration. *Journal of Petroleum Geology*, **16**(4), 465-482.

Lima, H.P., Aranha, L.G.F. and Feijó, F.J. (1994). Bacias de Bragança-Viseu, São Luís e gráben de Ilha Nova. *Boletim de Geociências da Petrobras*, **8**(1), 111-116.

Linares, E. (1977). *Catálogo de edades radimétricas determinadas para la República Argentina*. Asociación Geológica Argentina, publicacion especial B-4, Buenos Aires, Argentina.

Linsser, H. (1974). Interpretação das anomalias gravimétricas regionais na área amazônica. *Boletim Técnico da Petrobras*, **17**(1), 3-15.

López-Gamundí, O.R., Conaghan, P.J., Rossello, E.A. and Cobbold, P.R. (1995). The Tunas Formation (Permian) in the Sierras Australes Foldbelt, east central Argentina: evidence for syntectonic sedimentation in a foreland basin. *Journal of South American Earth Sciences* **8**(2), 129-142.

López-Gamundí, O.R., Espejo, I.S., Conaghan, P.J. and Powell, C.M.A. (1994). Southern South America. In: Permian-Triassic Pangean basins and foldbelts along the Panthalassan margin of Gondwanaland, eds. Veevers, J.J. and Powell, C.M.A. *Geological Society of America, Memoir*, Boulder, USA, **184**, 281-329.

Lugo, J. and Mann, P. (1995). Jurassic-Eocene tectonic evolution of Maracaibo Basin, Venezuela. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J.

American Association of Petroleum Geologists, Memoir, Tulsa, USA, **62**, 699-725.

Macellari, C.E. (1988). Cretaceous paleogeography and depositional cycles of western South America. *Journal of South American Earth Sciences*, **1**, 373-418.

Macellari, C.E. (1995). Cenozoic sedimentation and tectonics of the southwestern Caribbean pull-apart basin, Venezuela and Colombia. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 757-780.

Magnavita, L.P. (1992). *Geometry and kinematics of the Recôncavo-Tucano-Jatobá rift, NE Brazil*: University of Oxford, PhD thesis, Oxford, UK (unpublished).

Manspeizer, W. (1988). Triassic-Jurassic rifting and opening of the Atlantic: an overview. In: Triassic-Jurassic rifting, ed. Manspeizer, W., pp. 41-79. Elsevier, Developments. In: *Geotectonics* 22, Amsterdam, the Netherlands.

Marinelli, R.V. and Franzin, H.J. (1996). Cuencas de Rawson y Península de Valdés. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 159-169. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Marocco, R., Lavenu, A., and Baudino, R. (1995). Intermontane Late Paleogene-Neogene basins of the Andes of Ecuador and Peru: sedimentologic and tectonic characteristics. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 597-613.

Mascle, J., Blarez, E., and Marinho, M. (1988). The shallow structures of the Guinea and Ivory Coast-Ghana transform margins: their bearing on the Equatorial Atlantic Mesozoic evolution. *Tectonophysics*, **155**, 193-209.

Mascle, J. and Renard, V. (1976). The marginal São Paulo plateau, comparison with the southern Angolan margin. *Anais da Academia Brasileira de Ciências*, **48**, 180-190.

Massabie, A.C. and Rossello, E.A. (1984). La discordancia pre-Formación Sauce Grande y su entorno estratigráfico, Sierras Australes de la Província de Buenos Aires. *Actas de lo IX Congreso Geológico Argentino* 1, Buenos Aires, Argentina, 337-352.

Mathalone, J.M.P. (1996) *South American sedimentary basins (map, scale 1:10,000,000)*. PetroSantander Inc., Houston, USA.

Mathalone, J.M.P. and Montoya, M. (1995). Petroleum geology of the sub-Andean basins of Peru. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir* ,Tulsa, USA, **62**, 423-444.

Matos, R.M.D. (1998). The transversal zone: a key feature between NE Brazil and W Africa. *Proceedings of the Rio'98 AAPG International Conference* and *Exhibition*, Rio de Janeiro, Brazil, 426-427.

Matos, R.M.D. and Waick, R.N. (1998). A unique transform margin: the Equatorial Atlantic. *Proceedings of the Rio'98 AAPG International Conference* and *Exhibition*, Rio de Janeiro, Brazil, 798-799.

Melo, J.H.G. (1988). The Malvinokaffric realm in the Devonian of Brazil. In: *Devonian of the world*, eds. McMillan, N.J., Embry, A.F. and Glass, D.J., pp. 669-704. Canadian Society of Petroleum Geologists, Memoir 14, I, Calgary, Canada.

Milani, E. J. (1985). Tectônica cisalhante na evolução do rift do Recôncavo-Tucano-Jatobá. *Revista Brasileira de Geociências*, **15**(4), 287-292.

Milani, E.J. (1987). *Aspectos da evolução tectônica das Bacias do Recôncavo e Tucano Sul, Bahia, Brasil*. Petrobras, Série Ciência-Técnica-Petróleo 18, Rio de Janeiro, Brasil.

Milani, E.J. (1991). Anomalias gravimétricas em bacias do tipo rifte: exemplos brasileiros. *Anais do II Congresso Internacional da Sociedade Brasileira de Geofísica* 1, Salvador, Brasil, 172-176.

Milani, E.J. (1997). *Evolução tectono-estratigráfica da Bacia do Paraná e seu relacionamento com a geodinâmica fanerozóica do Gondwana sul-ocidental*. Universidade Federal do Rio Grande do Sul, PhD thesis, Porto Alegre, Brasil (unpublished).

Milani, E.J. (1999). Terrenos, orogenias e a evolução tectono-estratigráfica da Bacia do Paraná. *Anais do VII Simpósio Nacional de Estudos Tectônicos*, pp. 3-6, Lençóis, Brasil.

Milani, E.J., França, A.B., and Schneider, R.L. (1994). Bacia do Paraná. *Boletim de Geociências da Petrobras*, **8**(1), 69-82.

Milani, E.J., Lana, M.C., and Szatmari, P. (1988). Mesozoic rift basins around the Northeast Brazilian microplate (Recôncavo-Tucano-Jatobá, Sergipe-Alagoas). In: *Triassic-Jurassic rifting*, ed. Manspeizer, W., pp. 833-858. Elsevier, Developments In: Geotectonics 22, Amsterdam, the Netherlands.

Milani, E.J. and Latgé, M.A.L. (1987). Modelagem gravimétrica da porção terrestre da Bacia Potiguar e suas implicações geotectônicas. *Boletim de Geociências da Petrobras* **1**(1), 75-85.

Milani, E.J. and Ramos, V.A. (1998). Orogenias paleozóicas no domínio sul-ocidental do Gondwana e os ciclos de subsidência da Bacia do Paraná. *Revista Brasileira de Geociências*, **28**(4), 527-544.

Milani, E.J. and Szatmari, P. (1998). Rift dynamics in Northeastern Brazil. *Proceedings of the Rio'98 AAPG International Conference* and *Exhibition*, pp. 368-369, Rio de Janeiro, Brazil.

Milani, E.J. and Zalán, P.V. (1998). *Brazilian Geology Part I: the Geology of Paleozoic Cratonic Basins* and *Mesozoic Interior Rifts of Brazil*. Rio'98 AAPG International Conference and Exhibition, American Association of Petroleum Geologists/ Associação Brasileira de Geólogos de Petróleo, Short Course Notes, Rio de Janeiro, Brasil.

Milani, E.J. and Zalán, P.V. (1999). An outline of the geology and petroleum systems of the Paleozoic interior basins of South America. *Episodes*, **22**(3), 199-205.

Miura, K. and Barbosa, J.C. (1972). Geologia da plataforma continental do Maranhão, Piauí, Ceará e Rio Grande do Norte. *Anais do XXVI Congresso Brasileiro de Geologia* 2, Belém, Brasil, 57-66.

Mohriak, W.U., Macedo, J.M., Castellani, R.T., Rangel, H.D., Barros, A.Z.N., Latgé, M.A.L., Rizzi, J.A., Mizusaki, A.M.P., Szatmari, P., Rizzo, J.G., and Aires, J.R. (1995). Salt tectonics and structural styles in the deep-water province of the Cabo Frio region, Rio de Janeiro, Brazil. In: Salt tectonics: a global perspective, eds. Jackson, M.P.A., Roberts, D.G., and Snelson, S. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **65**, 273-304.

Moratello, J.H.(1993). Cuenca Cuyana. In: *Geologia y recursos naturales de Mendoza*, ed. Ramos, V.A., pp. 367-375. Asociación Geológica Argentina/Instituto Argentino de Petróleo, Buenos Aires, Argentina.

Mossmann, R., Falkenhein, F.U.H., Gonçalves, A., and Nepomuceno Filho, F. (1986). Oil and gas potential of the Amazon Paleozoic Basins. In: Future petroleum provinces of the world, ed. Halbouty, M.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **40**, 207-241.

Netto, A.S.T., Filho, J.R.W., and Feijó, F.J. (1994). Bacias de Jacuípe, Camamu e Almada. *Boletim de Geociências da Petrobras*, **8**(1), 173-184.

Nullo, F.E., Dimieri, L., and Kusiak, M.E. (1996). Geologia de las Islas Malvinas. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 213-224. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Ojeda, H.A. (1982). Structural framework, stratigraphy and evolution of Brazilian marginal basins. *American Association of Petroleum Geologists Bulletin*, **66**, 732-749.

Oliveira, C.M.M. (1994) *Estilos estruturais e evolução tectônica da bacia do Acre*. Universidade Federal de Ouro Preto, MSc thesis, Ouro Preto, Brasil (unpublished).

Pareja, J.L., Vargas, C.F., Suárez, R.S., Ballón, R.A., Carrasco, R.C., and Villaroel, C.A. (1978). *Mapa geológico de Bolívia - memoria explicativa*. Yacimientos Petroliferos Fiscales Bolivianos/ Servicio Geológico de Bolivia, La Paz, Bolívia.

Parnaud, F., Gou, Y., Pascual, J.-C., Truskowski, I., Gallango, O., Passalacqua, H. and Roure, F. (1995a). Petroleum geology of the central part of the Eastern Venezuelan Basin. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 741-756.

Parnaud, F., Gou, Y., Pascual, J.-C., Capello, M.A., Truskowski, I. and Passalacqua, H. (1995b). Stratigraphic synthesis of western Venezuela. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 681-698.

Pereira, M.J. and Feijó, F.J. (1994). Bacia de Santos. *Boletim de Geociências da Petrobras*, **8**(1), 219-234.

Perez de Mejía, D., Kiser, G.D., Maximowitsch, B., and Young, G.A. (1980). Geologia de Venezuela. In: *Evaluacion de formaciones en Venezuela*, ed. Felder, B., pp. 1-23. Schlumberger, Caracas, Venezuela.

Peters, K.E., Wagner, J.B., Carpenter, D.G. and Conrad, K.T. (1997). Potencial devoniano de primera clase se evidencia al oriente de la cuenca Madre de Dios. *Oil* and *Gas Journal Revista Latinoamericana*, **3**, 43-53.

Pezzi, E.E. and Mozetic, M.E. (1989). Cuencas sedimentarias de la region Chacoparanense. In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 65-78. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Ramos, V.A. (1988). Late Proterozoic - Early Paleozoic of South America: a collisional history. *Episodes*, **11**(3), 168-174.

Ramos, V.A. (1990). *Field guide to geology of the Central* Andes *(31° - 33° SL)*. Universidad de Buenos Aires, Central Andes Field Seminar notes, Buenos Aires, Argentina.

Ramos, V.A. (1992). Control geotectónico de las cuencas triásicas de Cuyo. *Boletin de Informaciones Petroleras*, **9**(31), 2-9.

Ramos, V.A. (1993). Interpretación tectónica. In: *Geologia y recursos naturales de Mendoza*, ed. Ramos, V.A., pp. 257-268. Asociación Geológica Argentina/Instituto Argentino de Petróleo, Buenos Aires, Argentina.

Ramos, V.A. (1996). Evolución tectonica de la plataforma continental. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 385-404. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Ramos, V.A., Jordan, T.E., Allmendinger, R.W., Mpodozis, C., Kay, S.M., Cortés, J.M. and Palma, M. (1986). Paleozoic terranes of the central Argentine-Chilean Andes. *Tectonics*, **5**(6), 855-880.

Rangel, H.D., Martins, F.A.L., Esteves, F.R., and Feijó, F.J. (1994). Bacia de Campos. *Boletim de Geociências da Petrobras*, **8**(1), 203-217.

Rapela, C.W. and Pankhurst, R. (1994). Variaciones temporales y composicionales del magmatismo jurásico de la Patagonia. *Actas de lo VII Congreso Geológico Chileno* II, Concepción, Chile, 1433-1437.

Rast, N. (1988). Variscan-Alleghenian orogen. In: *Triassic-Jurassic rifting*, ed. Manspeizer, W., pp. 1-27. Elsevier, Developments in Geotectonics 22, Amsterdam, the Netherlands.

Robbiano, J.A., Arbe, H.A., and Gangui, A. (1996). Cuenca Austral marina. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 323-342. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Roeder, D. (1988). Andean-age structure of Eastern Cordillera (Province of La Paz, Bolivia). *Tectonics* **7**, 23-39.

Ross, J.G., Pinchin, J., Griffin, D.G., Dinkelman, M.G., Turic, M.A. and Nevistic, V.A. (1996). Cuenca de Malvinas Norte. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 253-272. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Russo, A., Archangelsky, S., Andreis, R.A. and Cuerde, A. (1987). Cuenca Chacoparanense. In: *El sistema carbonifero en la Republica Argentina*, ed. Archangelsky, S., pp. 197-212. Academia Nacional de Ciencias, Cordoba, Argentina.

Santos, C.F., Cupertino, J.A. and Braga, J.A.E. (1990). Síntese sobre a geologia das bacias do Recôncavo, Tucano and Jatobá. In: *Origem e evolução de bacias sedimentares*, eds. Raja Gabaglia, G.P and Milani, E.J., pp. 235-266. Petrobras, Rio de Janeiro, Brasil.

Santos, C.F., Gontijo, R.C., Araújo, M.B., and Feijó, F.J. (1994). Bacias de Cumuruxatiba e Jequitinhonha. *Boletim de Geociências da Petrobras*, **8**(1), 185-190.

Schamel, S. (1991). Middle and Upper Magdalena Basins, Colombia. In: Active margin basins, ed. Biddle, K.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **52**, 283-302.

Sempere, T. (1995). Phanerozoic evolution of Bolivia and adjacent regions. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 207-230.

Starck, D. (1995). Silurian-Jurassic stratigraphy and basin evolution of northwestern Argentina. In: *Petroleum basins of South America*, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 251-267

Stoakes, F.A., Campbell, C.V., Cass, R. and Ucha, N. (1991). Seismic stratigraphic analysis of the Punta del Este Basin, offshore Uruguay. *American Association of Petroleum Geologists Bulletin*, 75(2), 219-240.

Szatmari, P., Françolin, J.B.L., Zanotto, O.A., and Wolff, S. (1987). Evolução tectônica da margem equatorial brasileira. *Revista Brasileira de Geociências*, 17, 180-188.

Szatmari, P. and Milani, E.J. (1999). Microplate rotation in northeast Brazil during South Atlantic rifting: Analogies with the Sinai microplate. *Geology*, **27**(12): 1115-1118.

Szatmari, P., Milani, E.J., Lana, M.C., Conceição, J.C.J. and Lobo, A.P. (1984) How South Atlantic rifting affects Brazilian oil reserves distribution. *Oil and Gas Journal* **83**, 107-113.

Tankard, A.J., Uliana, M.A., Welsink, H.J., Ramos, V.A., Turic, M., França, A.B., Milani, E.J., Brito Neves, B.B., Eyles, N., Skarmeta, J., Santa Ana, H., Wiens, F., Cirbián, M., López-Paulsen, O., Germs, G.J.B., De Wit, M.J., Machacha, T. and Miller, R.M.G. (1995). Structural and tectonic controls of basin evolution in Southwestern Gondwana during the Phanerozoic. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 5-52.

Tavella, G.F. and Wright, C.G. (1996). Cuenca del Salado. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 95-116. Asociacion Geologica Argentina/Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Thomaz Filho, A., Cordani, U.G., and Kawashita, K. (1976). Aplicação do método Rb/Sr na datação de rochas sedimentares argilosas da Bacia do Paraná. *Anais do XXIX Congresso Brasileiro* *de Geologia* 2, Ouro Preto, Brasil, 290-302.

Turner, S., Regelous, M., Kelley, S., Hawkesworth, C., and Mantovani, M.S.M. (1994). Magmatism and continental break-up in the South Atlantic: high precision Ar/Ar geochronology. *Earth and Planetary Science Letters*, 121, 333-348.

Uchupi, E. and Emery, K.O. (1991). Pangaean divergent margins: historical perspective. *Marine Geology*, **102**, 1-2.

Uliana, M.A. and Biddle, K.T. (1988). Mesozoic-Cenozoic paleogeographic and geodynamic evolution of southern South America. *Revista Brasileira de Geociências*, **18**(2), 172-190.

Uliana, M.A. and Legarreta, L. (1993). Hydrocarbon habitat in a Triassic-to-Cretaceous sub-Andean setting: Neuquén Basin, Argentina. *Journal of Petroleum Geology*, **16**(4), 397-420.

Urien, C.M. and Zambrano, J.J. (1996). Estructura del margen continental. In: *Geologia y recursos naturales de la plataforma continental argentina*, eds. Ramos, V.A. and Turic, M.A., pp. 29-66. Asociacion Geologica Argentina/ Instituto Argentino del Petroleo, Buenos Aires, Argentina.

Urien, C.M., Zambrano, J.J., and Yrigoyen, M.R. (1995). Petroleum *basins of southern South America: an overview. In: Petroleum* basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 63-77.

Vail, P.R., Mitchum, R.M. and Thompson, S. (1977). Seismic stratigraphy and global changes of sea level, part 3: relative changes of sea level from coastal onlap. In: Seismic stratigraphy - applications to hydrocarbon exploration, ed. Payton, C.E. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **26**, 63-81.

Veevers, J.J., Powell, C.M.A., Collinson, J.W. and López-Gamundí, O.R. (1994). Synthesis. In: Permian-Triassic Pangean basins and foldbelts along the Panthalassan margin of Gondwanaland, eds. Veevers, J.J. and Powell, C.M.A. *Geological Society of America, Memoir*, Boulder, USA, **184**, 331-353.

Vergani, G.D., Tankard, A.J., Belotti, H.J. and Welsink, H.J. (1995). Tectonic evolution and paleogeography of the Neuquén Basin, Argentina. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, 62, 383-402.

Vieira, R.A.B., Mendes, M.P., Vieira, P.E., Costa, L.A.R., Tagliari, C.V., Bacelar, A.P., and Feijó, F.J. (1994). Bacias de Espírito Santo e Mucuri. *Boletim de Geociências da Petrobras*, **8**(1), 191-202.

Von Huene, R. and Lallemand, S. (1990). Tectonic erosion along the Japan and Peru convergent margins. *Geological Society of America Bulletin*, **102**, 704-720.

Von Huene, R. and Scholl, D.W. (1991) Observations at convergent margins concerning sediment subduction, subduction erosion, and the growth of continental crust. *Reviews of Geophysics*, **29**, 279-316.

Welsink, H.J., Franco, A.M., and Oviedo, A.G. (1995a). Andean and pre-andean deformation, Boomerang hills area, Bolivia. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 481-499.

Welsink, H.J., Martinez, E., Aranibar, O., and Jarandilla, J. (1995b). Structural inversion of a Cretaceous rift basin, southern Altiplano, Bolivia. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 305-324.

Wiens, F. (1995). *Phanerozoic tectonics and sedimentation in the* Chaco Basin of Paraguay, with comments on hydrocarbon potential. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 185-205.

Yrigoyen, M.R. (1989). Cuencas de Rawson y Península Valdés. In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 467-476. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Yrigoyen, M.R. (1989). Cuenca de Malvinas. In: *Cuencas sedimentarias argentinas*, eds. Chebli, G. and Spalletti, L., pp. 481-491. Universidad Nacional de Tucuman, Serie Correlacion Geologica 6, Tucuman, Argentina.

Zalán, P.V. (1985). *Tectonics* and *sedimentation of the Piauí-Camocin: sub-basin, Ceará Basin, offshore northeastern Brazil.* Petrobras, Série Ciência-Técnica-Petróleo 17, Rio de Janeiro, Brasil.

Zalán, P.V., Wolff, S., Astolfi, M.A., Vieira, I.S., Conceição, J.C., Appi, V.T., Neto, E.V.S., Cerqueira, J.R., and Marques, A. (1990). The Paraná Basin, Brazil. In: Interior cratonic basins, eds. Leighton, M.W., Kolata, D.R., Oltz, D.F. and Eidel, J.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **51**, 681-708.

The Andean Belt

"The Andes is the type section of a non-collisional orogen that formed a mountain chain by subduction of oceanic crust under a continental plate... The Andean history will be divided in four major stages. The first is related to the reconstruction of the proto-margin of Gondwana, and consists of amalgamation and collision of different terranes against the Late Proterozoic margin of Gondwana. The second stage in the Late Paleozoic is linked to the formation of the Gondwanides, the first mountain chains developed along the Pacific margin by an Andean-type subduction, and the Alleghanides, related to the closing of the Iapetus ocean and the formation of the Pangea Supercontinent. The third stage is related to a generalized extension during Pangea break-up, that predates the opening of the South Atlantic and related oceans, an it is punctuated by collision of island arcs in the Northern Andes. The last stage is responsible of the present orogen, and includes a great variety of tectonic processes from collision of island arcs, seismic and aseismic ridges, as well as normal subduction of oceanic crust under the South American Plate, that defines the proper Andean-type."

(Ramos and Aleman, this volume)

NORTHERN ANDES

Antenor Aleman and Victor A. Ramos

From the time of the pioneer work of Gansser (1973) it was clear that there existed a sharp geological boundary at the latitude of the Gulf of Guayaquil (figs.1 and 2). North of this boundary, accreted oceanic terranes and pervasive strike-slip deformation played an important role in the evolution of the Andes. In contrast, the Central Andes was built on an ensialic crust with dominant orthogonal shortening. Furthermore, 200 km S of the crustal boundary of the Gulf of Guayaquil, the Huancabamba deflection marked an abrupt change in structural trend from NW to NNE (Gerth, 1955; Ham and Herrera, 1963). These two lines of evidence were used initially to establish a clear distinction between the Northern Andes and the Central Andes (Gansser, 1973; Aubouin, 1973).

Recently, this boundary has been corroborated by geophysical and geochemical studies. Paleomagnetic studies confirmed the allochthonous nature of the Piñon/Dagua Terrane (Roperch *et al.*, 1987). Bouguer gravity maps provided independent evidence to establish the oceanic and ensialic boundary (Feininger and Seguin, 1983; Mooney, 1979). In addition to the above, wide-angle seismic reflection profiles in the Western Cordillera of Colombia showed upper crustal velocities similar to those found in oceanic crust (Mooney, 1979). These geophysical data have also been used to conduct two-dimensional seismic and gravity modelling and infer the crustal structure of the Andes (Feininger and Seguin, 1983; Mooney, 1979). Preliminary geochemical studies established the Mesozoic age and suggested ocean-ridge tholeiites with higher-than-expected K, Sr, and Rb content (Barrero, 1979; Goossens and Rose, 1973; Goossens *et al.*, 1977). This was followed by studies that refined and provided alternative explanations for the origin of these mafic rocks (Lebrat *et al.*, 1985; Bourgois *et al.*, 1987; Marriner and Millward, 1984). Recently, several papers support a multiple origin for these allochthonous terranes that vary from intra-plate oceanic settings, to back-arc, island-arc, and normal MORB tholeiites (Spadea and Espinosa, 1996; Cosma *et al.*, 1998). However, it is important to note that it was the Mesozoic high pressure/low temperature metamorphic event associated with emplacement of these rocks that is unique to the Northern Andes.

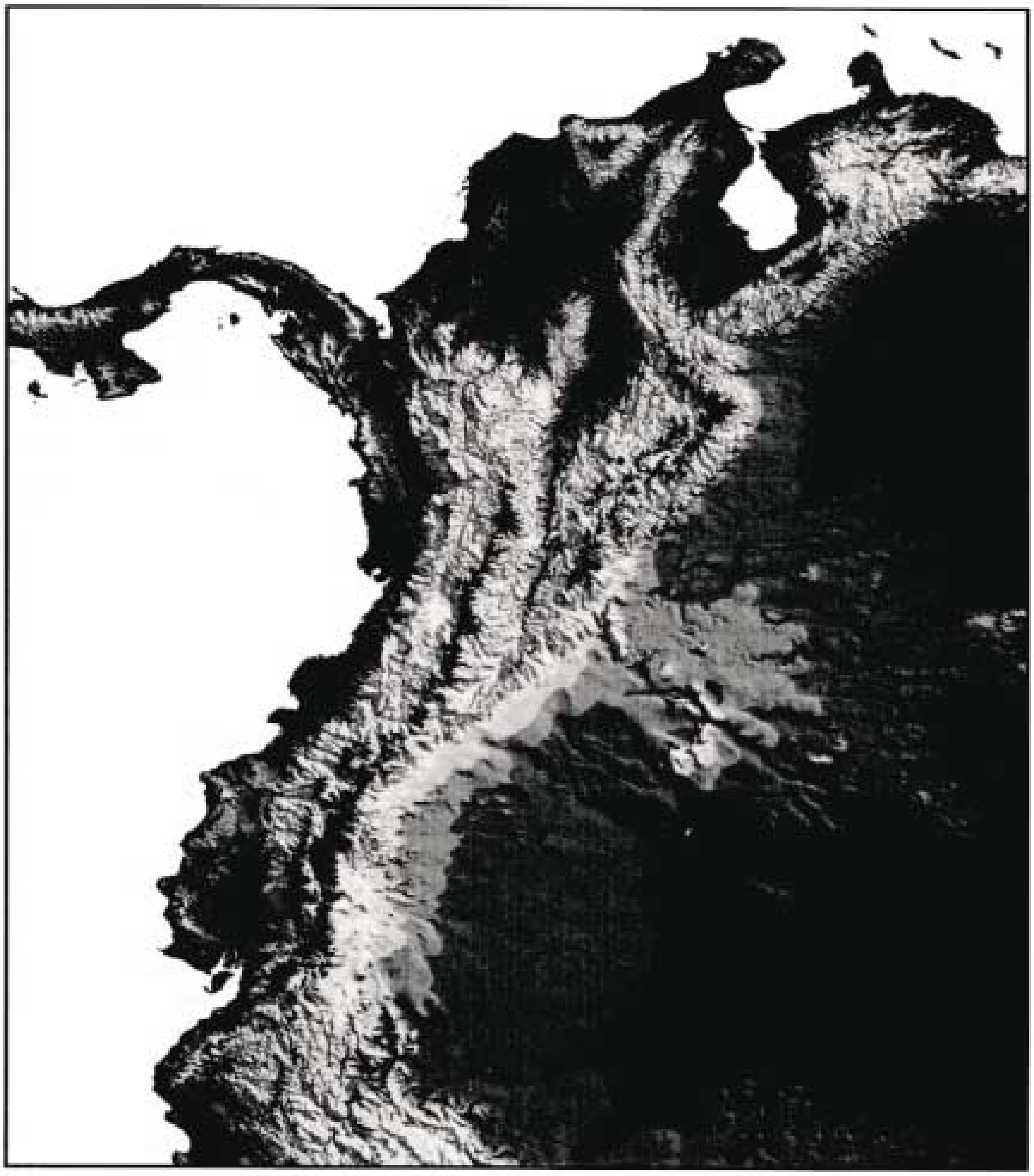
The present Northern Andes may be subdivided into three main segments based on their geological and structural characteristics: the Venezuelan Andes (or Mérida Andes), the Colombian Andes, and the Ecuadorian Andes (Fig.2). Their present political divisions roughly coincide with major changes in the nature and virgation of the different mountain chains. For instance, the Colombian and Ecuadorian Andes correspond to a typical subduction-related mountain chain developed along the continental margin. On the other hand, the intracontinental Mérida Andes were formed as a result of the interaction between the Paleogene Caribbean thrusting and Neogene tectonic inversion during Andean compression. These structures were greatly affected by a complex system of strike-slip faults and folds.

Along-strike changes in the Colombian and Ecuadorian Andes resulted from variations in the Mesozoic extension as well as the fact that Western Colombia was situated along the path of the Caribbean Plate. Pronounced Mesozoic extension in Colombia accounted for the separation of the Central and Eastern Cordilleras. Late Cretaceous collision of the Piñon/Dagua Terrane caused tectonic inversion and uplift of the Central Cordillera with mild deformation in the Eastern Cordillera of Colombia. Although the Ecuadorian Andes were also affected by extension, Late Cretaceous docking of the Piñon/Dagua Terrane caused shortening and flaking of Mesozoic oceanic crust and subsequent amalgamation of the Eastern Cordillera into a single geomorphical unit. Pacific extrusion of the Caribbean Plate was associated with middle Eocene and late Miocene accretion in the Western Cordillera. Complex plate interaction prior to the formation of the Caribbean Plate left a series of small drifting terranes that were later accreted in the northern part of Western Colombia. This periodic terrane docking, enhanced by oblique convergence, caused strain partitioning and continuously reactivated old suture zones.

Because the political frontiers approximately match the geological boundaries, the Northern Andes will be described from the Venezuelan (Mérida) Andes in the N to the Colombian and Ecuadorian Andes to the S in the following sections.

Venezuelan (Mérida) Andes

Gansser (1973), in his review paper, refers to the Mérida Andes as a distinct mountain chain, separated from the main Eastern Cordillera. This distinction is based on important strike changes, and the presence of Precambrian crystalline rocks. Bucher (1952) proposed for its origin a mega-anticline, limited on both flanks by high-angle reverse faults, while Rod (1956) proposed a mushroom-like transpressional uplift with shear deformation along the Boconó Fault System. Imbricate thrusting occurs toward both Andean fronts (Deratmiroff, 1971). Kellogg and Bonini (1982) postulated that crustal shortening along a thrust fault dipping to the SE caused the uplift of the Mérida Andes. Recently, Colletta *et al.* (1997) interpreted the Mérida Andes as a transpressional intercratonic orogenic belt, developed in response to oblique convergence between two continental lithosphere blocks (Fig. 3).

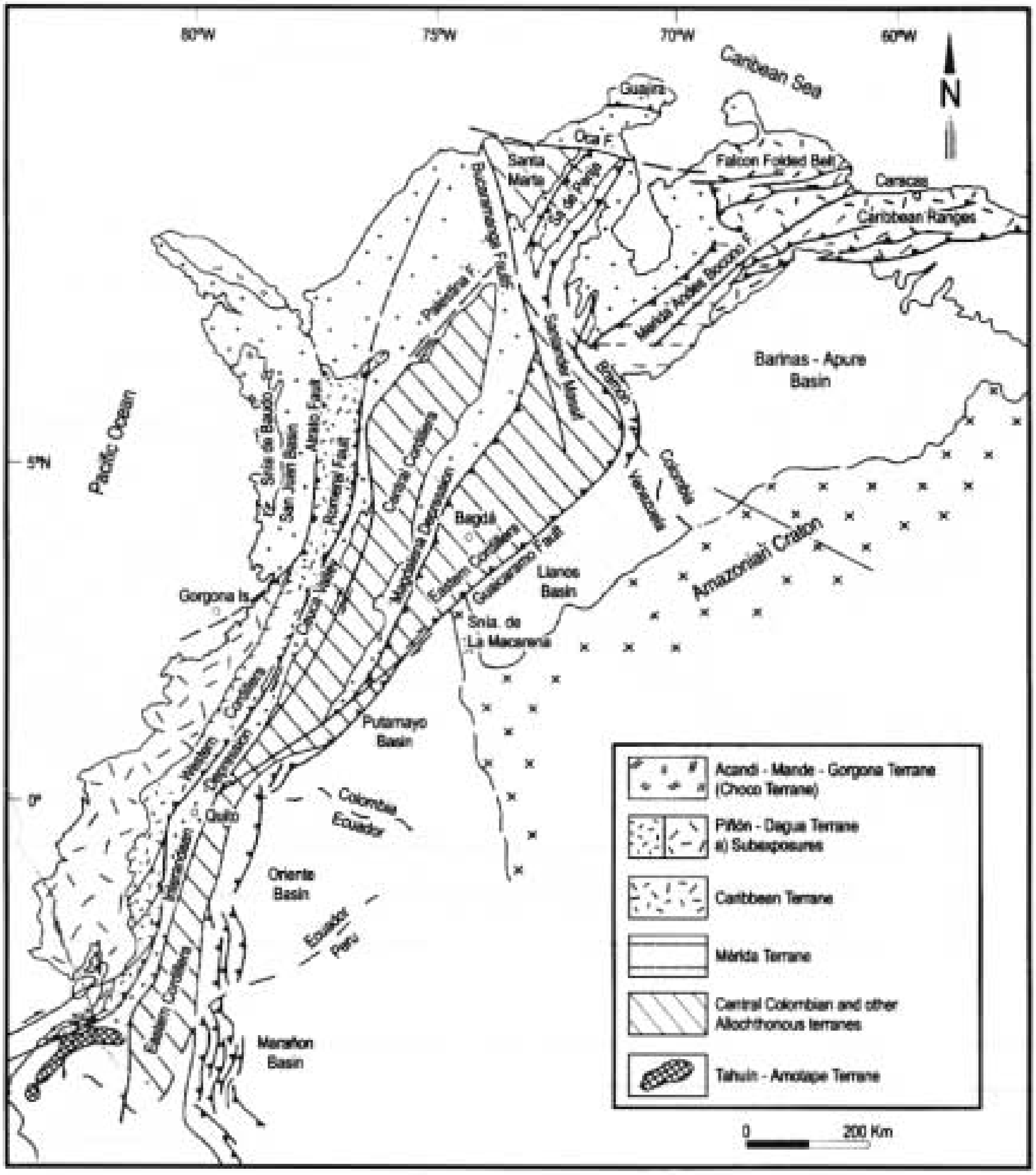


The last deformation episode, which created the present day Mérida Andes, commenced in the late Oligocene and continues to the present. Recent tectonics are dominated by uplift manifest by the tilting of Quaternary terraces (Meier *et al.*, 1987), and by studies of focal mechanism (Malavé and Suárez, 1993). The mountain belt is bounded on the SW by the Santo Domingo Reentrant of the Llanos Basin which abuts against the Santander Massif along the Bramón Fault (Laubscher, 1987). To the NE, the range terminates at the Barquisimeto Depression where it is replaced by the Caribbean or Coastal Ranges. Bellizzia and Rodríguez (1968) emphasized the dominant role of the Boconó Fault in the structural development of the northern Mérida Andes. Although Shagam (1975) described dominant vertical displacement along this fault, 80 km of right-lateral displacement has been reported, mostly occurring in recent times (Kellogg and Bonini, 1982; Stephan, 1982).

White (1985) published the first CDP seismic data

FIGURE 1 - Digital topographic map of the Northern Andes (from USGS web site). Note the tectonic boundaries among the Santa Marta Block, the Mérida Andes, and Eastern, Central, and Western Cordilleras of Colombia and Ecuador. Compare with the main structural provinces depicted in Fig. 2.

FIGURE 2 - Main structural provinces and terranes of the Northern Andes (based on Colleta et al., 1997; Bourgois et al., 1987; Restrepo and Toussaint, 1988; Litherland et al., 1994).



illustrating the structural style. Recently, Audemard (1991), De Toni and Kellogg (1993), Colletta *et al.* (1997), have made significant contributions to deciphering and understanding the structural styles of this foldbelt.

Tectono-stratigraphic evolution

Precambrian

Several units in the Mérida Andes have been ascribed to the Precambrian. Unfortunately, the isotope dates display large error brackets. The U/Pb data plotted on the concordia diagram (Burkley, 1976) were re-interpreted by Marechal (1983) who argued against ages older than the Bellavista Formation (650 - 580 Ma). These problematic units include zircon ages for the El Carmen Granite (891 ± 100 Ma), the Quiu Granite (1.061 ± 0.1 Ga), the El Cambur Granite (1.118 ± 0.25 Ga) and the Estanques Granite (1.601 ± 0.35 Ga) (Burkley, 1976). Robust U/Pb Precambrian ages have been reported for the Santander Massif to the S (Restrepo-Pace *et al.*, 1997). However, the lack of granulite precludes the presence of the Grenville Orogeny that is documented in Colombia (Alvarez and Cordani, 1980; Kroonenberg, 1982; Priem *et al.*, 1989; Restrepo-Pace *et al.*, 1997). Thus, these older ages, with large error brackets, only provide a clue to the age of the protolith.

Equally important is the large age error for the Colorado Massif, situated in the Caparo Block. Rocks from the Bellavista Formation described by Burkley (1976), gave 650 - 580 Ma (Bellizzia and Pimentel, 1994). These rocks consist of chlorite-muscovite quartz schist and metachert with subordinate amphibolite and olivine basalt. They underwent a metamorphic event in which there occurred widespread granitic plutonism and penetrative deformation associated with the El Topo Gneiss (660 Ma), the Valera Granite (593 ± 16 Ma) and the Rio Caparo Granite (615 ± 30 Ma). This

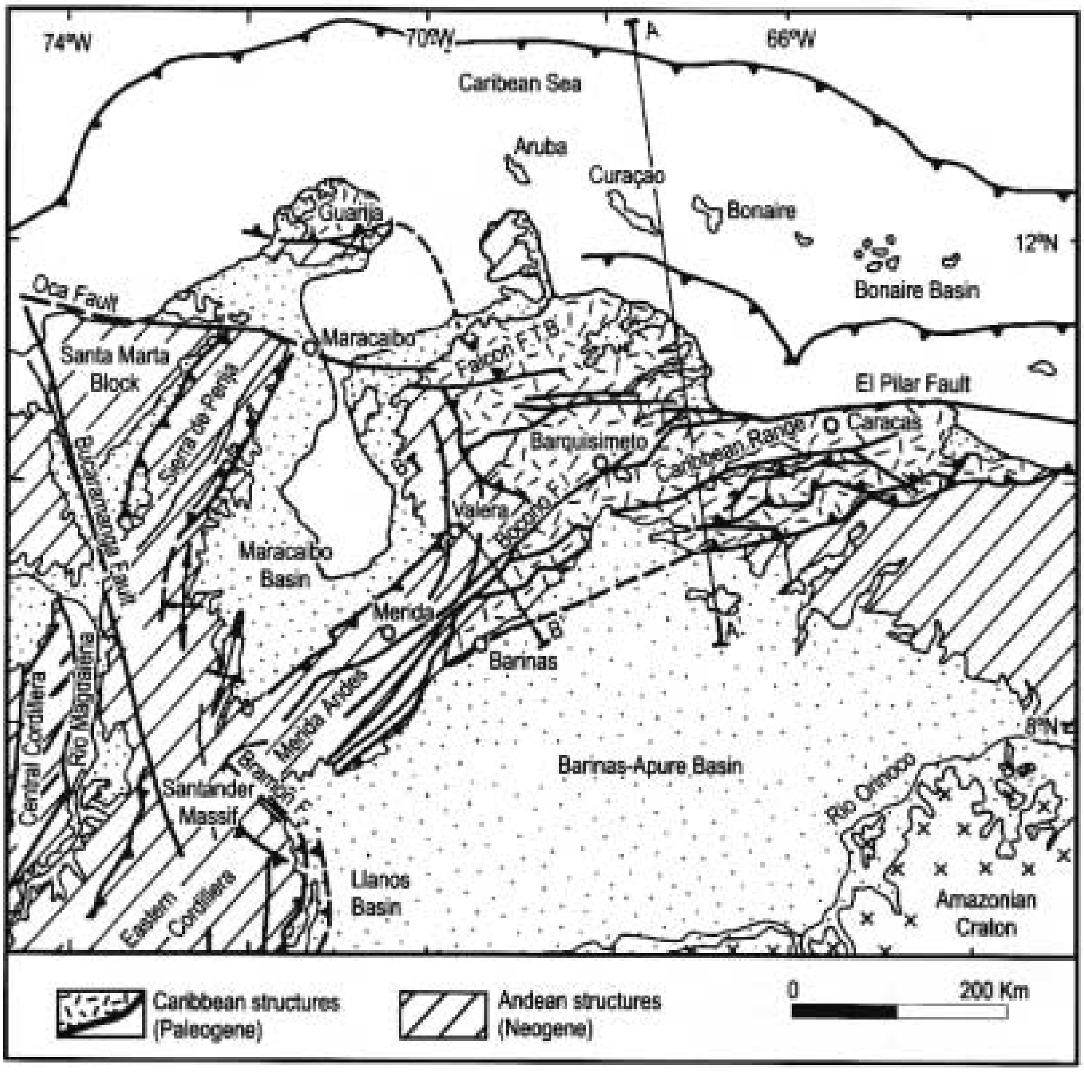


FIGURE 3 - Tectonic setting of the Mérida Andes and the Caribbean Range of Venezuela with main geologic units (based on Kellogg and Bonnini, 1982; Bosch and Rodríguez, 1992; Colleta et al., 1997).

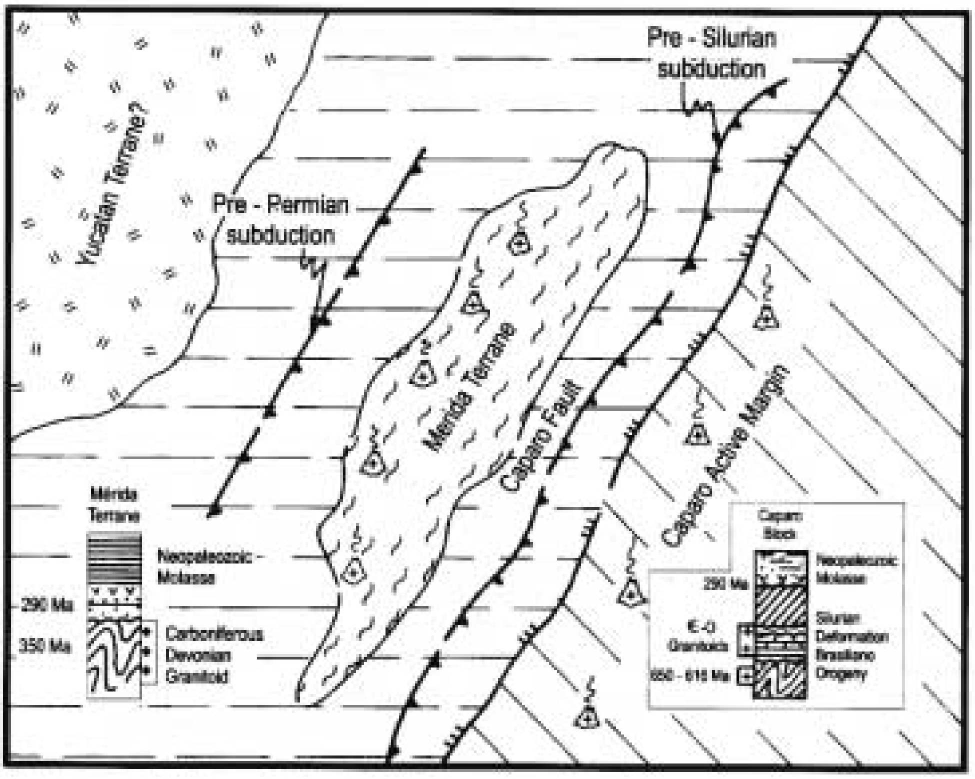


FIGURE 4 - Paleozoic tectonic setting of the Northern Andes of Venezuela (based on Bellizzia and Pimentel, 1994; Pindel, 1985).

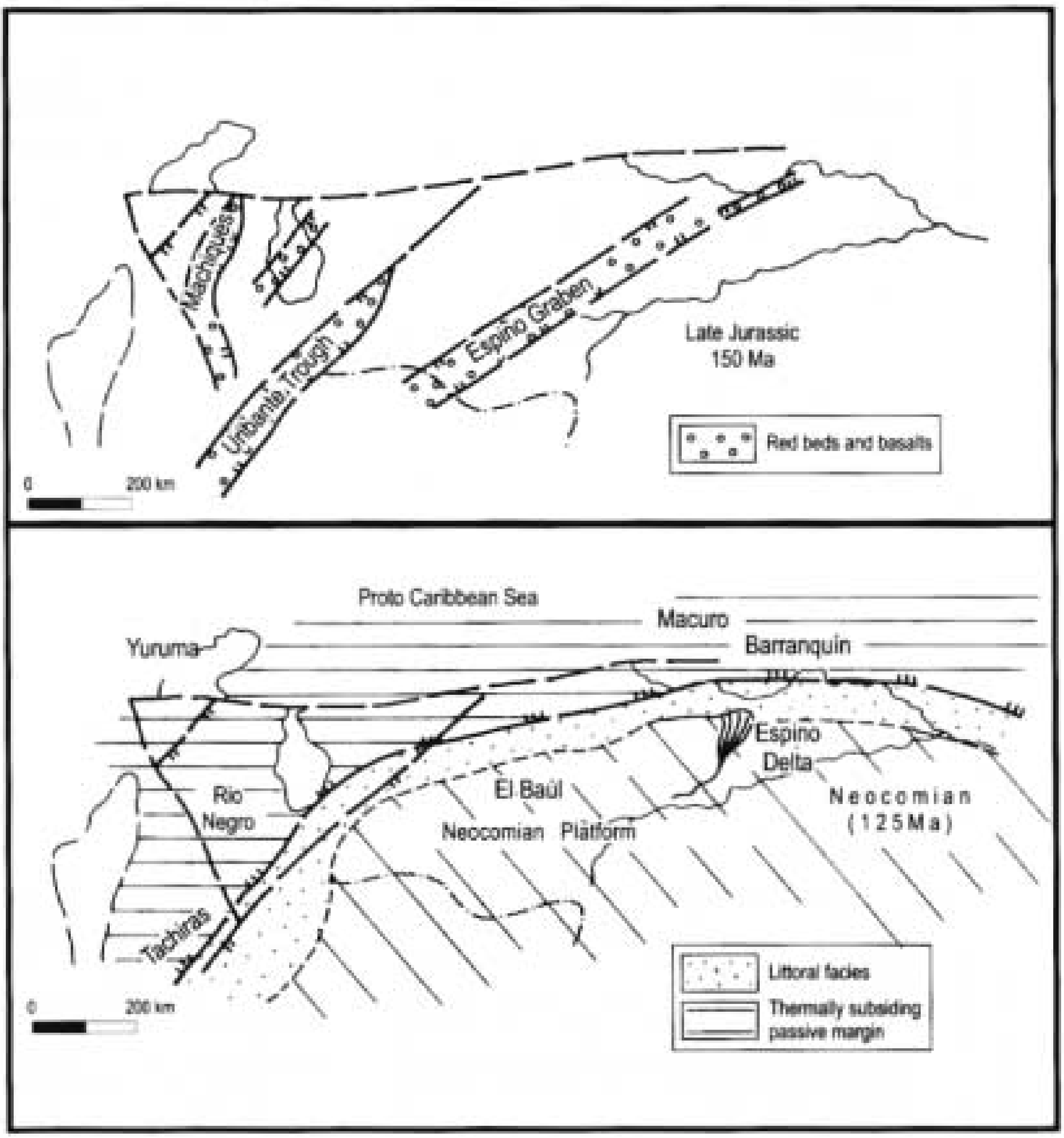


FIGURE 5 - Mesozoic paleogeography of Venezuela (based on Pindell and Barret, 1990; Bartok, 1993).

metamorphic event correlates in time with the Brasiliano Orogeny (Herz *et al.*, 1989).

Paleozoic

Bellitzia and Pimentel (1994) described two distinct geological provinces, separated by the Caparo Fault, and named them the autochthonous Caparo Block and the allochthonous Mérida Terrane (Fig. 4). Although paleomagnetic data to support this model are lacking, there are marked differences in pre-Sabaneta rocks across this fault. Thus, the Precambrian metamorphic rocks of the Caparo Block unconformably overlie 350 m of black shale, siltstone and sandstone of the Caparo Formation containing graptolites, trilobites and brachiopods of Late Ordovician age (Benedetto and Ramirez, 1982). Thinly laminated shale layers interbedded with siltstone, micaceous sandstone and siliceous limestone of the El Horno Formation in turn unconformably overlie the Caparo Formation. A rich fauna of graptolites, corals, trilobites and brachiopods (Boucout, 1972; Sánchez, 1984) indicates a Silurian age (Llandoverian to Ludlovian). These sedimentary rocks were intruded by syn-tectonic granites (Caparoensis Orogeny) from the Early Ordovician to Early Silurian (495 - 425 Ma), and by post-tectonic granites during the Devonian (Burkley, 1976; Marechal, 1983).

The basement of the allochthonous Mérida Terrane (Bellizzia and Pimentel, 1994) is made up of the Sierra Nevada Formation (Lower Paleozoic) consisting of three metamorphic facies, indicating a sedimentary protolith (Kovisars, 1971; Shagam, 1975; Bellizzia and Pimentel 1994; Marechal, 1983). These facies vary from quartz-feldspathic mica schist, sillimanite-bearing assemblages with rare staurolite, to micaceous quartz-feldspathic gneiss with rare sillimanite and garnet to hornblende-epidote amphibolite. Marechal (1983) has estimated temperatures for these rocks between 620 to 720 °C and pressures form 4.3 to 7 kbar. According to Marechal (1983), this formation changes stratigraphically upward to microfolded, bluish-grey quartz-feldspathic garnet schist, phyllite, amphibolite and metaconglomerate of the Tostosa Formation. This unit is followed by slate, thinly laminate phyllite interbedded with metasandstone, locally with conglomerate, crystalline limestone, metachert, felsic metavolcanic units and pyroclastic rocks of the Mucuchachi Formation (Bellizzia and Pimentel, 1994). The fauna and flora suggest a Carboniferous age, no older than Mississippian (Sánchez, 1984). These two units underwent deformation and metamorphism at the same time as the Sierra Nevada metamorphic facies, prior to Westphalian as determined by the concordant foliation planes (Marechal, 1983). This interpretation is based on the presence of unmetamorphosed Upper Paleozoic rocks in the Caparo Block as well as in the Mérida Terrane.

Subsequent docking of the Mérida Terrane was accompanied by molasse deposition of the Sabaneta Formation (Westphalian) consisting of interbedded red sandstone, shale and conglomerate (Thompson and Miller, 1949; Arnold, 1966; Sánchez, 1984). This formation is overlain by a sequence of interbedded sandstone, shale and fusulinid limestone of the Palmarito Formation (Stephanian to Guadalupian) (Pierce *et al.*, 1961; Sánchez, 1984). The Westphalian collision of the Mérida Terrane may have created salients and reentrants with different rheologies. Such heterogeneity controlled not only the relative subsidence rates but also the location and development of structural highs such as the orthogonal Mérida Arch (Lugo, 1994).

Mesozoic

A period of tectonic quiescence heralded the regional extension associated with deposition of varicolored beds of sandstone, shale, conglomerate and tuff of the La Quinta Formation. The Mérida Arch started to become active and continued active through the Early Cretaceous (Lugo, 1994). This was an important transfer zone during the structural development of the Mérida Andes. Jurassic grabens, trending NE-SW, (Fig. 5) are subparallel to the Uribante Trough (Renz, 1959). Felsic and basic volcanism, coeval with opening of the western Tethys Ocean and the Gulf of Mexico accompanied extension.

Passive margin sedimentation did not started until the Cretaceous (Barremian). It began with the deposition of coarse-grained, thick, cross-bedded arkosic to quartz-rich sandstone units of the Río Negro Formation (Zambrano *et al.*, 1971). The Cogollo Group (Aptian to Cenomanian) overlies this formation, which consists of interbedded limestone, shale and sandstone. Periodic flooding of the continental margin drowned the carbonate platform (Bartok *et al.*, 1981) with subsidence in the major troughs, active until the close of the Cenomanian. This accounts for the thick Lower Cretaceous section. The regional Turonian transgression was followed by periodic transgressions from the Coniacian to Early Santonian. Each transgression was marked by the deposition of black shales, rich in organic matter; and beds of limestone and chert of La Luna/Navay Formation (Tribovillard *et al.*, 1991; Parnaud *et al.*, 1995). An abrupt change in depositional style and rates occurred during Late Campanian-early Maastrichtian. Sediments consisted of thick shale units and thin bedded sandstone of the Colon/Mito Juan Formations which grade laterally to the Burguita Formation (Parnaud *et al.*, 1995), that consists mainly of sandstone.

Cenozoic

During the Caribbean Orogeny (Late Cretaceous to early Eocene), there occurred an orthogonal emplacement of allochthonous terranes and contemporaneous docking of the Central Cordillera of Colombia (Peruvian Phase), associated with the development of a dogleg-shaped foredeep. The trend of the foredeep changed from N-S to E-W on account of the orthogonal nature of these allochthonous terranes. Paleocene to early Eocene reactivation of the Mérida Arch suggests that it may have guided the collision-related thrusting of the Caribbean nappes (Lugo, 1994). Discrete unconformities within the thick molasse sequence of the Oroque Group (Paleocene) occur along the Mérida Andes, suggesting periodic orogenesis. Fluvio-deltaic deposits of sandstone, shale and coal seams of the Mirador/Gobernador and Carbonera/Paguey formations attest to a new pulse of molasse deposition during the Eocene. Marine shale and siltstone beds of the Leon Formation (Oligocene) overlie these deposits (Parnaud *et al.*, 1995).

Eastward movement of the Caribbean Plate from the middle Eocene (Pindell and Barret, 1990), provides a mechanism for the thrusting of the Carora-El Tocuyo nappes in the northern Mérida Andes (Stephan, 1982). These nappes consist of Paleocene to early Eocene flysch deposits and metamorphosed Cretaceous basinal rocks. Oligocene N-S overthrusting has been demonstrated from evidence found in the Guarumen-1S exploration well (Figueroa de Sánchez and Hernández, 1990). Such Caribbean nappes were often reactivated or refolded by younger Neogene Andean structures (Fig. 6a). Thus, an important boundary can be shown to occur across the Humocaro Fault where to the W, autochthonous and allochthonous structures have been overprinted by N-S to N45°E structures, coaxial with the Boconó Fault (Stephan, 1982). Stephan (1982) has estimated about 80 km of right-lateral displacement from the nappe offsets. Truncation of the Caribbean allochthon and foreland sequences S of the Andes suggests regional uplift at the Eocene/ Oligocene boundary.

The Miocene/Pliocene molasse in the southern foredeep, associated with deformation and uplift of the Mérida Andes (Fig. 6b), consists of fluvial conglomerate, sandstone and shale of the Parángula, Río Yuca and Guanapa formations. This southern flank shows both N and S verging basement-involved structures (Audemard, 1991; Colletta *et al.*, 1997). The equivalent molasse of the northern flexural basin has also a distinct fluvial facies of the Chama/Palmar, Isnoto and Betijote formations (Colletta *et al.*, 1997). However, the structural style consists of N-verging thrusts, mainly detached from the pre-Cretaceous substratum, that form an antiformal stack (Colletta *et al.*, 1997). Secondary detachment zones have also been reported in the Upper Cretaceous and Tertiary strata (Audemard, 1991). Multiple unconformities and fining-upward cycles in the molasse sequence record periodic deformation and uplift of this Andean segment. Indeed, a late Miocene intra-molasse unconformity heralds the beginning of the frontal monocline formation and the period of greatest uplift (De Toni and Kellogg, 1993).

According to Colletta *et al.* (1997), Plio-Quaternary

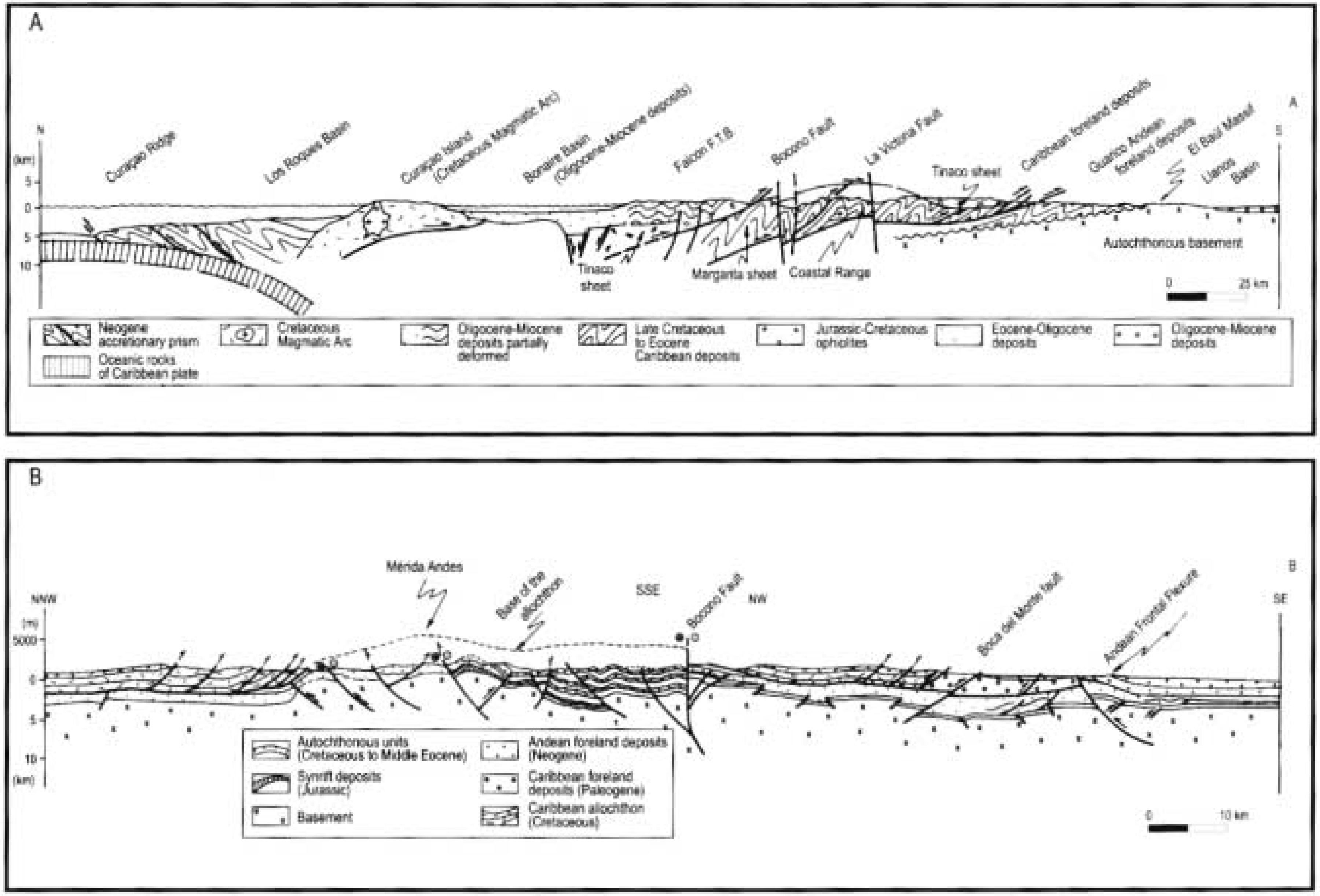


FIGURE 6 - Regional cross-sections of the Venezuelan Andes. a) Caribbean Ranges (after Stephan, 1982). b) Mérida Andes (after Hooper et al., 1999). Location in figure 3.

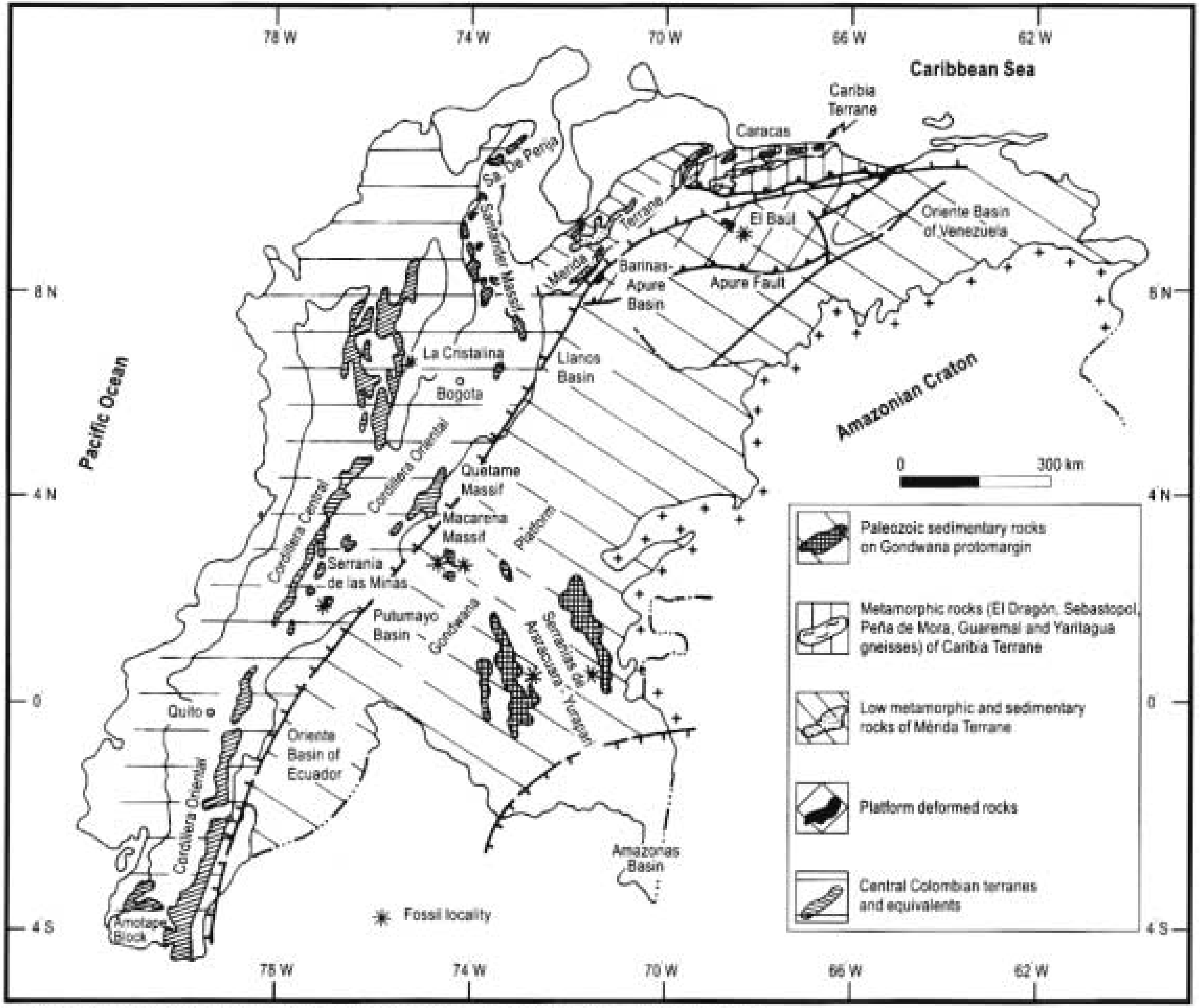


FIGURE 7 - Paleozoic basement of the Northern Andes and accreted terranes (based on Bellizzia and Pimentel, 1994).

strata pinch out against the Andean tectonic wedge. During the Pleistocene, a sequence of fluvio-glacial sandstone and conglomerate beds was deposited along the major valleys, and some of these were periodically uplifted to form distinct terraces. Other sequences have been used to infer Recent displacement along the Boconó Fault (Schubert, 1984).

Crustal growth and tectonic development of the Venezuelan (Mérida) Andes

Basement isotope ages in the Mérida Andes, S of the Caparo Fault, suggest the presence of older protoliths and perhaps a northern extension of the Guiana Shield. However, crustal growth processes did not commenced until the Neoproterozoic, as can be determined by the presence of chlorite-muscovite-quartz schist and metachert with subordinates amphibolite and olivine basalt of the Bellavista Formation (Burkley, 1976; Marechal, 1983; Bellizzia and Pimentel, 1994). The Bellavista Formation is associated with the Valera Granite (593 ± 16 Ma), Río Caparo Granite (615 ± 30 Ma) and El Topo Gneiss (660 Ma), and may correlate in time with the Brasiliano Orogeny (Herz *et al.*, 1989).

Discordantly overlying the basement rocks are Ordovician and Silurian strata with Gondwana-South American faunas (Benedetto and Ramírez, 1982; Benedetto *et al.*, 1992). Intense Lower Paleozoic magmatic activity (Burkley, 1976; Marechal, 1983) may be related to the Andean-type of margin during this time, and may represent the northernmost influence of the Ocloyic Orogen. Devonian age granites imply the proximity of Laurasia to Gondwana (Fig. 7), as supported by the fauna and flora described in the Perijá Mountains (Berry *et al.*, 1997). However, Mérida Terrane docking did not take place until the Westphalian as suggested by the presence of the Sabaneta and Palmarito formations. This collision event was an important episode of crustal growth accompanied by variable Barrowian metamorphism and penetrative deformation of the Mucuchachi and Tostosa formations. Tectonic transport, indicated by stretching lineation of originally horizontal folded structures, suggests a shear couple deformation during terrane emplacement (Kovisars, 1971; Marechal, 1983; Bellizzia and Pimentel, 1994). Culmination of this tectonic event involved development of long cylindrical folds in post-Westphalian rocks with NE-SW upright-oriented axis, or showing slight overturning toward the NW. This deformation was associated with widespread calc-alkaline plutonism from Early Permian to Triassic (290 - 225 Ma).

NW-SE Jurassic extension was coeval with opening of the western Tethys Ocean and Gulf of Mexico, and was marked by an early phase of passive margin development (Pindell and Barret, 1990; Bartok, 1993). Main depositional grabens were oriented subparallel to the present day Mérida Andes which follow the trend of the Uribante Trough (Renz, 1959). These grabens were filled with varicolored siliciclastic sediments, interbedded with felsic and basic volcanic units. However, accommodation zones such as the Mérida Arch were orthogonal to the mountain belt (Lugo, 1994). A fully developed passive margin was achieved during Early Cretaceous while high subsidence rates continued along extensional grabens until the Cenomanian. Late Cretaceous sedimentation took place in the setting of a passive margin. Nonetheless, Paleogene molasse deposition was mainly related to the interplay between thrusting of the Caribbean nappes and deformation in the Colombian Andes. The orthogonal Mérida Arch has been periodically active through early Eocene suggesting reactivation of pre-existing faults (Lugo, 1994)

Deformational processes responsible for the formation of the present-day Mérida Andes did not begin until the late Oligocene and were manifest by a double vergence foldbelt and foredeeps. According to Colletta *et al.* (1997), this paired foldbelt represents relatively minor intraplate readjustment between the Eastern Colombian Cordillera and the South Caribbean transform creating conjugate foreland basins and foothills. During the Neogene, the Maracaibo Block detached from the South American mainland to accommodate most of the deformation at the triple junction between the Pacific, Caribbean oceanic domains, and South America. Colletta *et al.* (1997) explained the surface thrust fronts parallel to the plate boundary, strike-slip motion in the allochthon along the Boconó Fault and asymmetric wedging at depth in terms of strain partitioning during Neogene oblique convergence. The large negative gravity anomaly in the northern Andean trough (- 150 mGal) suggests that this foredeep and the Andes are not in isostatic equilibrium (Kellogg and Bonini, 1982). This view supports the possibility of a SE-dipping subduction of parts of the Maracaibo continental lithosphere (Colletta *et al.*, 1997). However, the southern trough lacks a major gravity anomaly and its syn-orogenic sequence does not fill a flexural foredeep, but rather constitutes a gentle NW-dipping monocline recording Recent Andean uplift (Colletta *et al.*, 1997).

Deratmiroff (1971) mapped low-angle thrusts along the central segment of the northern Mérida front and between Torondoy and Valera. Castrillo and Hervouet (1996) have described complex imbricate thrusting and thin-skinned thrusting associated with strike-slip faults. Recently, Colletta *et al.* (1997) described the puzzling NE Brujas High that involves inversion of Jurassic grabens along high angle faults. Dominant frontal triangle zones are characterized by complex duplexes decoupled in the basement as well as in the Colón Formation (Late Cretaceous) and in Paleogene and Neogene shale beds (Deratmiroff, 1971; Audemard, 1991; De Toni and Kellogg, 1993; Colletta *et al.*, 1997). Miocene syn-flexural strata and backthrusting were involved in the shortening and wedging out toward Lake Maracaibo. However, Plio-Quaternary strata pinch out against the Andean tectonic wedge (Colletta *et al.*, 1997).

Although the bulk of shortening took place during the Miocene, there is still significant present day deformation in the Mérida Andes as recorded in tilted terraces, earthquake focal mechanism and apatite fission track data (Schubert, 1984; Kohn *et al.*, 1984). Thus, uplifting, tilting and lateral displacement of these terraces (Meier *et al.*, 1987; Schubert and Vivas, 1993) and focal mechanism studies (Malavé and Suárez, 1993) have been used to document current tectonic activity. Systematic determinations of the sense and magnitude of the displacement of Quaternary sediments and topographic features along the Boconó Fault have demonstrated vertical and right-lateral displacements exceeding 250 m. To the S, the Bocono Fault abuts against the Punzón de Pamplona reverse faults which ties in with the Eastern Cordillera Frontal Fault System (Beltrán, 1994). Crustal-scale balanced cross-sections have helped to constrain about 60 km of Neogene shortening (Colletta *et al.*, 1997).

The Colombian Andes

Several benchmark papers have been published on the Colombian Andes, including work by Nelson (1957), who made the first traverse across the Central and Western Cordillera from Ibague to Cali; and by Burgl (1961) who summarized his work on the stratigraphy and geological history of Colombia. Radeli (1967) published one of the most comprehensive reviews on the geology of the Colombian Andes based upon his extensive fieldwork. Other important works on the structural and stratigraphic evolution of the mountains include Irving (1971), Julivert (1973), Toussaint (1978), Etayo-Serna and Barrero (1983), and Restrepo and Toussaint (1988).

The evolution of the Colombian Andes encompasses a complex amalgamation of multiple allochthonous terranes in time and space. The mountain belt consists of three distinct and separate chains: the Western, Central, and Eastern Cordillera (Fig. 8), bounded to the E by the Borde Llanero Fault System (Forero-Suarez, 1990) which marks the boundary with the Proterozoic basement rocks of the Guayana Shield. The Western Cordillera consists mainly of Cretaceous tholeiitic basalt and deep-water sedimentary facies resting on oceanic crust (Barrero, 1979). Ophiolitic rocks have been reported from the Baudó Range in the Western Cordillera up to the western flank of the Central Cordillera (Toussaint, 1978). The Romeral crustal-scale fault zone contains lawsonite-glaucophane schist and eclogite (MacCourt and Feininger, 1984). The Romeral Fault separates the Western and Central Cordillera and has been interpreted as a suture or subduction zone (Toussaint and Restrepo, 1982). The Central and Eastern cordilleras are underlain by continental crust. Thus, the basement of the Central Cordillera consists of metapelite and metavolcanic units known as the Central Andean Terrane, with sedimentary cover ranging in age from Paleozoic to Tertiary. Whereas the Central Cordillera has undergone pervasive plutonic and magmatic activity with evidence for migration and widening of the volcanic arc, the Eastern Cordillera lacks extensive volcanism except for isolated, subordinate Cretaceous basic volcanism.

Tectono-stratigraphic evolution

Precambrian

Grenville age rocks have been reported in the Garzón, Santander and Sierra Nevada de Santa Marta massifs (Fig. 9). Perhaps these rocks also occur in the Serrania de Perijá (Tschanz *et al.*, 1974; Alvarez and Cordani, 1980; Kroonenberg, 1982; Priem *et al.*, 1989; Restrepo-Pace *et al.*, 1997). Noteworthy, also is the presence of granulite xenoliths in lava of the Nevado del Ruiz Volcano (Jaramillo, 1978). Restrepo and Toussaint (1978) have reported garnetiferous amphibolite from Caldas, on the eastern flank of the Central Cordillera, yielding a single hornblende K/Ar age of 1.67 ± 0.5 Ga. Trumpy (1943) reported high-grade metamorphic rocks, without geochronological support, in the Serrania de Macarena.

Radelli (1962) made some of the first petrologic descriptions of the rocks of the Garzón Massif and later Kroonenberg (1982) distinguished two petrogenetic units. The dominant Garzón Group consists of banded charnockitic and garnetiferous granulite, mafic granulite, and amphibolite. However, the less extensive Guatopon and Mancagua hornblende-biotite augengneiss are concordantly foliated with the hosting Garzón Group and are interpreted as metamorphosed syn-tectonic granites (Kroonenberg, 1982). Intercalation of calc-silicates and pelitic gneiss with this granulite suggests a sedimentary protolith (Kroonenberg, 1982; Priem *et al.*, 1989). Alvarez and Cordani (1980) have obtained a four points Rb/Sr isochron of 1.18 Ga in these granulite supracrustal rocks. Six of eight samples from this massif define a line corresponding to an age of 1.172 ± 0.09 Ga. However, there are also abundant 1.56 - 1.45 Ga old augen gneiss related to the underlying Parguazan tectonomagmatic event observed in the Guayana Shield (Priem *et al.*, 1982, 1989). The age of the protolith is confirmed by calculated Nd model ages showing a consistent average of *c.* 1.55 Ga (Restrepo-Pace *et al.*, 1997).

Slivers of Orinoquian age (*c.* 1.1 Ga) basement rocks outcrop in the northern part of the Santander Massif (Ward *et al.*, 1973). They include two distinctive petrotectonic units: a low to medium-grade metamorphic metapelitic-metapsammitic unit, and a structurally concordant foliated granite orthogneiss unit with intrafoliated amphibolite dykes displaying tholeiitic chemical signatures. Mineral paragenesis suggest low medium pressure (3 - 9 kbar) and high temperature (300 - 700 °C) greenschist to upper amphibolite facies rocks with local kyanite, implying medium pressures in restricted areas (Restrepo-Pace, 1992; Restrepo-Pace *et al.*, 1997). In like manner to the Garzón Massif, the protolith of these gneiss are considered to be sedimentary rocks.

The isolated, triangular massif of the Sierra de Santa Marta contains Precambrian rocks which consist of quartz-perthite granulite; and intermediate, mafic, ultramafic, calcareous and garnetiferous granulite with mineral paragenesis similar to the Garzón Massif (Tschanz *et al.*, 1974). Strong Rb/Sr isochrons of 1.37 - 1.27 Ga have been recalculated from gneiss (MacDonald and Hurley, 1969; Kroonenberg, 1982). Furthermore, whole rock Rb/Sr data of 1.273 Ga and 736 Ma have been reported from granulite as well as a hornblende K/Ar age of 949 ± 36 Ma (Tschanz *et al.*, 1974). In the Guajira Peninsula, 200 km NE of Santa Marta, a 1.25 Ga U/Pb zircon age has been determined on the Jojocinto Leucogranite-Gneiss (Case and MacDonald, 1973). Calculated Nd model ages show a consistent average of *c.* 1.7 Ga, somewhat older than the Garzón samples (Restrepo-Pace *et al.*, 1997).

Although Kroonenberg (1982), precluded a long metamorphic crustal history for the Garzón and Santa Marta massifs, based on low initial strontium isotope ratios, he recognized an important metamorphic event around 1.2 Ga

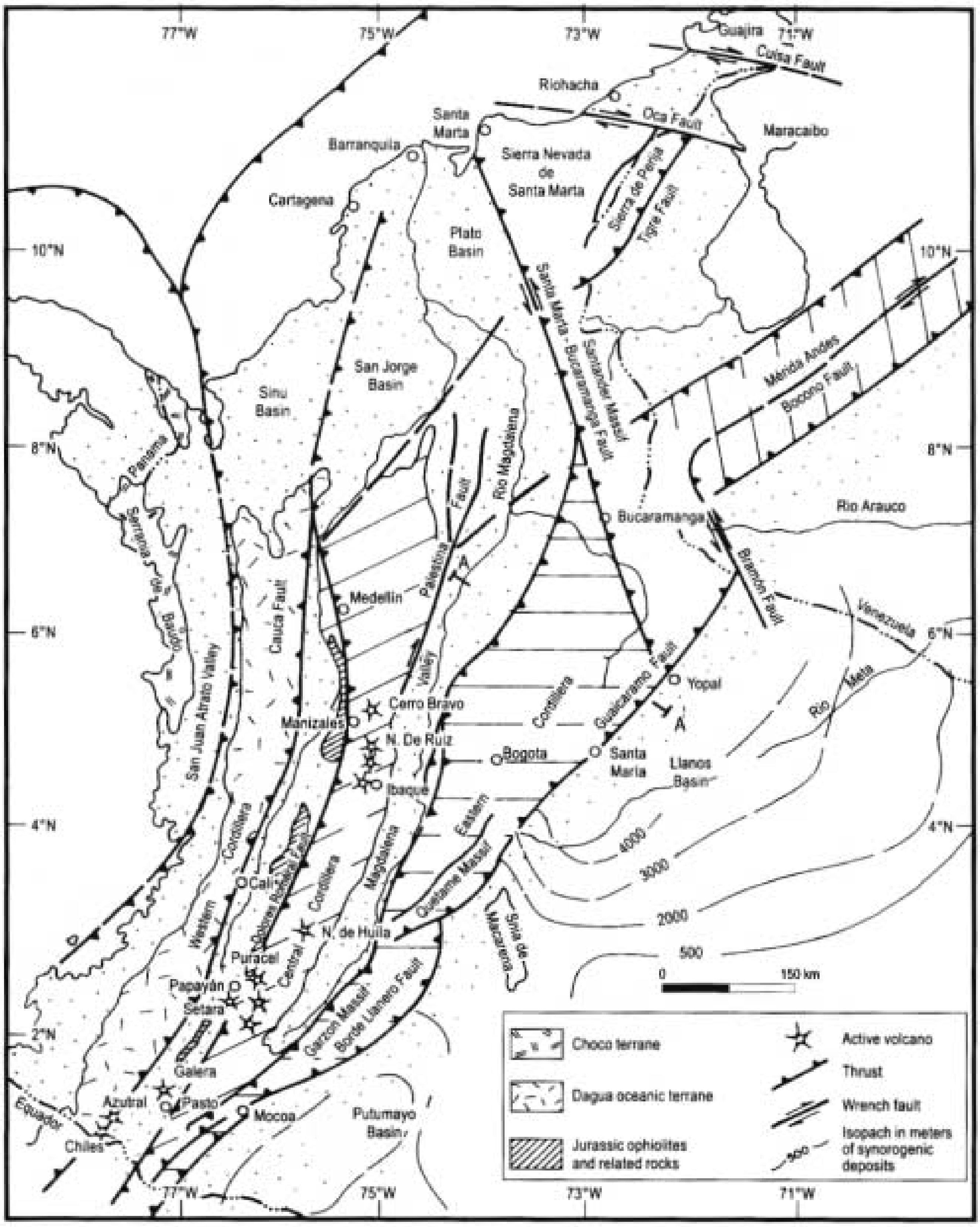


Fig. 8 - Major geologic provinces of the Colombian Andes (based on Bourgois et al., 1987; Restrepo and Toussaint, 1988; Forero-Suárez, 1990).

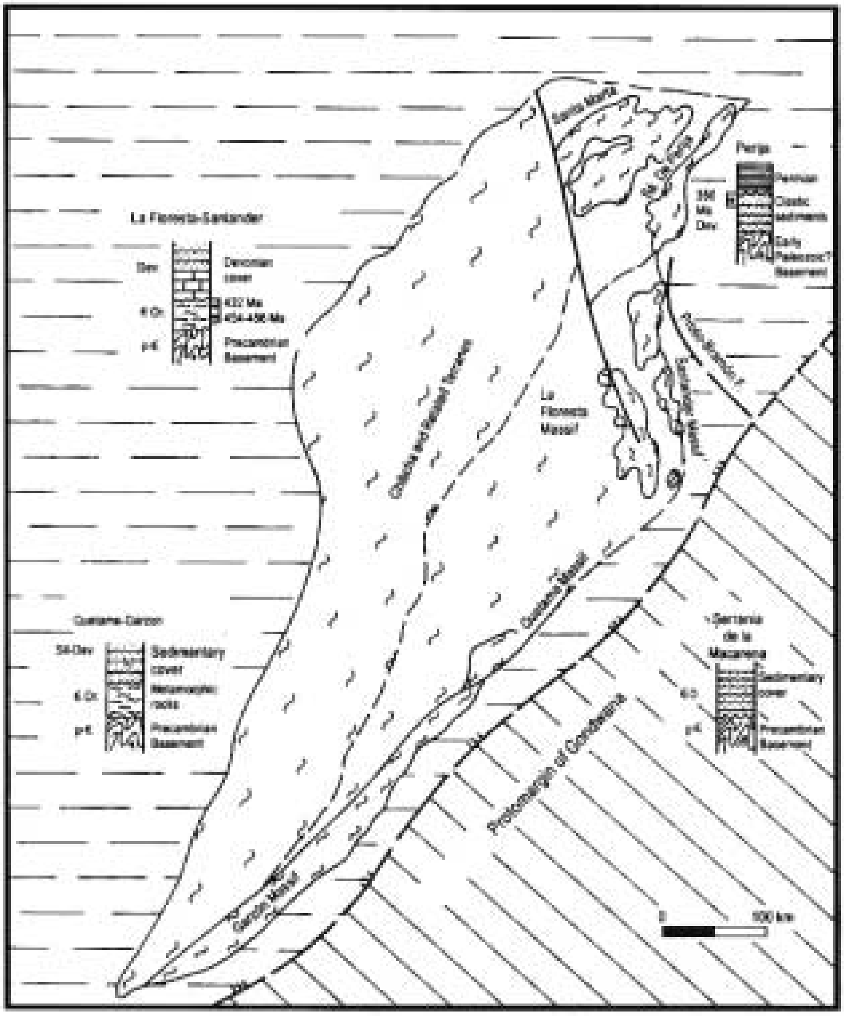
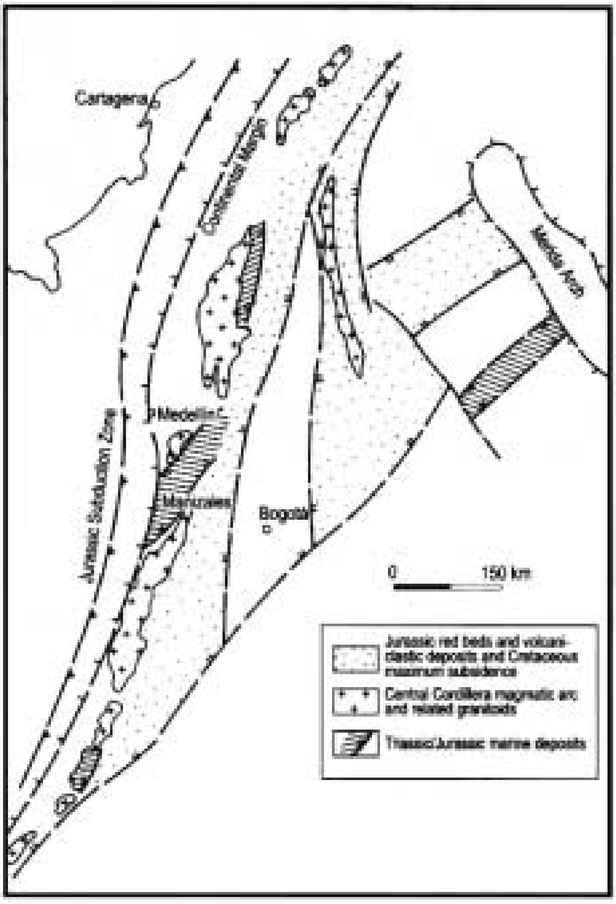


FIGURE 9 - Paleozoic tectonic setting of the Colombian Andes with main Precambrian outcrops. The Chibcha (or Central Colombian) and related terranes have docked to the proto-margin of Gondwana during the Late Ordovician (based on Etayo-Serna and Barrero, 1983; Toussaint, 1993). The Santa Marta, Santander, and Perijá blocks have been palinspatically restored to their Paleozoic position.

FIGURE 10 - Mesozoic tectonic settings of the Colombian Andes (based on Mojica et al., 1996). Maximum Cretaceous sudsidence was located along Jurassic grabens, and continued until Cenomanian.



along the Guaiana Shield. He interpreted this granulite belt as having formed in the western border of the Paleo to Mesoproterozoic nucleus of the Guiana Shield by an orogenic event around 1.2 - 1.4 Ga (Orinoquian). Thus, the lithology of the belt matches an ensialic calc-alkaline volcanic suite with minor pelitic, semipelitic and calcareous intercalations, probably deposited in a shallow marine environment, and cut by small ultramafic and anorthositic intrusives (Kroonenberg, 1982).

In contrast, Priem *et al.* (1989) recognized a *c.* 1.6 Ga old augen gneiss (Guatopon and Mancagua) in the Garzón Massif, similar to that observed in the adjacent parts of the Guiana Shield (Parguasan tectonomagmatic episode). They also described a supracrustal sequence metamorphosed in the granulite facies around 1.2 Ga (Orinoquian Orogeny), and a set of pegmatite dykes (*c.* 850 Ma). The low initial strontium isotope has been interpreted, alternatively, as an indication that the metamorphism took place shortly after deposition. Thus, most of the sporadic Precambrian slivers distinguished thus far along the Colombian Andes have shared the same tectonic history as the Garzón Massif. Because the Orinoquian Orogeny (1.2 Ga) trended subparallel to the present Andes (Litherland *et al.*, 1985), it was originally interpreted to represent a continental collision between the western margin of the Guiana Shield and the eastern margin of the Canadian Shield (Priem *et al.*, 1989). However, an alternative interpretation suggests a closer relationship with the Oaxacan Complex of southern Mexico (Keppie and Ortega-Gutierrez, 1995; Restrepo-Pace *et al.*, 1997).

Paleozoic

In the Serrania de la Macarena, Trumpy (1943) reported Middle Cambrian to Lower Ordovician trilobites, brachiopods and graptolites in the Güéjar Group (Fig. 9); also described by Harrington and Kay (1951). This group consists of dark mudstone, phyllite to slate interbedded with quartz-rich sandstone and greywacke. Trumpy (1943) speculated that the Quetame Group might be the metamorphic equivalent of the Güéjar Group. Radelli (1967) later confirmed this. To the N, quartzite, micaschist and gneissic schist of the Perijá Formation; and to the W, micaceous gneiss, mica schist, phyllite and marble of the La Ceja Group in the Central Cordillera, have been also ascribed to the Cambro-Ordovician and may correlate with the Güéjar Group. Villaroel *et al.* (1997) described on the western flank of the Eastern Cordillera a sequence of dark grey micaceous shale beds, interbedded with arkósic sandstone and polymictic conglomerate beds of the Venados Formation (Middle Ordovician).

To the E of the Borde Llanero Fault, Ulloa *et al.* (1982) described trilobites, brachiopods and graptolites in grey to black micaceous shale beds that grade upward to interbedded siltstone and sandstone units with subordinates limestone members of the Lower Negritos Formation (Lower Ordovician). Along the Serrania de Las Minas, in the Central Cordillera, Mojica *et al.* (1988) reported graptolites and trilobites in fine grained arkosic sandstone beds, that grade upward to dark grey to black shale beds of the Higado Formation (Middle Ordovician), informally described as La Cristalina Formation. These units may be equivalent to the graphitic schist of the Cajamarca Group (Nelson, 1957; McCourt *et al.*, 1984). Structurally, they are seen as well-developed isoclinal folds with penetrative schistosity, including crenulation cleavage indicative of a polyphase deformational history.

This deformation could be related to a major orogenic event at the close of the Ordovician, correlated with the Ocloyic Orogeny of northern Argentina (Ramos, 1988). The deformation accounts for a pronounced angular unconformity between Cambro/Ordovician rocks and the Devonian basal conglomerates. In the Santander Massif, another major unconformity also separates Middle Devonian siliciclastic rocks from potentially laterally equivalent metamorphic Ordovician rocks (Boinet *et al.*, 1986). Indeed, new isotope ages from the Santander Plutonic Group suggest Late Ordovician (457 Ma) to Early Devonian (413 Ma) for some of the granites, and a Late Ordovician age (456 ± 22.8 Ma) for one gabbro (Boinet *et al.*, 1985). Late Silurian (Ludlovian) palynomorphs have recently been described by Grösser and Prössl (1991) in interbedded shales of the Areniscas de Guttierrez of the Quetame Massif. These were originally ascribed to the Middle Devonian (Renzoni, 1968).

Devonian rocks are restricted to the Eastern Cordillera and its northward extension in the Sierra de Perijá. However, in the Central Cordillera, near the town of Rovira, a sequence of black shale beds with irregular interbedded sandstone units of the Amoya Formation contains Middle Devonian palynomorphs (Prössl and Grösser, 1995). To the E, in the Santander Massif, the Floresta Formation consists of a basal conglomerate that grades upward to interbedded quartz-rich sandstone and shale units with abundant Early to Middle Devonian fauna (Boinet *et al.*, 1986; Barrett, 1988). According to Boinet *et al.* (1986), the Floresta Formation is also intruded by the Onzaga Monzonite (394 ± 23 Ma). Northwards, in the Serrania de Perijá, the thick Rio Cachiri Group (Devonian) consists of more than 1300 m of interbedded shale and sandstone with a few biostromal limestone units containing a Givetian to Frasnian fauna and flora (Berry *et al.*, 1997). It is noteworthy that Late Devonian isotopic ages have been reported for the Las Lajas Granite (370 ± 20 Ma).

Carboniferous rocks are also well represented in the Eastern Cordillera and its northern extension. Late Carboniferous fauna has been described in interbedded sandstone, conglomerate and graphitic shale units of the Gachala Formation in the Quetame Massif and in the Garzón Massif. To the N, in the Santander Massif the stratigraphic thickness increases to a maximum of 2900 m of red colored sandstone and shale with subordinated limestone and conglomerate beds known as the Labateca Formation. Farther to the N, in the Serrania de Perijá, a sequence of thin-bedded black shales interbedded with thick massive fossiliferous limestone members and varicolored siltstone and sandstone beds of the Tinacoa Formation (Carboniferous) is reported.

Permian rocks are absent in the Garzón and Quetame massifs. However, in the Garzón Massif, Mojica *et al.* (1988) reported limestone members of possible Wolfcampian age. In the Santander Massif, fine-grained sandstone and shale is interbedded with thick-bedded fusulinid-bearing limestone units of the Diamante Formation. These sediments are overlain by massive limestone conglomerate beds of the Tiburón Formation of Lower Permian age (Ward *et al.*, 1973). To the N, in the Sierra de Perijá, siltstone and sandstone beds grade upward to dark grey fossiliferous limestone units of the Palmarito Formation (Late Wolfcampian to Early Leonardian; Thompson and Miller, 1949). The northernmost Permian outcrops are in the Sierra Nevada Massif, and consist of crystalline limestone and marble units with wollastonite interbedded with quartz-rich sandstone and conglomerate that Gansser (1955) has assigned to the uppermost Paleozoic. These Permian rocks have been intruded by Late Permian to Early Jurassic calc-alkaline granites varying from I-type to S-type. The S-type granites are related to shear zones before Late Triassic separation of North and South America (Pindell and Barrett, 1990). Indeed, amphibolite and sheared ultrabasic pods in the Rosario Complex along with the metagabbro and amphibolite of the Bolo Azul Complex, may represent the accretion of Upper Paleozoic rocks onto the South American Craton along the Palestina Fault (Feininger, 1970; McCourt and Feininger, 1984).

Mesozoic

A small, foliated, calc-alkaline plutonic precursor of Late Triassic age situated in the Magdalena Valley was related to ensialic back-arc extension similar to that proposed for the Choyoi Group in Central Argentina (Ramos and Kay, 1990). Indeed, Jurassic isotope ages (163 ± 10 Ma and 131 ± 9 Ma) for the Cauca Ophiolitic Complex (Restrepo and Toussaint, 1973) and the blueschist facies associated with the Romeral Fault System provide evidence for a subduction zone. They also explain the presence of abundant Jurassic batholiths in the Eastern Cordillera (Fig. 10). The NE-SW oriented Triassic/Jurassic grabens (Cediel, 1981; Mojica *et al.*, 1996) are filled with pre-Norian arkosic sandstone beds, polymictic conglomerate and breccia (Luisa Formation); dark grey limestone units, locally fossiliferous (Payandé Formation); Rhaetic to Middle Jurassic pyroclastic, ash and lapilli tuff deposits, and rhyolitic to dacitic agglomerates (Saldaña Formation). This sequence is cut by several diorite and gabbro bodies coeval with the San Augustin and Gallego Granites (172 - 159 Ma). In the Santander Massif, to the E, varicolored mudstone beds, siltstone, and sandstone units interbedded with welded tuff and conglomerate of the Jordan Formation (Late Triassic) filled the graben. In turn, varicolored conglomerate beds, sandstone and shale beds of the Girón Formation overlie this unit. It is locally intruded by the Aguablanca (196 ± 7 Ma) and Mogotes (193 ± 6 Ma) batholiths (Ward *et al.*, 1973).

In the Serrania de Perijá, NW-SE oriented grabens are filled by a sequence of varicolored conglomerate units, interbedded with arkosic sandstone, tuff and mudstone of the La Quinta Formation. This formation is also present in the Guajira Peninsula where polymictic conglomerate beds change from predominantly volcanic to mainly granitic and metamorphic with basaltic andesite and abundant ash layers near the base (Maze, 1984). The northern and southern Central Cordillera contains Late Triassic grabens filled with red beds (El Sudán Formation) and overlain by Jurassic dark grey shale, interbedded with volcaniclastic sediments (Geyer, 1980). However, the central part of the Central Cordillera contains abundant Early Triassic to Jurassic stocks and plutons that are also present in the Eastern Cordillera from the Garzón Massif to the Sierra Nevada (Maya, 1992). Mantle derived, calc-alkaline Jurassic plutonism is also present in the western margin of the Central Cordillera such as at the Mocoa (180 - 170 Ma), Segovia (160 Ma) and Ibague (150 - 140 Ma) batholiths.

A new pulse of ensialic extension took place during Early Cretaceous in two separate grabens. To the E, and to the S of the Machiques Trough, a new graben developed between the paleo Guaicáramo and Chiscas faults and was filled by Berriassian to Aptian conglomerate, sandstone and shale. To the W, the graben along the Mundo Nuevo Syncline continued to subside and was filled by Late Valanginian transgressive sandstone units of the Tambor/Yavi formations (Fabre, 1987; Etayo-Serna, 1979). This period of crustal extension was perhaps related to an increase in the subduction angle and concomitant trenchward volcanic arc migration from Jurassic to Late Cretaceous (McCourt and Feininger, 1984, Toussaint and Restrepo, 1982). A regional sagging phase took place during the Aptian, and was accompanied by widespread subsidence and deposition in the Eastern Cordillera. Regional flooding during the Albian resulted in the deposition of dark grey shale beds and limestone units. In the Cenomanian to Turonian there occurred the deposition of limestone and shale in a high-upwelling regime during a series of transgressive pulses (Villamil and Arango, 1998). Maastrichtian regressive sandstone beds were associated with the rise of the Central Cordillera.

To the W, a marginal basin to island-arc tholeiitic assemblage developed between the Central Cordillera and the Baudó Range (Bourgois *et al.*, 1987; Spadea and Espinosa, 1996). Aptian to Senonian deposition in this basin included siliceous and carbonaceous phyllite and slate of the Dagua Group; basaltic lava flows, hyaloclastic breccia, dolerite, gabbro and minor pyroclastics of the Diabase Group; and a very thick sequence of subaqueous pyroclastics and volcaniclastic sandstone beds, interbedded with mudstone and chert of the Espinal Formation. Interpretation of this suite of rocks vary from oceanic island arc (Barrero, 1979) to ocean floor (Pitchler *et al.*, 1974) and oceanic flood basalt (Millward *et al.*, 1984). Late Cretaceous tectonic escape and formation of the Caribbean Plate (Stephan *et al.*, 1990) caused obduction of this marginal basin sequence as indicated by complex folding and thrusting. Alpine-type nappes were thrusted from NW to SE with development of recumbent folds, isoclinal folding and crenulation cleavage (Bourgois *et al.*, 1987). Indeed, according to Burgois *et al.* (1987), the structure of the Western Cordillera consists of a stack of nappes deformed to an antiform.

First obduction over the Central Cordillera took place during Late Jurassic or Early Cretaceous (Fig. 8), as suggested by the Jambaló Glaucophane Schist (Feininger, 1982) and other metaophiolite complexes with westward dipping foliation along the Romeral Fault (Bourgois *et al.*, 1987). A second phase of obduction can be interpreted from the Yurumal Complex that consists of serpentinite, peridotite, gabbro, massive tholeiitic basalt flows, pillow basalt associated with chert, tuff, and volcaniclastic turbidite beds toward the top (Restrepo and Toussaint, 1977). The complex is intruded by the Antioquia Batholith (80 - 60 Ma). During this obduction phase, the Atrato fore-arc basin was formed, and high subsidence rates provided the favourable conditions for marine deposition throughout Pliocene. Oblique Eocene collision of the Acandí-Mandé-Gorgona Terrane enhanced subsidence rates in fore-arc basins and triggered strain partitioning causing reactivation of the Romeral and Cauca-Patia paleo-sutures as strike-slip faults.

The trailing edge of the Western Cordillera, represented by Late Cretaceous komatiitic rocks of the Gorgona Terrane (Echeverría, 1980; Etayo-Serna and Barrero, 1983). These mafic rocks were interpreted to represent an oceanic plateau formed by extensive decompression melting of an uprising deep mantle plume with similar REE content to rocks in Curaçao (Spadea and Espinosa, 1996; Kerr *et al.*, 1996). Farther to the W, the Late Cretaceous arc sequence of the Baudó Range docked during the late Miocene.

Cenozoic

The Calima Orogeny of Late Cretaceous to Paleocene times (Barrero, 1979) was heralded by uplift of the Central Cordillera. This resulted in the deposition of a molasse wedge, interrupted by local basement uplifts such as that of the Santander Massif in the Eastern Cordillera. Deformation was coeval with the formation of nappes, the extrusion of Cretaceous volcanic rocks in the Western Cordillera, and the emplacement of the Antioquia Batholith in the Central Cordillera. Whereas Tertiary sedimentation in the Eastern Cordillera was dominated by deposition of fluvio-deltaic clastic deposits, sedimenation in the Western Cordillera was marine, and included Miocene deep-water facies in the Atrato Basin. The Central Cordillera, on other hand, was uplifted and affected by pervasive plutonism and subaerial volcanism throughout the Tertiary.

Oblique convergence between the Farallón and South American plates during the Eocene resulted in strain partitioning into E-W compression in the Magdalena Valley and pull-apart basin formation along the Cauca/Patia-Romeral Fault System. Upper Eocene to lower Miocene fluvial sandstone and shale beds were deposited in these basins. N-S strike-slip faulting related to the Romeral/Palestina Fault System developed during the rise of the Western Cordillera. Docking of the Acandí-Mandé-Gorgona Terrane caused a trench jump, starting a new arc-trench system and renewed transpressional deformation and uplift in the Eastern Cordillera as occurs in the Magdalena Valley (Butler and Schamel, 1988). This is interpreted from mesoscopic structures cut by 38 - 35 Ma hydrothermal veins (Cheilletz *et al.*, 1994; Branquet *et al.*, 1996) as well as the creation of independent foredeeps in the Magdalena Valley and the Sabana de Bogotá basins and an increase of Eocene subsidence in the fore-arc Atrato Basin (Incaic Orogeny). Moreover, from late Eocene to Miocene, there is a well-documented eastward migration of the volcanic arc (Toussaint and Restrepo, 1982).

During the Oligocene there occurred extensional collapse of the Northern Central Cordillera with concomitant formation of the Plato and San Jorge basins. A low rate of convergence (Pardo-Casas and Molnar, 1987) and a decrease in subduction angle of the Nazca Plate may explain the lack of magmatic activity, and relative tectonic quiescence during the Oligocene. A major marine transgression in the Eastern Cordillera foredeep is manifest in widespread deposition of shale beds.

During the Miocene, a time of maximum paroxysm and mountain building in Colombia, there occurred renewed folding and thrusting in the Eastern Cordillera (Fig. 11a), with the deposition of a thick molasse sequence in the present-day foredeep (Dengo and Covey, 1993) and by an increase in the subsidence in the Middle and Upper Magdalena basins. Strain partitioning during oblique convergence resulted in significant compressional deformation and reactivation of pre-existing dip-slip faults as strike-slip faults in the Magdalena Valley. Eastern Cordillera shortening was coeval with emplacement of the Baudó Range, development of orogenic parallel strike-slip faults and sinistral strike-slip displacement along the Santa Marta Bucaramanga Fault (Campbell, 1968). Miocene plutonic and volcanic activity varying from alkaline to calc-alkaline in composition was widespread in the Western Cordillera, in the upper reaches of the Magdalena River Valley, and along the axis of the Central Cordillera (Alvarez, 1983; McCourt *et al.*, 1984). Thus, the fluvial conglomerate and sandstone of the Combia Formation (Restrepo *et al.*, 1981), deposited in an extensional basin trending ENE-E, contain andesitic lava flows and pyroclastic beds with island-arc tholeiitic affinities at the top of the section.

Plio-Pleistocene calc-alkaline volcanism continued in the Central Cordillera and along the Cauca/Patia and Romeral faults. Segmentation of the arc-trench system explains the lack of volcanism in the Bucaramanga region, N of the Atrato-San Juan lineament, where earthquake focal mechanisms suggest a shallow subduction angle (Pennington, 1981). Upper mantle Pliocene to Pleistocene alkali basaltic and ultramafic magmatism has been observed in the uppermost regions of the Magdalena Valley and may be attributed to subduction-related deep crustal fractures (Kroonenberg *et al.*, 1982). This volcanism is coeval with rhyolite ignimbrites found in the Central Cordillera.

Crustal growth and tectonic development of the Colombian Andes

Based on Nd model ages, early crustal growth processes in the Colombian Andes began during the Parguasan (1.45 Ga) tectonomagmatic event (Restrepo-Pace *et al.*, 1997). Similar ages have been reported in the Guiana Shield (Priem *et al.*, 1982, 1989) suggesting that these basement rocks extend under the Colombian Andes. A second tectonic metamorphic event occurred at approximately 1.1 Ga involving formation of a granulite belt around the nucleus of the Guyana Schield in the Paleo to Mesoproterozoic times during the collisional Grenville Orogeny (Kroonenberg, 1982; Priem *et al.*, 1982, 1989; Restrepo-Pace *et al.*, 1997).

Cambro-Ordovician siliciclastic rocks were deposited on a passive margin, devoid of volcanic activity. A Late Ordovician Cordilleran-type margin is interpreted from the

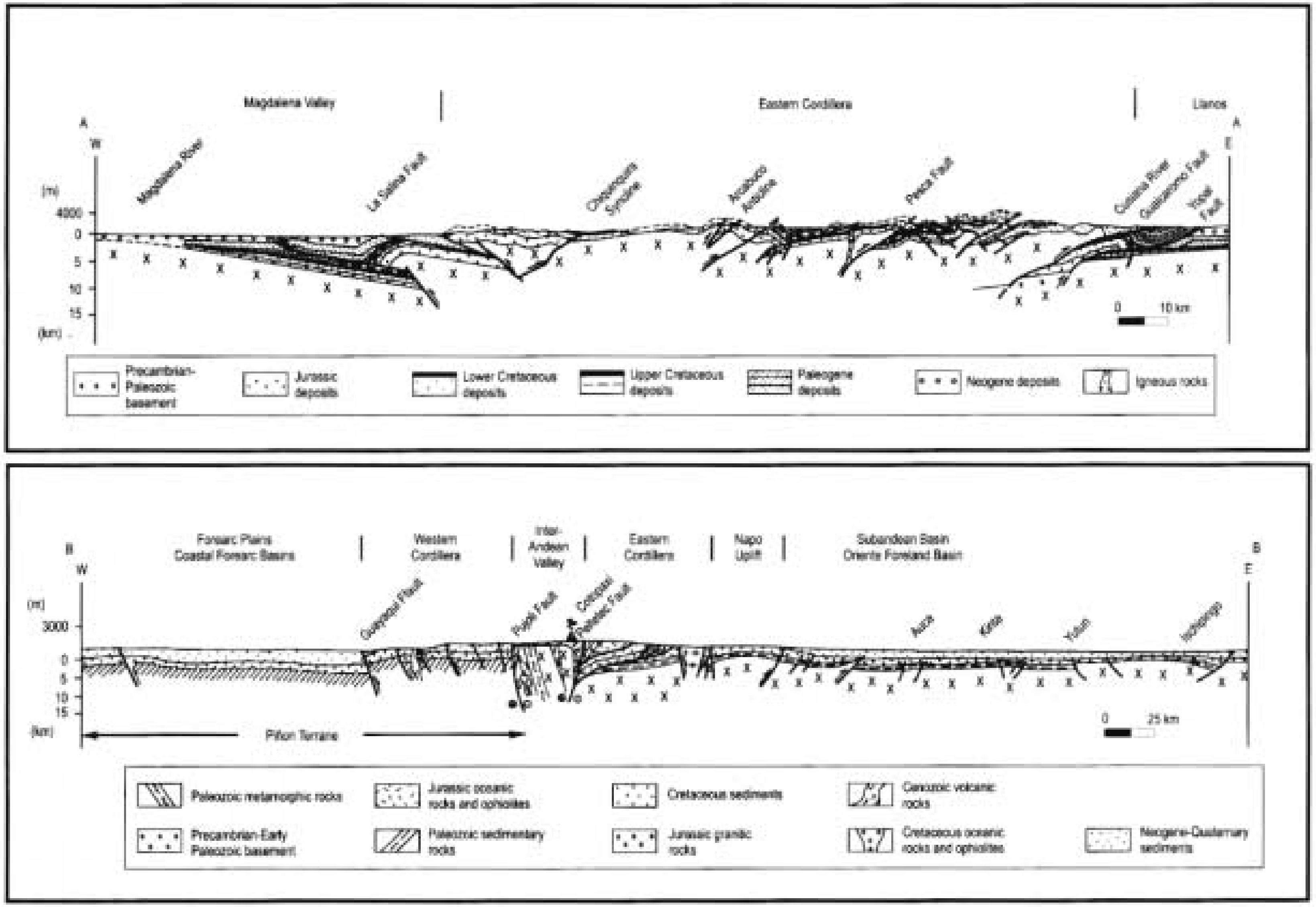


FIGURE 11 - Regional cross-sections: a) Eastern Cordillera of Colombia (based on Colletta et al., 1990). b) Ecuadorian Andes. Location in figs. 8 and 12.

presence of gneiss, schist, migmatite and granite (Santander Plutonic Group) of the Central and Eastern Cordillera (Trumpy, 1943; Harrington and Kay, 1951; Boinet *et al.*, 1985) and may correlate with the Ocloyic Orogenic event of northern Argentina (Ramos, 1988). The beginning of Gondwana assemblage and the closing of the ocean between Gondwana and Laurasia is suggested by Lower Paleozoic plutons in the Mérida Andes, the Serrania de Perijá, and the Santander, Floresta and Quetame massifs in the Eastern Cordillera (Boinet *et al.*, 1985). This event was followed by *deposition of siliciclastic sediments and subordinate* carbonate rocks containing a Devonian flora and fauna with affinity to the Eastern Americas Realm, suggesting the absence of barriers between Euramerica and Gondwana during the Middle Devonian (Berry *et al.*, 1997). Carboniferous deposition took place in a foredeep filled with thick siliciclastic sediments that are capped by Permian shallow marine carbonates. A second collisional event in northern South America is suggested by widespread S-type granite magmatism during Late Permian to Early Triassic. This collision took place along the Palestina Fault where McCourt and Feininger (1984) suggested the presence of a paleo-suture.

Right-lateral Triassic shear of the Northern Andes away from North America accompanied by crustal-scale transpression has been speculated for the northern Andes (Jaillard *et al.*, 1990; Aspden *et al.*, 1992). However, the small, foliated, calc-alkaline plutonic precursors of regional Late Triassic extension in the Magdalena Valley could be related to ensialic back-arc extension similar to those proposed for the Choyoi Group in Central Argentina (Ramos and Kay, 1990). In fact, this event was followed by ensialic Jurassic back-arc extension associated with high heat flow and volcanism, which thinned the crust and formed a series of N-NW oriented grabens. These new Mesozoic basins were the depositional loci of volcanic and sedimentary rocks, that later were inverted during the Andean Orogeny. Calc-alkaline *volcanism migrated seaward during trench rollback as* interpreted by the position of the Jurassic Ibague and Segovia batholiths, relative to the position to the Late Cretaceous Antioquia Batholith.

The most important crustal growth process was the Late Cretaceous accretion of the Central and Western Cordilleras during oblique convergence of the Farallon Plate (Pardo-Casas and Molnar, 1987). This event involved intensive magmatism, eastward arc migration, *regional metamorphism, and* reactivation of paleo-sutures such as the Romeral and Palestina faults (Feininger, 1970; McCourt and Feininger, 1984; *Restrepo and Toussaint, 1988). This tectonic event also* provided the mechanism for separation of the upper part of the Magdalena Valley from the Putumayo Basin.

The middle Eocene westward shift of the loci of volcanic activity was followed by eastward arc migration (Toussaint, 1978) associated with a significant increase in convergence rates (Pardo-Casas and Molnar, 1987). An arc-trench system, developed after emplacement of the Late Cretaceous *Gorgona Komatiites, encompassed pervasive calc-alkaline* to tholeiitic magmatism and marine deposition in the Atrato fore-arc basin. Middle Eocene deformation was recorded by compressional deformation and molasse deposition. Equally important was the dual Middle Magdalena and Sabana de Bogotá foredeeps where synchronous, thick molasse was deposited. Compressional deformation and uplift of the Sierra de Perijá was initiated as long ago as the early Eocene (Kellogg, 1984). At that time, the Cauca Patia and Romeral sutures were reactivated as strike-slip faults with the formation of pull-apart basins, filled with terrestrial deposits along this inter Andean depression.

Although convergence rates decreased significantly during Oligocene (Pardo-Casas and Molnar, 1987), transpressive deformation continued in the Eastern Cordillera *with formation of E-verging thrusts. Cessation of igneous* activity in the Colombian Andes may be explained by decreased subduction angle. Eastward thrust propagation and foredeep migration were hampered by Precambrian and Paleozoic rocks in the Eastern Cordillera, which behaved as a backstop along the paleo-Guaicáramo Fault. The Sierra de Perijá also underwent compressional deformation and uplift (Kellogg, 1984), and the Caño Limón strike-slip fault, in the present-day Llanos Basin, was active. Crustal thickening and uplift of the Central Cordillera was related to igneous underplating during the process of collision/subduction. This uplift accompanied subsidence in pull-apart basins along the Cauca/Patia and Romeral faults as well as in the Magdalena Valley intermontane basins.

Break-up of the Farallon Plate during the early Miocene was associated with a change in the angle of convergence from oblique to orthogonal, with a concomitant increase in rate of plate convergence (Handschumacher, 1976; Pardo-Casas and Molnar, 1987). This event was accompanied by accretion of the Baudó Range, deformation of the pull-apart basins along the Cauca Valley, the beginning of new magmatic cycle, and diachronous tectonic uplift of the three cordilleras. The Panamá-Baudó Arc that dips to the S and W was accreted along a major strike slip fault-suture that later was intruded by ultramafic rocks and komatiitic basalt (20 Ma) of subcrustal origin (Salinas and Tistl, 1991). Shortening caused the inversion of Mesozoic half grabens *characterized by high-angle, basement-involved reverse* faults, folding, thrusting and uplifting of the Eastern Cordillera (Colletta *et al.*, 1990; Dengo and Covey, 1993).

Widespread deformation of the Eastern Cordillera during the middle Miocene was displayed by large antiforms and pop-up structures formed by opposite verging thrust faults, short and large wavelength and folds detached in Cretaceous rocks and complex imbricate thrusts (Butler and *Schamel, 1988; Colletta et al., 1990; Dengo and Covey, 1993).* The foldbelt asymmetry was related to the presence of two deformation fronts and two foreland provinces (Colletta *et al., 1990). Indeed, loading of thrust sheets increased* subsidence in the Magdalena and the Llanos basins. Northwards, the Sierra de Perijá underwent a third period of compressional deformation and inversion (Kellogg, 1984). About 125 km of left-lateral displacement of the Santa Marta strike-slip fault in the northern part of the Central Cordillera (Campbell, 1968) was perhaps related to the eastward motion of the Caribbean Plate.

Compressional deformation continued in the Colombian Andes throughout Plio-Pleistocene. It was accompanied by calc-alkaline volcanism along the Cauca Patia and Romeral faults; rhyolite volcanism in the Central Cordillera; and alkali basaltic and ultramafic magmatism in the uppermost part of the Magdalena Valley (Kroonenberg, 1982). Whereas compressional processes continued uplifting the Eastern Cordillera, igneous underplating enhanced uplift in the Western and Central Cordilleras. Equally important was the late Miocene-Pleistocene uplift of the Garzón Massif (Van der Wiel, 1991). Strain partitioning caused reverse faults reactivation as strike-slip faults in the Magdalena Valley, separation and individualization of the Neiva Basin, and formation of orogen-parallel strike-slip faults such as the Apiáy Fault in the Llanos Basin. Eastward propagation of compressional deformation of the Eastern Cordillera Foldbelt, across the Precambrian to Paleozoic backstop, took place by out of sequence movement of the Guaicáramo Fault, which splayed into the Yopal and San Juan de Ariporo frontal thrusts. The last two faults are bedding-parallel within the Eocene molasse, and this accounts for significant disharmonic folding and blind thrusting in the frontal thrust.

Ecuadorian Andes

Unlike the Colombian Andes, the evolution of the Ecuadorian Andes can be described in terms of only the Western Cordillera and Eastern Cordillera (Cordillera Real), each with an unique rock assemblage (Fig. 12). The Central and Eastern Cordillera of Colombia merge near the Ecuadorian border. A well-defined inter-Andean graben, formed by strike-slip processes, separates these cordilleras, and is bounded by the Pujili and Peltetec faults (Campbell, 1974; Baldock and Longo, 1982).

The Western Cordillera consists of a strongly deformed Cretaceous allochthonous terrane capped by Tertiary continental arc rocks and a Paleogene marine clastic and carbonate sequence intruded by Tertiary plutons. The eastern boundary is sharp and marked by the Pujili Fault, whereas the western flank consists of accreted Mesozoic back-arc rocks overlain by Tertiary fore-arc deposits. The Cordillera Real consists of highly metamorphosed, thrusted and folded Paleozoic to Mesozoic rocks, partially covered by Tertiary intrusive and volcanic rocks. The Peltetec Fault and the Oriente Basin mark the western and eastern boundaries, respectively. The Inter-Andean graben, between the two Cordilleras, is a tectonic depression approximately 50 km wide, which formed during late Miocene transtension, and filled with non-marine clastic deposits. This depression is the loci of large number of volcanoes located along major bounding faults (Litherland and Aspden, 1992).

Some of the first contributions to the geology of the Ecuadorian Andes include those by Sauer (1971) who summarized his fieldwork. Other pioneers include Colony and Sinclair (1928), Sheppard (1937), Barrington (1938), Tschopp (1953), and Marchant (1961). Faucher and Savoyat (1973) updated Sauer's classical paper in their review of Ecuador paleogeography and orogenic events. Later a posthumous publication by Kennerley (1980) provided a significant advance in the understanding of the geological chronology of Ecuador. This was followed by the explanatory paper by Baldock and Longo (1982) of the geological map of Ecuador. Work by the British Geological Survey has been incorporated in a new geological map of Ecuador (Litherland *et al.*, 1994). This work includes a large number of new isotope ages, petrographic and geochemical analysis, and documentation of several terranes, which provides a more coherent picture of the tectonic evolution of the Ecuadorian Andes.

Tectono-stratigraphic evolution

Precambrian

There are no Precambrian rocks cropping out in the Ecuadorian Andes. However, xenocrystic zircon from analyzed rocks suggest that mafic and granitic magmas were contaminated during crustal ascent, or were partially derived from Guiana Shield crustal recycling (Noble *et al.*, 1997). Thus, inherited zircon ages from the Marcabeli Pluton (540 Ma, 2.22 Ga, and 2.876 Ga) as well as zircon upper intercept ages from Tres Lagunas granitoids (2.9 to 2.85 Ga) in the southern Cordillera Real are interpreted to represent detrital input from metasedimentary precursors. Furthermore, Nd-depleted mantle model ages (T_{DM}) for the Tres Lagunas and Marcabeli granitoids support 1.6 - 1.4 Ga ages for the protoliths (Noble *et al.*, 1997).

Similar Sm/Nd ratios and T_{DM} ages have been calculated by Restrepo-Pace *et al.* (1997) for Orinoquian granulite in the Colombian Eastern Cordillera. Priem *et al.* (1989) have reported similar Rb/Sr whole rock ages for the Parguazan tectonomagmatic episode for the Mitu Migmatitic Complex in the Guiana Shield. Although no granulites have been reported, isotope ages seem to suggest the presence of the Guiana Shield under the Ecuadorian Andes.

Paleozoic

Rocks of probable Lower Paleozoic age were overprinted by pervasive anatexis during the emplacement of Triassic S-type granite and extensive Jurassic cordilleran I-type granitoid (Aspden *et al.*, 1992). The rocks include quartzite, metasiltstone, graphitic schist, black phyllite and rare metagreywacke of the Chiguinda Unit which dominates the Southern Cordillera Real (Litherland *et al.*, 1994). Bedding and primary cleavage of this unit are subparallel and steeply dipping, and follow the Andean trend with evidence for more than one phase of deformation (Litherland *et al.*, 1994). Indeed, Kennerley (1980) correlated this unit with low-grade metamorphic phyllite and slate beds of the Salas Formation (Ordovician-Silurian) in the Olmos Massif in Peru (Mourier, 1988). To the NE, near Baños, there are also some 417 Ma pelitic schist beds of the Agoyan Unit, which have been correlated with the Chiguinda Unit by Litherland *et al.* (1994) who also reported reworked Early Ordovician acritarchs in the Jurassic Maguazo tectonic unit.

Upper Paleozoic rocks are present in the Cutucú High and in the Tahuin/Amotape Terrane. The oldest Paleozoic unit, the Devonian Pumbuiza Formation, consists of highly folded, grey to black slate with fine-grained quartz sandstone that form the core of the Cutucú High. Similar rocks have been reported to the W in the Tahuin/ Amotape Terrane (Martinez, 1970). This unit is unconformably overlain by the Macuma Formation (Lower Pennsylvanian) that consists of interbedded bioclastic limestone and black shale beds. These two units may represent shallow-water

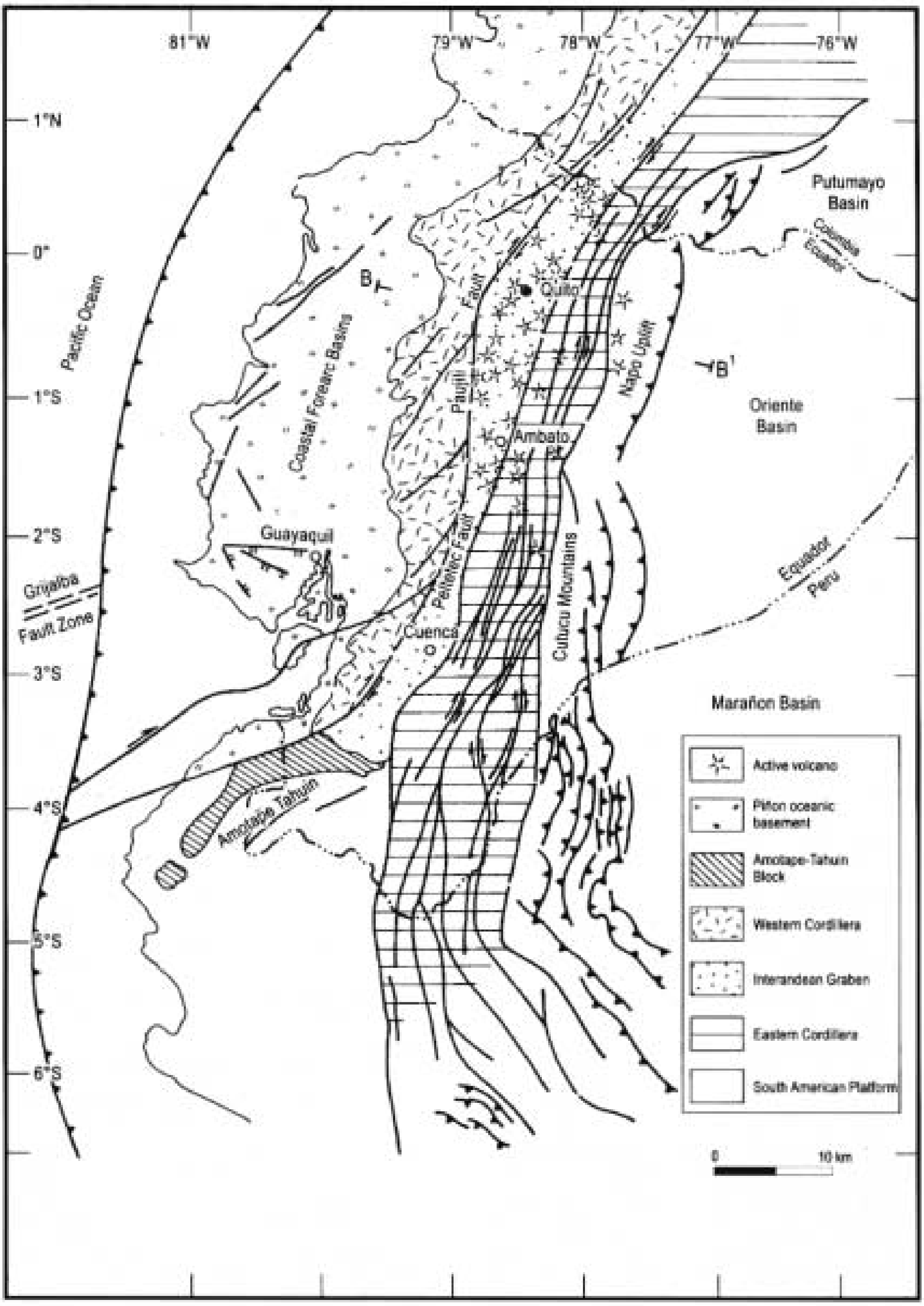


Fig. 12 - Major geological provinces of the Ecuadorian Andes with active volcanoes (based on Litherland et al., 1994).

facies equivalent to the thick Devonian to Pennsylvanian quartz-rich arkose and feldspathic wackes interbedded with shale, interpreted as a flysch sequence in the Tahuin/Amotape Block (Martinez, 1970; Mourier, 1988).

Litherland *et al.* (1994) described within the Loja Terrane, in the southern Cordillera Real, the Isimachi Unit of possible Upper Paleozoic age consisting of low-grade phyllite and marble. They also described the Monte Olivo amphibolite, which include an amphibolite dyke with relict igneous texture of Upper Devonian age (363 ± 9 Ma and 371 ± 10 Ma), and a garnet amphibolite showing Carboniferous ages (306 ± 10 Ma and 342 ± 23 Ma). These amphibolites are associated with the Chiguinda and Agoyan units of the Loja Terrane.

Mesozoic

Litherland *et al.* (1994) described the Triassic Piuntza unit as a continental/marine volcaniclastic sequence, between the Macuma and Santiago formations. This is equivalent to the Upper Triassic high-grade migmatite of the Sabanilla Complex, which is part of the Tres Lagunas S-type granite suite (227.6 ± 3.2 Ma) in the Cordillera Real, and the Moromoro Granite (220 ± 6 Ma) in the El Oro Metamorphic Belt (Litherland *et al.*, 1994). The Tres Lagunas blue quartz granite has been traced laterally into gneissic belts related to steep westward-dipping Andean-trending shear zones with widely developed S-C mylonites. Such shear zones suggest at least several phases of dextral transpression (Aspden *et al.*, 1992; Litherland *et al.*, 1994). To the S the Sabanilla Orthogneiss, consisting of foliated biotite-muscovite garnet granite, contains metasedimentary xenoliths in various stages of digestion with $^{87}Rb/^{86}Sr$ age of 224 ± 37 Ma (Litherland *et al.*, 1994). These S-type granites are interpreted to have formed during the separation of North and South America during the Triassic. However, they may represent a Late Triassic delayed phase collision between Gondwana and Laurasia followed by right-lateral shear. The pegmatitic and saussuritized Piedras Amphibolite (221 ± 16 Ma), originally a low-K basalt, is present within the El Oro Metamorphic Complex, and has been interpreted as a slice of metamorphosed oceanic crust (Litherland *et al.*, 1994).

Jurassic deposition started to the E of the present Eastern Cordillera with deposition of dark grey limestone units, calcareous sandstone and shale beds of the Santiago Formation (Lower Jurassic) (Geyer, 1974). This unit is overlain by more than 2500 m of varicolored shale beds and sandstone beds with thin zones of dolomite and gypsum of the Chapiza Formation. Basaltic to rhyolitic lava flows, interbedded with dacitic to rhyolitic ignimbrites and unwelded lapilli tuff of the Misahualli Formation (Middle Jurassic) cap this extensional cycle on the eastern flank of the Eastern Cordillera (Tschopp, 1953; Romeuf *et al.*, 1995). REE and multi-element scans of these calc-alkaline rocks suggest an arc-trench system (Romeuf *et al.*, 1995) and perhaps deposition in an ensialic back-arc setting. Thus, the Misahualli Formation (172.3 ± 2.1 Ma) is cut by the Abitagua Granite (162 ± 1 Ma). To the S these volcanic rocks are equivalent to the Zamora Batholith (171 ± 6 Ma and 152 to 180 Ma) as documented by Litherland *et al.* (1994). Furthermore, low initial Sr isotope ratios and major element analysis of the Late Jurassic Abitagua and Zamora granitoids suggest cordilleran I-type plutons representing the southern extension of the Colombia Magmatic Belt (Aspden *et al.*, 1992).

In the northern part of the Cordillera Real, there is a continuous and narrow belt (15 km wide) of the Jurassic Upano Unit. According to Litherland *et al.* (1994), this belt consists of andesitic greenstone, greenschist and metagreywacke intercalated with pelitic and graphitic schist with steep tectonic foliation. To the W, they also described grey to black, graphitic-muscovite schist of the Cayuja Unit in tectonic contact with metamorphosed black limestone beds and black calcareous phyllite, with some marble and calc-silicate rocks, of the Cerro Hermozo Unit. These units may represent metamorphosed equivalents of the Santiago, Chapiza and Misahualli formations. To the S, along the Cordillera Real, there occurs a sequence of agglomerate beds and green phyllite of volcaniclastic origin of the Jurassic Alao-Paute Unit (Litherland *et al.*, 1994). These authors also describe in fault contact, the Jurassic (Middle Callovian to Middle Oxfordian) Manguazo Unit, a slightly metamorphosed turbidite sequence interbedded with andesitic basalt, black phyllite and chert, interpreted as a marine fore-arc sequence.

A narrow zone of highly deformed, steeply-dipping ophiolitic rocks along the Peltetec Fault, on the western slope of the Cordillera Real, is interpreted as a mélange with island arc signatures and oceanic crust structure (Litherland *et al.*, 1994). The sheared metagabbro, metabasalt, foliated tremolitic serpentinite, spilitized dolerite, and volcaniclastic mélange are interpreted as a Jurassic subduction zone or suture similar to those described along the Romeral Fault in Colombia (McCourt *et al.*, 1984; Restrepo and Toussaint, 1988). Noteworthy, is the Raspas Formation in the El Oro Ophiolite Complex consisting of garnetiferous pelitic schist and layers of glaucophane schist, quartzite layers, eclogite (132 ± 5 Ma) and eclogite amphibolite (Feininger, 1980). This high pressure/low temperature unit was probably formed during accretion and clockwise rotation of the Tahuin Block.

Cretaceous sedimentation in the Cordillera Real started with deposition of the fluvial to shallow marine, quartz rich sandstone beds of the Hollin Formation. This was followed during the Middle Albian to Early Maastrichtian by deposition of dark grey shale and limestone interbedded with sandstone of the Napo Formation. A regional unconformity separates the Napo Formation from conglomerate, sandstone and shale of the Tena Formation (Maastrichtian/Paleocene). To the W, along the coast, the Piñon Terrane consist of basaltic pillow lava flows, hyaloclastite, dolerite, boninite, and cumulate gabbro of the Piñon Formation (Goossens *et al.*, 1977; Lebrat *et al.*, 1987; Van Thournout *et al.*, 1992). In turn the Piñon Formation is overlain by more than 2000 m of deepwater volcaniclastic sandstone, pyroclastic breccia, basaltic andesite lava flows and subordinate limestone, rich in organic material, of the Cayo Formation (Cenomanian to Maastrichtian). This formation is transitionally overlain by rhythmically-bedded silicified tuff and tuffaceous shale of the Guayaquil Formation (Late Maastrichtian to Early Paleocene).

In the Western Cordillera, the Cretaceous rocks consist of mafic lava flows and marine greywacke of the Toachi and Pilaton units, respectively (Macuchi Formation), and mainly shale, greywacke, tuff and andesitic sills and lava flows of the Yunguilla Formation (Campanian) and the Callo Rumi Formation (Maastrichtian) (Cosma *et al.*, 1998). To the S, in the Celica-Lancones Basin, E of the Tahuin-Amotape Block, there are massive andesitic lava flows, welded tuff beds, pillow breccias and agglomerates of the Celica Formation (Albian). The Celica Formation is overlain by more than 2000 m of coarse-grained deep-water volcaniclastic sediments of the Alamor Formation (Cenomian-Coniacian). Unconformably overlying these beds is a sequence of interbedded greywacke, shale and marl of the Naranjo Formation (Santonian/Campanian) and thin bedded greywackes and nodular limestone with conglomerate lenses of the Casanga Formation (Jaillard *et al.*, 1996).

Cenozoic

Deformation in the Cordillera Real is manifest as thick molasse deposition from Maastrichtian to Miocene of the Tena, Tiyuyacu, Orteguaza/Chalcana, Arajuno and Chambira formations in the Subandean Basin. However, the Western Cordillera and coastal basins contain Eocene to Miocene deep to shallow water siliciclastic sediments and carbonate rocks, deposited in a fore-arc setting, overprinted by strike-slip faults (Bourgois *et al.*, 1990).

The Cordillera Real, the Western Cordillera and the Inter-Andean Graben were sites of widespread, pervasive volcanism during the Tertiary. A higher percentage of calc-alkaline rocks is present in the Central and Southern Andes suggesting a thicker crust (Hörmann and Pichler, 1982). Basalt, andesite and dacite of the Eocene Tandapi Formation (Cosma *et al.*, 1998) are the dominant rocks of the Western Cordillera. Miliolid-rich limestone beds of the Unacota Formation (middle Eocene) and its lateral equivalent deep-water volcaniclastic sediments of the Apagua Formation, overlie this formation.

The middle Eocene is unconformably overlain by Oligocene volcaniclastic sediments and lava flows of the Saraguro Formation (Eguez, 1986; Bourgois *et al.*, 1990). In the Cordillera Real, Lavenu *et al.* (1992) described calc-alkaline volcaniclastic and lava flows of the Sacapalca (Paleocene-Eocene), the Saraguro (Oligocene), the Pisayambo (Mio-Pliocene) and the Cotopaxi (Plio-Quaternary) volcanic events. Post-Oligocene Saraguro, Pisayambo and Cotopaxi volcanic events are also present along the Inter-Andean Graben and consist of volcaniclastic sediments and lava flows.

Today, most active volcanoes are along the Peltetec and Pujili bounding faults. Miocene-Pliocene reactivation of these major faults generated pull-apart basins along the Inter-Andean Graben. These basins are filled by volcaniclastic sediments derived from the Cordillera Real and deposited in fluvial and lacustrine environments from the middle Miocene onward. Late Miocene to Pliocene compressional deformation of the Andean Orogeny, has inverted these basins (Barragán *et al.*, 1996; Hungerbuehler *et al.*, 1996).

Crustal growth, tectonic evolution and structural styles of the Ecuadorian Andes

As indicated by inherited older zircons ages from the Marcabeli, Tres Lagunas Granite and El Oro Complex, crustal growth in the Ecuadorian Andes may have begun during the collisional Precambrian Orinoquian (Grenville) Orogeny. Indeed, Nd-depleted mantle model ages (T_{DM}) from the Tres Lagunas and Marcabeli granitoids indicate a 1.6 - 1.4 Ga age for the protoliths (Litherland *et al.*, 1994; Noble *et al.*, 1997). This collisional belt may be a continuation of the granulite belt in Colombia, where similar Sm/Nd ratios and T_{DM} ages have been calculated (Restrepo-Pace *et al.*, 1997).

Like the Precambrian event, the Ocloyic Orogeny (Late Ordovician) is poorly age-constrained since Lower Paleozoic rocks have undergone pervasive metamorphism and anatexis, often with schistosity following the Andean trend (Fig. 7). However, Kennerley (1980) has correlated the Chiquinda Unit of the Cordillera Real with the low grade Ordovician phyllite and slate from the Olmos Massif. Furthermore, reworked Ordovician acritarchs have been reported in the Jurassic Maguazo tectonic unit (Litherland *et al.*, 1994).

A possible Late Devonian to Lower Carboniferous tectonic event is also suggested by amphibolite dykes (363 ± 9 Ma) and garnet amphibolite (342 ± 23 Ma) associated with the Chiquinda and Agoyan units of the Loja Terrane (Litherland *et al.*, 1994). This event may correlate with an incipient Gondwana-Laurasia assemblage (Berry *et al.*, 1997). Mixed siliciclastic and carbonate sedimentary rocks of Devonian and Permo-Carboniferous age are reported in the Cutucú Mountains and Tahuin/ Amotape Terrane. In the Tahuin/Amotape Terrain these rocks have undergone significant Late Paleozoic folding and thrusting related to final Late Paleozoic Pangea assemblage.

Late Triassic anatectic processes are recorded in the Tres Lagunas and Moromoro S-type granites of the Cordillera Real and El Oro Metamorphic Belt. S-C mylonites in these plutons suggest several episodes of dextral transpression (Litherland *et al.*, 1994) when North America rifted away from South America (Aspden *et al.*, 1992). However, the Piedras amphibolites (221 ± 16 Ma) within the El Oro Metamorphic Complex suggest a collisional event (Feininger, 1980) during the closure of an oceanic basin. Alternatively, S-type granites could represent collisional plutons, and S-C mylonites could have been formed thereafter by transpressional processes. This tectonic event has been named as the Moromoro Orogenic Event (Litherland *et al.*, 1994).

An arc-trench was established during the Middle Jurassic as interpreted from the presence of cordilleran I-type Zamora (171 ± 6 Ma and 180 to 152 Ma) and Abitagua (162 ± 1 Ma) granites (Litherland *et al.*, 1994). Such granites were coeval with ensialic back-arc extension and thick red-bed deposition interbedded with bimodal volcanics. The subduction zone may have been situated along the Peltetec Fault where inliers of sheared metagabbro, metabasalt, foliated tremolitic serpentinite, spilitized dolerite and volcaniclastic rocks occur. The assemblage may represent an accretionary prism. Subduction continued in the S in the El Oro Ophiolite Complex, which contains foliated eclogite

(132 ± 5 Ma) described by Feininger (1980). Indeed, this deformation may be related to a 110° clockwise rotation of the Amotape/Tahuin Block, followed by dextral shearing prior to the Neocomian (Mourier *et al.*, 1988).

Perhaps the most important tectonic event in the Ecuadorian Andes evolution was related to the diachronous emplacement of mafic to ultramafic oceanic back-arc and intra-oceanic arc sequences of the Piñon Terrane (Goossens *et al.*, 1977; Feininger and Bristow, 1980; Lebrat *et al.*, 1987; Cosma *et al.*, 1998). This event has been identified as the Peruvian Phase of the Andean Orogeny (Peltetec Orogeny). Paleomagnetic data suggest large-scale (70°) clockwise rotation of the Piñon Terrane (Roperch *et al.*, 1987). Regionally, docking of this allochthonous terrane was accompanied by Late Cretaceous uplift and erosion of the proto-Cordillera Real as can be seen by the presence of the Tena Formation molasse of Late Maastrichtian age, and by inversion of half grabens in the Oriente Basin (Marksteiner and Alemán, 1997). This collision, manifest by major shearing and thrusting of Mesozoic rocks in the Eastern Cordillera, was accompanied by Barrovian-type upper amphibolite to greenschist facies metamorphism and the resetting of isotopic ages (Litherland *et al.*, 1994). The main Late Cretaceous suture was located along the Pujili Fault where an antiformal duplex of allochthonous rocks has been emplaced in the Western Cordillera. Basement involved deformation began and continued with different intensities throughout the Plio-Pleistocene. However, W of the Cosanga Fault, a stack of nappes involving crystalline basement was emplaced. To the S, in the Celica-Lancones Basin, this orogenic event is represented by an abrupt change in provenance from a dominantly volcanic provenance of the Alamor and El Naranjo formations to basement-derived conglomerate of the Cosanga Formation.

Termination of Piñon Terrane accretion was accompanied by westward trench migration, increased convergence rates (Pardo-Casas and Molnar, 1987) and the emplacement of the calc-alkaline Tandapi Arc (Eguez, 1986; Cosma *et al.*, 1998) during the Incaic Phase of the Andean Orogeny. This event was coeval with deposition of miliolid-rich limestone units around the arc and deepwater volcaniclastic beds in the Western Cordillera. The Cordillera Real was uplifted and the Oriente Basin was the site of Eocene Tiyuyacu molasse deposition. Inversion of half grabens, compressional deformation and uplift of the Napo and Cutucú Mountains continued (Marksteiner and Alemán, 1997). Excellent evidence for this event is found to the W of Macas, where 42 Ma aplite dykes cut mesoscopic folds in the Hollin Formation. To the W, along the coastal basins, the deepwater quartz-rich sandstone of the Ancon Group, containing abundant Piñon Terrane olistoliths record this orogenic event.

The Oligocene was a time of relatively tectonic quiescence and low convergence rate described as the Aymara Phase of the Andean Orogeny (Pardo-Casas and Molnar, 1987; Soler and Bonhomme, 1990). The Oligocene Saraguro volcanic arc sequence suggests continuous eastward arc migration (Lavenu *et al.*, 1992). This volcaniclastic sequence was deposited along the Western Cordillera and the western flank of the Cordillera Real.

The Andean Orogeny started as a consequence of the break-up of the Farallon Plate about 25 Ma (Handschumacher, 1976). Oblique compression and relatively high rates of plate convergence (Pardo-Casas and Molnar, 1987) account for early Miocene reactivation of the Peltetec and Pujili faults and the development of pull-apart basins along the Inter-Andean Graben (Litherland and Aspden, 1992). These basins filled with fluvial and lacustrine deposits derived from the East, suggesting that the Western Cordillera did not reach its present elevation until the Pliocene (Hungerbuehler *et al.*, 1996). Subduction of the Grijalva Fracture Zone (Lonsdale, 1978) coincides with the northern limit of low angle subduction and volcanic gap that characterized the Ecuadorian Andes segmentation (Hall and Wood, 1985; Gutscher *et al.*, 1999). Indeed, Miocene volcanic activity was restricted to the southern and central Ecuadorian Andes (Lavenu *et al.*, 1992).

Late Miocene/Pliocene uplift and inversion of transtensional basins along the Inter-Andean Graben was marked by large scale thrusting, inverse faulting and folding as determined by apatite fission track dating (Hungerbuehler *et al.*, 1996). This deformation was accompanied by continuous volcanic activity located along the present volcanic front from Cuenca in the S to central Colombia in the N (Barberi *et al.*, 1988). Additional fission track studies along the Abitagua Batholith suggest transpressional uplift in the last 5 Ma. During this time, the volcanic axis shifted eastwards to the Cordillera Real. Since the early Quaternary the volcanic belt has widened to the Western Cordillera and along the length of the Inter-Andean Graben (Fig. 11b) where a large number of volcanoes are present (Barberi *et al.*, 1988). Pliocene to Quaternary volcanoes along this depression contain andesites with a clear calc-alkaline affinity and lower K_2O content than those of the Cordillera Real (Barberi *et al.*, 1988). This may be related to deep crustal fractures controlling the emplacement of volcanoes along the Inter-Andean Graben. Large volumes of Holocene lahars formed during dome collapse, which melted part of the volcano's icecap and transformed rapidly into a debris flow. These lahar deposits have been documented by Mothes *et al.* (1998), in the Cotopaxi Volcano (4700 years BP) and by Samaniego *et al.* (1998) in the Cayambe Volcano (4000 - 650 years BP).

REFERENCES

Alvarez, J. (1981). Determinación de la edad Rb-Sr en rocas del Macizo de Garzón, Cordillera Oriental, Colombia. *Geología Norandina*, **4**, 31-38.

Alvarez, J. (1983). Geología de la Cordillera Central y Occidente colombiano y petroquímica de los intrusivos granitoides mesocenozoicos. *Boletín Geológico*, Bogotá, **26**, 1-175.

Alvarez, J. and Cordani, U.G. (1980). Precambrian basement within the septentrional Andes: age and geological evolution. *Abstracts, XXVI International Geological Congress, 1, 10, Paris.*

Arnold, H.C. (1966). Upper Paleozoic Sabaneta-Palmarito Sequence of Merida Andes, Venezuela. *American Association of Petroleum Geologists Bulletin*, **50**(11), 2366-2387.

Aspden, J.A., Fortey, N., Litherland, M., Viteri, F., and Harrison, S.M. (1992). Regional S-type granites in the Ecuadorian Andes: Possible remnants of the breakup of western Gondwana. *Journal of South American Earth Science*, **6**, 123-132.

Aubouin, J. (1973). Présentation de la Cordillère des Andes. *Revue de Géographie Physique et de Géologie Dynamique*, X, 5-10.

Audemard, F. (1991). *Tectonics of Western Venezuela*. Ph.D. Thesis, Rice University, 245 p.

Baldock, J. W., and Longo, R. (1982). *Geology of Ecuador (Explanation of the 1:1,000,000 Geological Map), Quito. Dirección General de Geología y Minas*, Ministerio de Recursos Naturales y Energéticos, 70 p.

Barberi, F., Coltelli, M., Ferrara, G., Inocenti, F., Navarro, J. M., and Santacroce, R. (1988). Plio-Quaternary volcanism in Ecuador. *Geological Magazine*, **125**, 1-14.

Barrett, S. (1988). The Devonian system in Colombia. In: *Devonian of the World*, eds. McMillan, N.J., Embry, A.F. and Glass, D.J. *Canadian Society of Petroleum. Geologists Memoir*, **14**(I), 705-717.

Barragán, R., Baudino, R., and Marocco, R. (1996). Geodynamic evolution of the Neogene intermontane Chota basin, Northern Andes of Ecuador. *Journal of South American Earth Sciences*, **9**, 309-319.

Barrero, D. (1979). Geology of the Western Cordillera, west of Buga and Roldanillo, Colombia. *Publicaciones Geológicas Especiales del Ingeominas*, 4, 1-75.

Barrington, B.C. (1938). On a theory of gravitational sliding applied to the Tertiary of Ancon, Ecuador. *Journal of the Geological Society, London*, **81**, 454-460.

Bartok, P. (1993). Prebreakup geology of the Gulf of Mexico-Caribbean: Its relation to Triassic and Jurassic rift systems of the region. *Tectonics*, **12**, 441-459.

Bartok, P., Reijers, T.J.A., and Juhasz, I. (1981). Lower Cretaceous Cogollo Group, Maracaibo Basin, Venezuela: Sedimentology, diagenesis and petrophysics. *American Association of Petroleum Geologists Bulletin*, **65**, 1110-1134.

Bellizzia, A. and Pimentel, N. (1994). Terreno Mérida: un cinturón alóctono Herciniano en la Cordillera de Los Andes de Venezuela. *V Simposio Bolivariano Exploración Petrolera en las Cuencas Subandinas, Memoria*, 271-299.

Bellizzia, A. and Rodriguez, G. D. (1968). Consideraciones sobre la estratigrafía de los Estados de Lara, Yaracuy, Cojedes y Carabobo, Venezuela. *Boletín de Geología, Caracas*, **9**(18), 515-563.

Beltrán, N. (1994). Trazas activas y síntesis neotectónica de Venezuela a Escala 1:2.000.000. *7° Congreso Venezolano de Geofísica*, 541-547.

Benedetto, J.L. and Odreman, O.E. (1977). Bioestratigrafia y paleoecología de las unidades permocarbónicas aflorantes en el área de Carache-Agua de Obispo, Estado Trujillo, Venezuela. *5° Congreso Geológico Venezolano*, 253-288.

Benedetto, J.L. and Ramirez Puig E. (1982). La secuencia sedimentaria Precambrico-Paleozoico Inferior pericratónica del extremo norte de Sudamérica y sus relaciones con las cuencas del Norte de Africa. *V° Congreso Latinoamericano de Geología*, **2**, 411-425.

Benedetto, J.L., Sánchez, T.M., and Brussa, E.D. (1992). Las Cuencas Silúricas de América Latina. In: *Paleozoico Inferior de Ibero-América*, eds. Gutiérrez Marcos, J.C., Saavedra, J. and Rabano, I. Universidad de Estremadura, 119-148.

Berry, C., Casas, J., Moody, J. M., and Young, G. (1997). Síntesis estratigráfica y paleontológica de la Formación Campo Chico (Devónico Medio-superior), Sierra de Perijá. *7° Congreso Geológico Venezolano*, **I**, 125-132.

Bosch, M. and Rodriguez, I. (1992). North Venezuelan collisional crustal block: The boundary between the Caribbean and South American plates. *Journal of South American Earth Sciences*, **6**(3), 133-143.

Boinet, T., Bourgois, J., Bellon, H. and Toussaint, J-F. (1985). Age et répartition du magmatisme Prémèsozoïque des Andes de Colombie. *Comptes Rendus de l'Académie des Sciences de Paris*, **300**, Serie II (10), 445-450.

Boinet, T., Babin, C., Bourgois, J., Broutin, J., Lardeux, H., Pons, D. and Rachæuf, P. (1986). Les grandes ètapes de l'évolition paléozoïque du Massif de Santander (Andes de Colombie): signification de la discordance du Dévonien moyen. *Comptes Rendus de l'Académie des Sciences de Paris*, **303**, Serie II (8), 707-712.

Boucot, A.J. (1972). Silurian of Venezuela. In: Correlation of the South American Silurian Rocks, eds. Berry, W.B.N. and Boucot, A.J. *Geological Society of America Special Paper*, **133**, 53-54.

Bourgois, J., Eguez, Butterlin, J. and De Wever, P. (1990). Évolution géodynamique de la Cordillère Occidentale des Andes d'Équateur: la découverte de la formation éocene d'Apagua. *Comptes Rendus de l'Académie des Sciences de Paris*, **311**, Serie II, 173-180.

Bourgois, J., Toussaint, J. F., Gonzalez, H., Azema, J., Calle, B., Desmet, A., Murcia, L. A., Acevedo, A. P., Parra, E., and Tourton, J. (1987). Geological history of the Cretaceous ophiolitic complexes of northwestern South America (Colombian Andes). *Tectonophysics*, **143**, 307-327.

Branquet, Y., Laumonier, B., Lopes, B., Cheilletz, A., Giuliani, G. and Rueda, F. (1996). Evidences of compressive structures in the Muzo and Coscuez Emerald Deposits, Eastern Cordillera of Colombia. *3rd International Symposium of Andean Geodynamics*, 675-677.

Bucher, W.H. (1952). Structure and orogenic history of Venezuela. *Geological Society of America Memoir*, **49**, 1-113.

Burgl, H. (1961). Historia geológica de Colombia. *Revista de la Academia Colombiana de Ciencias Exactas, Físicas y Naturales*, **11**, 137-194.

Burkley, L.A. (1976). *Geochronology of the central Venezuelan Andes*. Ph.D. Thesis, Case Western University, 150 p.

Butler, K. and Schamel, S. (1988). Structure along the eastern margin of the Central Cordillera, Upper Magdalena valley, Colombia. *Journal of South American Earth Sciences*, **1**(1), 109-120.

Campbell, C.J. (1968). The Santa Marta wrench fault of Colombia and its regional setting. *Transactions of the 4th Caribbean Geology Conference*, 247-261.

Campbell, C.J. (1974). Ecuadorian Andes. In: Mesozoic-Cenozoic Orogenic Belts, ed. Spencer, A.M. *Geological Society of London Special Publication*, **4**, 725-732.

Case, J.E. and MacDonald, W.D. (1973). Regional gravity anomalies and crustal structure in Northern Colombia. *Geological Society of America Bulletin*, **84**, 2905-2916.

Castrillo, J.T. and Hervouet, Y. (1996). Tectonic model of thrusts and imbricated thrusting wedge on the northwestern flank of the Merida Andes between Torondoy and Valera (Venezuela). *Andean Geodynamics, Troisième International Symposium sur la Géodynamique Andine, St. Malo, France*, 307-310.

Cediel, F. (1981). Las Formaciones Luisa, Payandé, Saldaña. Sus columnas estratigráficas characterísticas. *Geología Norandina*, **3**, 11-19.

Cediel, F., Mojica, J., and Macia, C. (1980). Definición estratigráfica del Triásico en Colombia, Suramérica, Formaciones, Luisa, Payandé y Saldaña. *Newsletter on Stratigraphy*, 9, 73-104.

Cheilletz, A., Feraud, G., Giuliani, G. and Rodriguez, C.T. (1994). Time-pressure and temperature constraints on the formation of Colombian Emeralds: An 40Ar/39 Ar laser microprobe and fluid inclusion study. *Economic Geology*, **89**, 361-380.

Chigne, N., Loureiro, D. and Rojas, E. (1997). El piedemonte de la Cordillera Oriental de Colombia y do los Andes de Mérida: Estilos estructurales y consideraciones sobre la génesis y migración de hidrocarburos. *VI Simposio Bolivariano, Memorias*, **1**, 457-477.

Colletta, B., Hebrard, F., Letouzey, J., Werner, P. and Rudkiewicz, J-L. (1990). Tectonic style and crustal structure of the Eastern Cordillera (Colombia) from a balanced cross-section. In: *Petroleum and Tectonics in Mobile Belts*, ed. Letouzey, J., pp. 81-100. Editions Technip.

Colletta, B., Roure, F., de Toni, B., Loureiro, D., Passalacqua H. and Gou, Y. (1997). Tectonic inheritance, crustal architecture and contrasting structural styles in the Venezuela Andes. *Tectonics*, **16**(5), 777-794.

Colony, R.J. and Sinclair, J.H. (1928). The lavas of the volcano Sumaco, eastern Ecuador, South America. *American Journal of Science*, **16**, 229-312.

Cosma, L., Lapierre, H., Jaillard, E., Laubacher, G., Bosch, D., Desmet, A., Mamberti, M. and Gabriele, P. (1998). Pétrographie et géochimie des unités magmatiques de la Cordillère occidentale d'Équateur (0°30'S): implications tectoniques. *Bulletin de la Société Geologique de France*, **169**, 739-751.

De Loczy, L. (1968). Geotectonic evolution of the Amazon, Parnaíba, and Paraná basins. *Anais da Academia Brasileira de Ciências*, 40 (suplemento), 231-249.

De Loczy, L. (1970). Tectonismo transversal na América do Sul e suas relações genéticas com as zonas de fraturas de cadeias meio-oceânicas. *Anais Academia Brasileira de Ciências*, **42**, 185-205.

Deratmiroff, G.N. (1971). Late Cenozoic imbricate thrusting in the Venezuelan Andes. *American Association of Petroleum Geologists Bulletin*, **55**(8), 1336-1355.

De Toni, B. and Kellogg, J. (1993). Seismic evidence for blind thrusting of the northwestern flank of the Venezuela Andes. *Tectonics*, **12**(6), 1393-1409.

Dengo, C.A., and Covey, M.C. (1993). Structure of the Eastern Cordillera of Colombia: Implications for traps and regional tectonics. *American Association of Petroleum Geologists Bulletin*, **77**, 1315-1337.

De Souza, H.A.F., Espinoza, A. and Delaloye, M. (1984). K/Ar ages of basic rocks in the Patia Valley, southwest Colombia. *Tectonophysics*, **107**, 135-145.

Echeverria, L.M. (1980). Tertiary or Mesozoic Komatiites from the *Gorgona Island, Colombia: Field relationship* and geochemistry. *Contributions to Mineralogy and Petrology*, 73, 253-266.

Eguez, A. (1986). *Evolution cénozoique de la Cordillère occidentale septentrionale d'Equateur (0°15'S à 1°10'S). Les minéralisation associées.* Thèse Doctoral Université Paris VI, 116 p.

Etayo-Serna, F. (1979). Zonation of the Cretaceous of Central Colombia by Ammonites. *Publicaciones Geológicas Especiales del Ingeominas*, **2**, 186 p.

Etayo-Serna, F., and Barrero, D. (1983). Mapa de terrenos geológicos de Colombia. *Instituto Nacional de Investigaciones Geológico-Mineras, Publicaciones Especiales*, **14**, 235 p.

Fabre, A. (1987). Tectonique et génération d'hydrocarbures: un modèle de l'évolution de la Cordillère Orientale de Colombie et du bassin des Llanos pendant le Crétacé et le Tertiaire. *Archives des Sciences, Geneve*, 40, 145-190.

Faucher, B. and Savoyat, E. (1973). Esquisse géologique des Andes de l'Équateur. *Revue de Géographie Physique et de Géologie Dynamique*, **15**, 115-142.

Feininger, T. (1970). The Palestina Fault, Colombia. *Geological Society of America Bulletin*, **81**, 1201-1216.

Feininger, T. (1980). Eclogite and related high-pressure regional metamorphic rocks from the Andes of Ecuador. *Journal of Petrology*, **21**, 107-140.

Feininger, T. (1982). Glaucophane schists in the Andes of Jambaló, Colombia. *Canadian Mineralogist*, **20**, 41-47.

Feininger, T. and Bristow, C.R. (1980). Cretaceous and Paleogene geology of coastal Ecuador. *Geologische Rundschau*, **69**, 849-874.

Feininger, T., and Seguin, M. K. (1983). Simple Bouguer gravity anomaly field and the inferred crustal structure of continental Ecuador. *Geology*, **11**, 40-44.

Figueroa de Sánchez, L. and Hernández, L. (1990). Exploración geofísica-geológica del área de Guarumen. *Transations 5th Venezuelan Geophysical Congress*, 219-222.

Forero-Suarez, A. (1990). The basement of the Eastern Cordillera, Colombia: an allochthonous terrane in northwestern South America. *Journal of South American Earth Sciences*, **3**, 141-151.

Gansser, A. (1955). Ein Beitrag zur Geologie und Petrographie der *Sierra Nevada de Santa Marta (Kolumbien, Sudamerika). Schweizerische Mineralogische und Petrographische Mitteilungen*, **35**, 209-279.

Gansser, A. (1973). Facts and theories on the Andes. *Journal of the Geological Society of London*, **129**, 93-131.

Gerth, H. (1955). *Der Geologische Bau der Südamerikanischen Kordillere.* Berlin: Gebr. Borntraeger, 264 p.

Geyer, O.F. (1974). Der Unterjura (Santiago-Formation) von Ekuador. *Neues Jarhbuch für Geologie und Paläontologie Monatshefte*, **9**, 525-541.

Geyer, O. (1980). Die Mesozoische Magnafazies-Abfolge in den Nordanden (Peru, Ekuador, Kolumbien). *Geologische Rundschau*, **69**(3), 875-891.

Gonzáles de Juana, C. (1952). Introducción al estudio de la geología de Venezuela. *Boletín de Geología, Caracas*, **2**, 407-416.

Gose, W., Castillo, J. and Pararnau, A. (1989). Resultados paleomagnéticos de rocas sedimentarias cretácicas de Venezuela. *VII Congreso Geológico Venezolano*, 725-730.

Goossens, P.J. and Rose, W.I. (1973). Chemical composition and age determination of tholeiitic rocks in the basic Igneous Complex, Ecuador. *Geological Society of America Bulletin*, **84**, 1043-1052.

Goossens, P. J., Rose, W. I., and Flores, D. (1977). Geochemistry of tholeiites of the Basic igneous complex, Ecuador. *Geological Society of America Bulletin*, **88**, 1711-1720.

Grauch, R. (1972). Preliminary report of a Late(?) Paleozoic metamorphic event in the Venezuelan Andes. In: Studies in Earth and Space Science, ed. Hess, H.H. *Geological Society of America Memoir*, **132**, 465-473.

Grösser, J.R. and Prössl, K.F. (1991). First evidence of Silurian in Colombia: Stratigraphical data from the Quetame Massif, Cordillera Oriental. *Journal of South American Earth Sciences*, **4**, 139-157.

Gutscher, M.-A., Malavieille, J., Lallemand, S. and Collot, J.-Y. (1999) Tectonic segmentation of the North Andean margin: impact of the Carnegie Ridge collision. *Earth and Planetary Science Letters*, **168**, 255-270.

Hall, M.L., and Calle, J. (1982). Geochronological control for the main tectonic-magmatic events of Ecuador. *Earth Science Reviews*, **18**, 215-239.

Hall, M.L. and Wood, C.A. (1985). Volcano-tectonic segmentation of the Northern Andes, *Geology*, **13**, 203-207.

Ham, C.K., and Herrera Jr., L.J. (1963). Role of Subandean fault system in tectonics of eastern Peru and Ecuador. In: Backbone of the Americas, eds. Childs, O. and Beebe, B.W. *American Association of Petroleum Geologists Memoir*, **2**, 47-61.

Handschumacher, D.W. (1976). Post-Eocene plate tectonics of the eastern Pacific. In: Sutton, G.H. et al. (eds.) *American Geophysical Union Monograph*, **19**, 177-202.

Harrington, H.J., and Kay, M. (1951). Cambrian and Ordovician faunas of Eastern Colombia. *Journal of Paleontology*, **25**, 655-668.

Herz, N., Hasui, Y, Sena Costa, J.B. and da Silva Mattta, M.A. (1989). The Araguá Foldbelt, Brazil: A reactivated Braziliano-Pan African cycle (550 Ma) geosuture. *Precambrian Research*, **42**, 371-386.

Hooper, R.J., Urreiztieta, M. de, Eichenseer, H.T., Ballard, J-F. and Suter, J. (1999). The tectono-stratigraphic development of the Merida Andes, Western Venezuela, and the interaction between Caribbean and Andean Tectonics. In: McClay, K. (ed.). *Thurst Tectonics Conference, London, Abstracts*, 115-118.

Hörmann, P.K. and Pichler, H. (1982). Geochemistry, petrology and origin of the Cenozoic rocks of the northern Andes in Ecuador. *Journal of Volcanology and Geothermal Research*, 12, 259-282.

Hungerbuehler, D., Steinmann, M., Winkler, W. and Seward, D. (1996). Neogene fission-track stratigraphy of the southern Ecuadorian basins: Implications for regional tectonic history. In: *3rd International Symposium of Andean Geodynamics-Saint-Malo*, ORSTON ed., 387-390.

Irving, E.M. (1971). La Evolución estructural de los Andes más septentrionales de Colombia. *Boletin de Geología, Bogotá*, **19**(2), 1-90.

Jaillard, E., Soler, P., Carlier, G., and Mourier, T. (1990). Geodynamic evolution of the Northern Andes and Central Andes during early to middle Mesozoic times. *Journal of the Geological Society of London*, **147**, 1009-1022.

Jaillard, E., Ordoñez, M., Bengtson, P., Berrones, G., Bonhomme, M., Jimenez, N. and Zambrano, I. (1996). Sedimentary and tectonic evolution of the arc zone of southwestern Ecuador during Late Cretaceous and Early Tertiary times. *Journal of South American Earth Sciences*, **9**, 131-140.

Jaramillo, J.M. (1978). Rocas metamórficas de alto grado-granulites- en algunas lavas del Nevado del Ruiz, Colombia. *Resúmenes del Segundo Congreso Colombiano de Geología*, 16 p.

Julivert, M. (1973). Les traits structuraux et l'évolution des Andes Colombiennes. *Revue de Géographie Physique et de Géologie Dynamique*, **15**, 143-156.

Julivert, M. and Tellez, N. (1963). Sobre la presencia de fallas de edad Precretácea y post-Girón (Jura-triássico) en el flanco W del Macizo de Santander (Cordillera Oriental, Colombia). *Boletín de Geología, Universidad Industrial de Santander, Bucaramanga*, **12**, 5-17.

Kellogg, J.N. (1984). Cenozoic tectonic history of the Sierra de Perijá, Venezuela-Colombia, and adjacent basins. In: The Caribbean-South American Plate Boundary and Regional Tectonics, eds. Bonini, W.E., Hargraves, R.B. and Shagam, R. *Geological Society of America Memoir*, **162**, 263-285.

Kellogg, J.N., and Bonini, W.E. (1982). Subduction of the Caribbean plate and basement uplifts in the overriding South American plate. *Tectonics*, **1**, 251-276.

Kennerley, J.B. (1980). Outline of the Geology of Ecuador. Institute of Geological Sciences (London). *Overseas Geology and Mineral Resources*, **55**, 1-17.

Kerr, A.C., Tarney, J., Marriner, G., Nivia, A., Klaver, G.Th. and Sauders, A.D. (1996). The geochemistry and tectonic setting of Late Cretaceous Caribbean and Colombian volcanism. *Journal of South American Earth Sciences*, **9**, 111-120.

Keppie, J.D. and Ortega-Gutierrez, F. (1995). Provenance of Mexican Terranes: Isotopic constraints. *International Geology Review*, 37, 813-824.

Kesler, S., Sutter, J., Issigonis, M.J., Jones, L.M. and Walker, R. (1977). Evolution of porphyry copper mineralization in an oceanic island arc; Panama; a discussion. In: Porphyry copper deposits of the southwestern Pacific islands and Australia, eds. Gustafson, L.B. and Titley, S.R. *Economic Geology and the Bulletin of the Society of Economic Geologists*, **72**, 1142-1153

Kohn, B.P., Shagam, R. and Subieta, T. (1984). Results and preliminary implications of sixteen fission-track ages from rocks of the Western Caribbean Mountains, Venezuela. In: The Caribbean South American Plate Boundary and Regional Tectonics, eds. Bonini, W.E., Hargraves, R.B. and Shagam, R. *Geological Society of America Memoir*, **162**, 415-421.

Kovisars, L. (1971). Geology of a portion of the North-Central Venezuela Andes. *Geological Society of America Bulletin*, **82**, 3111-3138.

Kroonenberg, S. (1982). A Grenvillian granulite belt in the Colombian Andes and its relationship to the Guayana Shield. *Geologie en Mijnbouw*, **61**, 325-333.

Kroonenberg, S., Pichler, H. and Diederix, H. (1982). Cenozoic alkalibasaltic to ultramafic volcanism in the uppermost Magdalena Valley, southern Huila Department, Colombia. *Geología Norandina*, 19-26.

Lavenu, A., Bonhomme, M.G., Eguez, A., Dugas, V. and Vivier, G. (1992). New k-Ar age dates of the Neogene and Quaternary volcanic rocks from the Ecuadorian Andes: Implications for the relationship between sedimentation, volcanism and tectonics. *Journal of South American Earth Sciences*, **5**, 309-320.

Laubscher, H.P. (1987). The kinematic puzzle of the Neogene northern Andes. In: *The Anatomy of Mountain Ranges*, eds. Schaer, J-P. and Rogers, J., pp. 211-227, chapter 11. Princeton University Press.

Lebrat, M., Megard, F., Dupuy, C. and Dostal, J. (1987). Geochemistry and tectonic setting of pre-collision Cretaceous and Paleogene volcanic rocks of Ecuador. *Geological Society of America Bulletin*, **99**, 569-578.

Lebrat, M., Mégard, F., Juteau, T. and Calle, J. (1985). Pre-orogenic volcanic assemblages and structure in the Western Cordillera of Ecuador between 1o40'S and 2°20'S. *Geologische Rundschau*, **74**, 343-351.

Liddle, R.A. (1946). *The geology of Venezuela and Trinidad*. Paleontological Research Institute, Ithaca, NY, 890 p.

Litherland, M. and Aspden, J.A. (1992). Terrane-boundary reactivation: A control on the evolution of the Northern Andes. *Journal of South American Earth Sciences*, **5**, 71-76.

Litherland, M., Aspden, J.A. and Jemielita, R.A. (1994). The metamorphic belts of Ecuador. *British Geological Survey Overseas Memoir*, **11**, 147 p.

Litherland, M., O'Connor, E.A. and Pitfield, P.E.J. (1985). Andean trending mobile belts in the Brazilian Shield. *Nature*, **314**, 345-348.

Lonsdale, P. (1978). Ecuadorian subduction system. *American Association of Petroleum Geologists Bulletin*, **62**, 2454-2477.

Lugo, J. (1994). The Merida Arch: Tectonic Control on Deposition From Late Mesozoic to Early Cenozoic in Western Venezuela. *V Simposio Bolivariano, Memoria*, 291-310.

Malavé, G. and Suárez, G. (1993). Recent seismicity (mb 5.4) in northwestern Venezuela: regional tectonic implications. *Second ISAG, Oxford (UK)*, 111-114.

MacDonald, W.D. and Hurley, P.M. (1969). Precambrian gneisses from northern Colombia, South America. *Geological Society of*

America Bulletin, **80**, 1867-1872.

Marechal, P. (1983). *Les Temoins de Chaîne Hercynienne dans le noyau ancien des Andes de Mérida (Vénézuéla)*. Ph.D. Thesis, Université de Bretagne Occidentale, Brest, 176 p.

Marchant, S. (1961). A photogeological analysis of the structure of the Western Guayas Province, Ecuador. *Journal of the Geological Society of London*, **115**, 317-338.

Marksteiner, R. and Alemán, A.M. (1997). Petroleum systems along the fold belt associated to the Marañon-Oriente-Putumayo (MOP) foreland basins. *6th Simposio Bolivariano-Petroleum Exploration in the Subandean Basins*, II, 63-74.

Martinez, M. (1970). *Montañas de Amotape y posible origen del* petróleo en rocas paleozoicas del noroeste del Perú. *Primer Congreso Latinoamericano de Geología, Lima, Geología del Petróleo*, **II**, 105-138.

Marriner, G.F. and Millward, D. (1984). The petrology and geochemistry of Cretaceous to Recent volcanism in Colombia: the magmatic history of an accretionary plate margin. *Journal of the Geological Society of London*, **141**, 473-486.

Maya, M. (1992). Catálogo de dataciones isotópicas en Colombia. *Boletín Geológico*, Bogotá, **32**, 135-187.

Maze, W.B. (1984). Jurassic La Quinta Formation in the Sierra de Perijá, northwestern Venezuela: Geology and tectonic environment of red beds and volcanic rocks. In: The Caribbean-South American Plate Boundary and regional Tectonics, eds. Bonini, W.E., Hargraves, R. and Shagam, R. *Geological Society of America Memoir*, **162**, 263-285.

McCourt, W.J. and Feininger, T. (1984) High-pressure metamorphic rocks in the Central Cordillera of Colombia. *British Geological Survey Report Series*, **16**(1), 28-35.

McCourt, W.J., Aspden, J.A. and Brook, M. (1984). New geological and geochronological data for the Colombian Andes: Continental growth by multiple accretion. *Journal of the Geological Society of London*, **141**, 831-845.

Meier, B.P., Shwander, M.M. and Laubscher, H.P. (1987). The tectonics of Táchira: a sample of north Andean tectonics: In: *The Anatomy of Mountain Ranges*, eds. Schaer, J-P. and J. Rodgers, pp. 228-237. Princeton University Press.

Millward, D., Marriner, G.F. and Saunders, A.D. (1984). Cretaceous tholeiitic volcanic rocks from the Western Cordillera of Colombia. *Journal of the Geological Society of London*, **141**, 847-860.

Mojica, J., Kammer, A. and Ujeta, G. (1996). El Jurásico del sector noroccidental de Suramérica y guía de la excursión al Valle Superior del Magdalena (Nov. 1-4/95), regiones de Payandé y Prado, Departamento del Tolima, Colombia. *Geología Colombiana*, **21**, 3-40.

Mojica, J., Villaroel, C., Cuerda, A. and Alfaro, M. (1988). La fauna de graptolitos de la Formación Higado (Llanvirniano-?Landeiliano), Serranía de Las Minas, Valle Superior del Magdalena, Colombia. *V° Congreso Geológico Chileno*, **II**, C189-C202.

Mooney, W.D. (1979). *Seismic refraction studies of the Western Cordillera, Colombia and the East Pacific-Caribbean Ridge during the Jurassic and Cretaceous and the evolution of Western Colombia*. Ph.D. Thesis, University of Wisconsin, 84 p.

Mothes, P.A., Hall, M.L., and Jada, R.J. (1998). The enormous Chillos Valley Lahar: and ash-flow generated debris flow from the Cotopaxi Volcano, Ecuador. *Bulletin of Volcanology*, **59**, 233-244.

Mourier, T. (1988). *La transition entre Andes marginales et Andes cordillèraines à ophiolites. Evolution sédimentaire, magmatique et structurale du relai de Huancabamba (3°-8°S) nord Pérou-Sud Equateur*. Ph.D Thesis, Université de Paris, 275 p.

Mourier, T., Laj, C., Megard, F., Roperch, P., Mitouard, P. and Farfan, M. (1988). An accreted continental terrane in northwestern Peru. *Earth and Planetary Science Letters*, **88**, 182-192.

Nelson, J. (1957). Contribution to the geology of the Central and Western Cordillera of Colombia in the section between Ibague and Cali. *Leidse Geologische Mededelingen, Leiden*, **22**, 1-76

Noble, S.R., Aspden, J.A., and Jemielita, R. (1997). Northern Andean crustal evolution: New U-Pb geochronological constraints from Ecuador. *Geological Society of America Bulletin*, **109**, 789-798.

Pardo-Casas, F. and Molnar, P. (1987). Relative motion of the Nazca plate (Farallon) and South American plates since Late Cretaceous. *Tectonics*, **6**, 233-248.

Parnaud, F., Gou, Y., Pascual, J-C, Capello, M.A., Truskowski, I. and H. Passalacqua (1995). Stratigraphic Synthesis of Western Venezuela. In: Petroleum Basins of South America, ed. Tankard et al. *American Association of Petroleum Geologists Memoir*, **62**, 681-698.

Pennington, W.D. (1981). Subduction of the Eastern Panama Basin and seismotectonics of Northwest South America. *Journal of Geophysical Research*, **86**, 10753-10770.

Pierce, G.R., Jefferson Jr., C.C. and Smith, W.R. (1961). Fossiliferous Paleozoic localities in the Merida Andes, Venezuela. *American Association of Petroleum Geologists Bulletin*, **45**(3), 342-375.

Pindell, J.L. (1985). Alleghenian reconstruction and subsequent evolution of the Gulf of Mexico, Bahamas, and Proto-Caribbean. *Tectonics*, **4**(1), 1-39.

Pindell, J.L. and Barrett, S.F. (1990). Geological evolution of the Caribbean region: a plate-tectonic perspective. In: *The Caribbean Region, The geology of North America*, eds. Dengo, G. and Case, J.E. *Geological Society of America*, H, Chapter 16, 405-432.

Pitchler, H., Stibane, F.R. and Weyl, R. (1974). Basischer Magmatismus und Krustenbahn im südlichen Mittelamerica, Kolumbien und Ecuador. *Neues Jahrbuch für Geologie und Paläontologie Monatshefte*, **1974**, 102-126.

Priem, H.N.A., Kroonenberg, S.B., Boelrijk, N.A.I.M., and Hebeda, E.H. (1989). Rb-Sr evidence for the presence of a 1.6 Ga. basement underlying the 1.2 Ga. Garzón-Santa Marta Granulite Belt in the Colombian Andes. *Precambrian Research*, **42**, 315-324.

Priem, H.N.A., Andriessen, P.A.M., Boelrijk, N.A.I.M., de Boorder, H., Hebeda, E.H., Huggett, A., Verdurmen, E.A.Th. and Verschure, R.H. (1982). Geochronology of the Precambrian in the Amazonas region of southeastern Colombia (Western Guiana Shield). *Geologie en Mijnbouw*, **61**, 229-242.

Prössl, K.F. and Grösser, J.R. (1995). The age of the Formación Amoya, Upper Magdalena Valley: Another piece in the Paleozoic stratigraphic puzzle of Colombia. *Newsletters on Stratigraphy*, **32**, 91-101.

Radelli, L. (1962). Introdución al estudio de la petrografía del Macizo de Garzón (Huila, Colombia). *Geología Colombiana*, **3**, 17-46.

Radelli, L. (1967). *Géologie des Andes Colombiennes*. Travauxs du Laboratorie de Géologie, Facultad de Sciences de Grenoble, 6, 457 p.

Ramos, V.A. (1988). Tectonics of the Late Proterozoic-Early Paleozoic: a collisional history of southern South America. *Episodes*, **11**(3), 168-174.

Ramos, V.A. and Kay, S.M. (1990). Triassic rifting and associated basalts in the Cuyo basin, central Argentina. In: Andean magmatism and its tectonic setting, eds. Harmon, R.S. and Rapela, C.W. *Geological Society of America Special Paper*, **285**, 79-91.

Renz, O. (1959). Estratigrafía del cretáceo en Venezuela. *Boletín de Geología*, Caracas, **5**(10), 3-48.

Renzoni, G. (1968). Geología del macizo de Quetame. *Geología Colombiana*, **5**, 75-127.

Restrepo-Pace, P.A. (1992). Petrotectonic characterization of the Central Andean Terrane, Colombia. *Journal of South American Earth Sciences*, 5, 97-116.

Restrepo, J.J. and Toussaint, J.F. (1973). Obducción cretácea en el occidente Colombiano. *Publicaciones Especiales de Geología, Universidad Nacional, Facultad de Minas, Medellín*, 3, 26 p.

Restrepo, J.J. and Toussaint, J-F. (1974). Obducción Cretácea en el Occidente Colombiano. *Anales de la Facultad de Minas*, Medellín, 58, 73-105.

Restrepo, J.J. and Toussaint, J-F. (1977). Anfibolitas de Caldas-Antioquia. *Boletín de Ciencias de la Tierra, Medellín*, 24 p.

Restrepo, J.J. and Toussaint, J-F. (1978). Ocurrencia del Precámbrico en las cercanías de Medellín, Cordillera Central de Colombia. *Publicaciones Especiales de Geología, Universidad Nacional, Facultad de Minas, Medellín*, 12, 1-11.

Restrepo, J.J. and Toussaint, J-F. (1988). Terranes and continental accretion in the Colombian Andes. *Episodes*, 11, 189-193.

Restrepo, J.J., Toussaint, J.F. and González, H. (1981). Edades mio-pliocenas del magmatismo asociado a la Formación Combia, Departamento de Antioquia y Caldas, Colombia. *Geología Norandina*, 3, 21-26.

Restrepo-Pace, P.A., Ruiz, J., Gehrels, G. and Cosca, M. (1997). Geochronology and Nd isotopic data of Grenville-age rocks in the Colombian Andes: new constraints for Late Proterozoic-Early Paleozoic paleocontinental reconstructions of the Americas. *Earth and Planetary Science Letters*, 150, 427-441.

Rod, E. (1956). Strike Slip Faults of Northern Venezuela. *American Association of Petroleum Geologists Bulletin*, 40(3), 457-475.

Romeuf, N., Aguirre, L., Soler, P., Feraud, G., Jaillard, E. and Ruffet, G. (1995). Middle Jurassic volcanism in the Northern and Central Andes. *Revista Geológica de Chile*, 22(2), 245-259.

Roperch, P. Mégard, F., Laj, C., Mourier, T., Clube, T. and Noblet, C. (1987). Rotated oceanic blocks in western Ecuador. *Geophysical Research Letters*, 14, 558-561.

Salinas, R. and Tistl, M. (1991). A Tertiary ultramafic complex and komatiitic basalts from Condoto, Chocó, NW Colombia. *Zentralblatt für Geologie und Paläontologie*, 6, 1659-1678.

Samaniego, P., Monzier, M., Robin, C., and Hall, M.L. (1998). Late Holocene eruptive activity at Nevado Cayambe Volcano, Ecuador. *Bulletin of Volcanology*, 59, 451-459.

Sánchez, T. (1984). Étude paléoécologique du Paléozoïque Supérieur dans le nord des Andes du Venezuela. *Collection Biostratigraphie du Paléozoïque, Université de Bretagne Occidentale*, 2, 1-165.

Sánchez, T. (1985). Algunos bivalvos de la Formación El Horno (Silúrico) de los Andes de Mérida (Venezuela). *6° Congreso Geológico Venezolano*, 1120-1134.

Sauer, W. (1971). *Geologie von Ecuador*. Berlin, Gebrüder Borntraeger, 316 p.

Schubert, C. (1971). Observaciones geomorfológicas y glaciales en el área de Pico Bolivar, Sierra Nevada de Mérida, Venezuela. *Boletín Informativo, Asociación Venezolana de Geología, Minería y Petróleo*, 14(8), 193-216.

Schubert, C. (1984). Basin formation along the Bocono Fault, El Moron-El Pilar System, Venezuela. *Journal of Geophysical Research*, 89, 5711-5718.

Schubert, C. and Vivas, L. (1993). *El Cuaternario de la Cordillera de Mérida, Andes Venezolanos*. Universidad de Los Andes, Fundación Polar, Mérida, 345 p.

Shagam, R. (1972). Andean research project, Venezuela: Principal data and tectonic implications. In: Studies in Earth and Space Science, ed. Hess, H.H. *Geological Society of America Memoir*, 132, 449-463.

Shagam, R. (1975). The northern termination of the Andes. In: *The Ocean basins and margins, v.3, Gulf of Mexico and the Caribbean*, eds. Nair, A.M. and Stehli, F.G., pp. 326-419, chap. 9. New York: Plenum Press.

Shagam, R. (1977). Stratigraphic models for the Northern Venezuela Andes. *5° Congreso Geológico Venezolano*, 855-877.

Shagam, R. and Hargraves, R.B. (1970). Geologic and paleomagnetic study of Permo-Carboniferous red beds (Sabaneta and Mérida facies) Venezuela Andes. *American Association of Petroleum Geologists Bulletin*, 54(12), 2336-2348.

Sheppard, G. (1937). *The Geology of Southwestern Ecuador*. London: Murby, 275 p.

Sillitoe, R.H., Jaramillo, L., Damon, P.E., Shafiqullah, M. and Escobar, R. (1982). Setting, Characteristics and age of the Andean porphyry copper belt in Colombia. *Economic Geology*, 77, 1837-1850.

Soler, P. and Bonhomme, M.G. (1990). Relations of magmatic activity to plate dynamics in Central Peru from Late Cretaceous to Present. In: Plutonism from Antarctica to Alaska, eds. Kay, D. and Rapela, C. *Geological Society of America Special Paper*, 241, 173-192.

Soulas, J.P. (1985). Neotectónica del flanco occidental de Los Andes de Venezuela entre 70°30' y 71°00' (Fallas Bocono de Valera, Tuname, Piñango y del Piemonte). *VI Congreso Geológico Venezolano, Memoria*, IV, 2687-2711.

Spadea, P. and Espinosa, A. (1996). Petrology and chemistry of Late Cretaceous volcanic rocks from the southernmost segment of the Western Cordillera of Colombia. *Journal of South American Earth Sciences*, 9, 79-90.

Stephan, J.F. (1982). *Evolution geodynamique du domaine caraibe Andes et chaines caraibe sur la transversele de Barquisimeto, Venézuéla*. Ph.D. Thesis, Université Pierre et Marie Curie, Paris, 1 and 2, 512 p.

Stephan, J.F., De Lepinay, B.M., Calais, E., Tardy, M., Beck, C., Carfantan, J.C., Oliver, J.L., Vila, J.M., Bouysse, P., Mauffret, A., Borgois, J., Thery, J.M., Toumon, J., Blanchet, R., and Dercourt, J. (1990). Paleogeodynamic maps of the Caribbean: 14 steps from Lias to Present. *Bulletin de la Société Géologique de France*, 8, 915-919.

Thompson, M.L. and Miller, A.K. (1949). Permian fusulinids and cephalopods from the vicinity of the Maracaibo Basin in Northern South America. *Journal of Paleontology*, 23(1), 1-23.

Toni, B. de and Kellogg, J. (1993). Seismic evidence for blind thrusting of the Northwestern Flank of the Venezuelan Andes. *Tectonics*, 12(6), 1393-1409.

Toussaint, J-F. (1978). Grandes rasgos geológicos de la parte septentrional del Occidente Colombiano. *Boletín de Ciencias de la Tierra, Universidad Nacional de Colombia, Medellín*, 3, 231 p.

Toussaint, J-F. (1993). *Evolución Geológica de Colombia, Precámbrico-Paleozoico*. Universidad Nacional de Colombia, Medellín, 229 p.

Toussaint, J-F. and Restrepo, J. J. (1982). Magmatic evolution of the northwestern Andes of Colombia. *Earth Science Reviews*, 18, 205-213.

Tribovillard, N-P., Stephan, J. F., Manivit, H., Reyre, Y., Cotillon, P. and Jautee, E. (1991). Cretaceous black shales of Venezuela Andes: preliminary results on stratigraphy and paleoenvironmental interpretations. *Palaeogeography, Palaeoclimatology and Palaeoecology*, 81, 313-321.

Trumpy, D. (1943). Pre-Cretaceous of Colombia. *Geological Society of America Bulletin*, 54, 1281-1304.

Tschanz, C.M., Marvin, R.F., Cruz, J., Mehnert, H.H. and Cebula, G.T. (1974). Geologic evolution of the Sierra Nevada de Santa Marta, Northeastern Colombia. *Geological Society of America Bulletin*, 85, 273-284.

Tschopp, H.J. (1953). Oil exploration in the Oriente of Ecuador. *American Association of Petroleum Geologists Bulletin*, 37, 2303-2347.

Ulloa, C.E., Perez, V. E. and Baldis, B.A.J. (1982). Unidades litoestratigráficas del Ordovícico de los Llanos Orientales de Colombia. *Cuarto Congreso Latinoamericano de Geología, Actas*, **I**, 109-120.

Van der Wiel, A.M. (1991). *Uplift and volcanism of the SE Colombian Andes in relation to Neogene sedimentation in the Upper Magdalena Valley*. Ph.D. Thesis, University of Wageningen, Amsterdam, 208 p.

Van Thournout, F., Hertogen, J. and Quevedo, L. (1992). Allochthonous terranes in northwestern Ecuador. *Tectonophysics*, **205**, 205-221.

Villamil, T. and Arango, C. (1998). Integrated stratigraphy of Latest Cenomanian and Early Turonian facies of Colombia. In: Pindell, J.L. and Drake, C.L. (eds.). *Paleogeographic Evolution and Non-glacial Eustasy, Northern South America*, SEPM Special Publication, **58**, 129-159.

Villarroel, C., Macia, C. and Brieva, J. (1997). Formación Venado, nueva unidad litoestratigráfica del Ordovícico Colombiano. *Geología Colombiana*, **22**, 41-49.

Ward, D.E., Goldsmith, R., Cruz, J., and Restrepo, H. (1973). Geología de los Cuadrángulos H-12 Bucaramanga y H13 Pamplona, Departamento de Santander. *Boletín Geológico*, **21**, 133 p.

White, C. (1985). Patron estructural y prospectividad del frente norandino entre La Fría y el Vigía. *Sexto Congreso Geológico Venezolano*, 5397-5416.

Yañez, P., Ruiz, J., Patchett, Ortega-Gutierrez, R. and Gehrels, G.E. (1991). Isotopic studies of the Acatlan complex, southern Mexico: Implications for Paleozoic North American Tectonics. *Geological Society of America Bulletin*, **103**, 817-828.

Zambrano, E., Vásques, E., Duval, B., Latrielle, M. and Coffinieres, B. (1971). Síntesis palogeográfica y petrolera del occidente de Venezuela. *Cuarto Congreso Geológico Venezolano, Memoria*, **1**, 481-552.

Tectonic Evolution of the Andes of Ecuador, Peru, Bolivia and Northernmost Chile

E. Jaillard, G. Hérail, T. Monfret, E. Díaz-Martínez, P. Baby, A. Lavenu, and J.F. Dumont

This chapter was prepared under the co-ordination of E. Jaillard. Together with G. Hérail and T. Monfret, he wrote the Introduction. Enrique Díaz-Martínez prepared the section on the Pre-Andean evolution of the Central Andes. Again Jaillard, on the Pre-orogenic evolution of the North-Central Andes. E. Jaillard, P. Baby, G. Hérail, A. Lavenu, and J.F. Dumont wrote the text on the orogenic evolution of the North-Central Andes. And, finally, Jaillard closed the manuscript with the conclusions.

Introduction:

The Present-Day North-Central Andes (1°N - 23°S)

The Andean Chain is the major morphological feature of the South American Continent. This 8000 km long mountain belt extends along the western border of the South American Plate, and can be divided into three segments of distinct orientations, separated by two major bends. The NNE-SSW trending Colombian-Ecuadorian segment (12°N - 5°S) is 2000 km long and includes part of northernmost Peru and easternmost Venezuela. It is separated from the Peruvian segment by the Huancabamba Bend (Mégard, 1987). Its northern end (eastern Venezuela) exhibits a change to an E-W orientation, due to its connection to the South Caribbean dextral transform system. The Peruvian segment (5°S - 18°S) is 2000 km long and its orientation is close to NW-SE. It includes northern Bolivia, and is separated from the Chilean segments by the Arica Bend. The Chilean segment (18°S - 56°S) is 4000 km long and trends N-S. Its southern tip exhibits a change to an E-W direction along the Scotia sinistral transform system.

The width of the Andean Chain varies from 250 km in northern Peru (5°S) or southernmost Chile (52°S - 55°S), to as much as 750 km in Bolivia (18°S).

The area described in this chapter includes Ecuador, Peru, Bolivia and, therefore, northernmost Chile (N of 23°S). It includes parts of the three segments described above, the orientation of which played a significant role in the pre-orogenic period.

The lowest pass of the studied area is situated in northern Peru (Abra de Porculla, 2300 m), a zone where the chain is very narrow. The highest average altitude is reached between 15°S and 23°S, where the Altiplano of Bolivia and southern Peru reaches a nearly 4000 m of average elevation, and corresponds to the widest part of the chain. The Andean Chain is usually highly asymmetric, with a steep western slope, and a large and complex eastern side. In Peru, the distance between the trench and the hydrographic divide varies from 240 to 300 km, whereas the distance between the hydrographic divide and the 200 m contour line ranges between 280 km (5°N) and about 1000 km (Lima Transect, 8°S - 12°S). In northern Chile and Argentina (23°S), these distances become 300 km and 500 km, respectively. In southern Peru, as little as 240 km separates the Coropuna Volcano (6425 m) from the Chile-Peru Trench (- 6865 m). This, together with the western location of the Andes relative to the South American Continent, explains why the rivers flowing toward the Pacific Ocean do not exceed 300 km long, whereas those flowing to the Atlantic Ocean reach 4000 km long. We have to note two exceptions to this rule.

In Ecuador this asymmetry disappears, due to peculiar tectonic history and deformational processes. The trench-hydrographic divide distance roughly equals the distance between the water divide and the 200 m contour line, and ranges between 280 and 350 km. Between 13°S and 24°S (southern Peru-Bolivia), there are two hydrographic divides, delimiting a wide, flat, endorheic basin known as the Altiplano, which coincides with the zone of highest average elevation and largest width of the chain (Fig. 1).

The highest summits are usually recent or active volcanoes situated on the deformed chain or on metamorphic or granitic slices uplifted by reverse faults. Among the former are the Cotopaxi (5897 m) and Chimborazo volcanoes (6310 m) in Ecuador, the Coropuna (6425 m) and Ampato volcanoes (6310 m) in Peru, and the Sajama Volcano (6520 m) in Bolivia, near the Bolivia-Chile border. Among the latter, we may note the mainly granitic Cordillera Blanca in Peru, which culminates at 6768 m (Nevado Huascarán), and the metamorphic Eastern Cordillera of southern Peru and northern Bolivia (Nevado Illimani, 6682 m; Nevado Illampu, 6485 m). However, deformed and uplifted sediments may also form high summits in Peru, such as the Nevado Yerupajá (6632 m) in the Cordillera Huayhuash and the Nevado Ausangate (6384 m) in southern Peru or the Cordillera de Apolobamba (Nevado Cololo, 5975 m) in Bolivia.

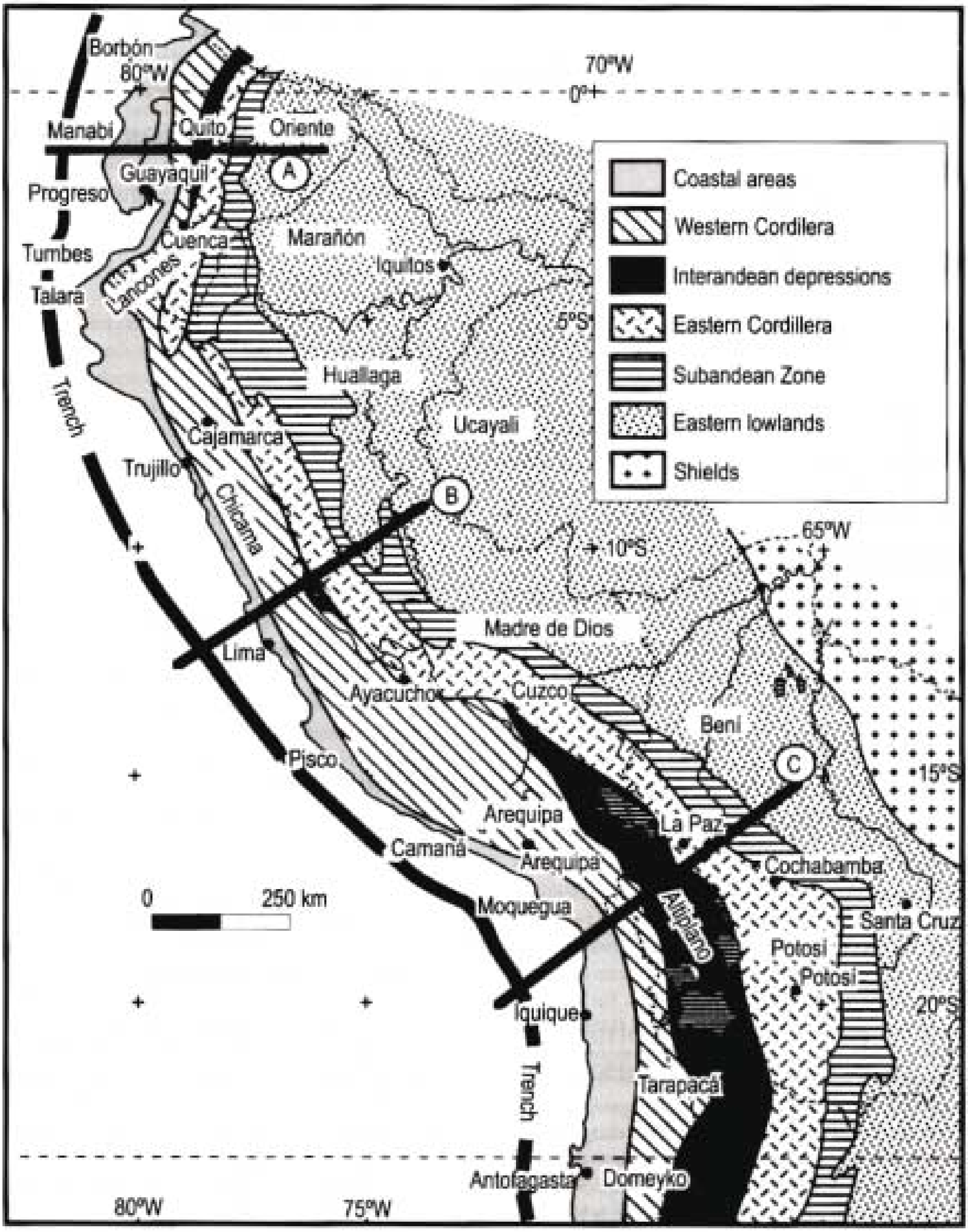
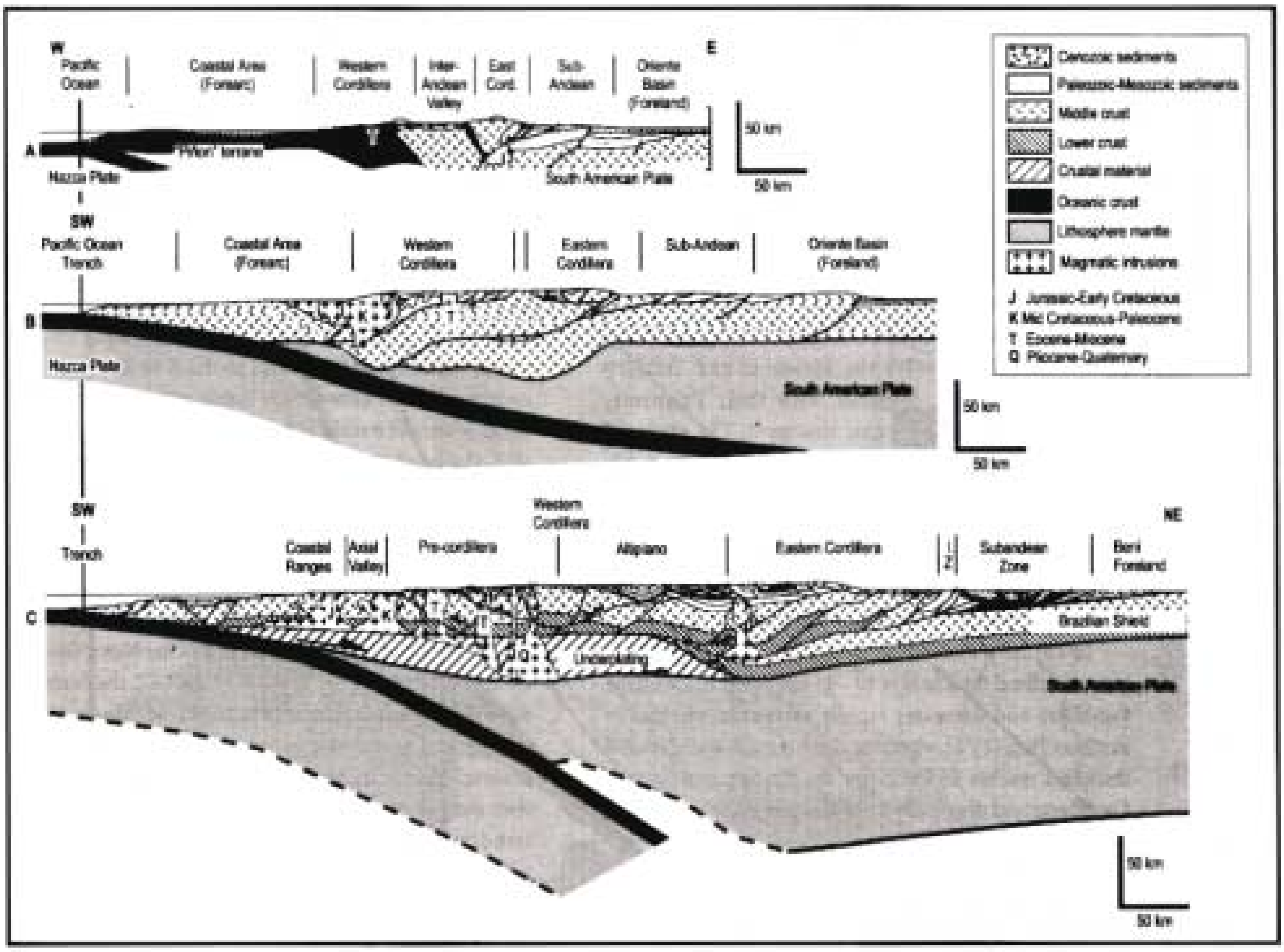


FIGURE 1 - Structural sketch of the Andes of Ecuador, Peru, Bolivia and northern Chile, showing the main morphostructural units. A, B, C is the location of profiles of Fig. 2.

FIGURE 2 - Structural sections of the Andes of A - Ecuador (after Mégard, 1987), B - Central Peru (after Moulin, 1989), and C - Bolivia (after Rochat et al., 1999). Location on Fig. 1.



Morphostructural units and crustal structure

The Ecuadorian segment (1°N - 3°S) exhibits four main longitudinal morpho-structural units (Fig. 1). In the 50 to 180 km wide coastal area relief is low (< 300 m), and sediments are accumulated in rather wide alluvial plains by rivers descending from the Andes. The Andean Chain is made of a Western Cordillera constituted by deformed oceanic rocks, and an Eastern Cordillera (Cordillera Real) made of Paleozoic to Mesozoic metamorphic rocks (Litherland *et al.*, 1994; Fig. 2). The eastern areas are made of a 50 - 80 km wide uplifted zone (500 to 1000 m) made of Mesozoic deformed sedimentary rocks (Subandean Zone), bordered to the E by gently dipping lowlands covered by rainforest, where Quaternary sediments derived from the Andes are presently being deposited.

North of 3°S, the Interandean Valley filled by Tertiary-Quaternary volcanic and volcaniclastic deposits separates the cordilleras. High, active or Recent volcanoes are built on the Cordilleras, the Interandean Valley and the eastern Subandean Zone. South of 3°S, the Interandean Valley, the western oceanic terranes and the active volcanoes disappear, and the Andean Chain is made of metamorphic rocks unconformably overlain by pre-Pliocene volcanic and subordinate sedimentary rocks. The Guayaquil Gulf indentation (3°S) roughly separates the northern fore-arc units (North Andean Block), made of oceanic terranes, from southern fore-arc zones made of continental blocks. In northern Peru (3°S - 5°S), the latter constitutes an isolated NE-SW trending coastal Cordillera (Amotape Massif). This zone corresponds to an important change in the chain orientation (Huancabamba Bend).

Farther S (5°S - 18°S), the NW-SE trending Peruvian margin is basically made of the same units, which exhibit, however, distinct features. Except in northern Peru (Sechura Plain), the coastal zone is narrow and disappears locally in southern Peru. It is made of the deformed edge of the South American Continental Plate, locally covered by fore-arc marine deposits, mainly of Tertiary age. The southern tip of this segment is defined by a sharp direction change from NW to N-S (Arica Bend, Fig. 1).

The Western Cordillera constitutes the continental water divide. It is mainly made of deformed metamorphic rocks and Mesozoic sediments, intruded to the west by Mesozoic plutons (Mesozoic arc) and unconformably covered to the E by abundant volcanic rocks (Tertiary arc), associated with localised Tertiary plutons (Cordillera Blanca). Active volcanoes are only known S of 16°S, thus defining a zone without present-day arc volcanism between 2°30'S and 16°S.

The Eastern Cordillera is chiefly made of pre-Mesozoic metamorphic rocks and subordinate Paleozoic to Tertiary intrusions and volcanic rock (inner arc of southern Peru; Kontak *et al.*, 1985). It is bordered to the W or SW by a nearly continuous E-verging thrust and fold belts (Marañón, Tíclio, Mañazo FTB) involving mainly Mesozoic sedimentary rocks. In southern Peru, the Eastern Cordillera is thrust over the Altiplano by means of SW-verging reverse faults. Longitudinal valleys, the trend of which is controlled by the main Andean structures, separate N of 13°S the Eastern and Western Cordilleras. S of 13°S, the Eastern and Western Cordilleras are more distant and are separated by large valleys (13° - 14°S) evolving southeastwards into an endorheic basin (Altiplano), which considerably enlarges in Bolivia.

The deformed subandean zone is much larger than in Ecuador (120 to 250 km) and enlarges southeastwards (Fig. 2). Deformation involve the Mesozoic and Tertiary sedimentary rocks, together with their Paleozoic, sedimentary or metamorphic basement. The eastward migration of deformation influenced the course of the rivers, which trend mainly parallel to the Andean structures. The eastern lowlands are made of either wide alluvial plains receiving Quaternary sediments (northern Peru), or slightly uplifted, fault-controlled zones underlain by the Brazilian Shield (southern Peru, Bolivia).

In northern Peru (7°S), where the chain is relatively narrow, crustal thickness is 40 - 45 km below the Western Cordillera and decreases rapidly eastwards, whereas in southern Peru (13°S), where the chain is much wider, crustal thickness reaches 65 km below the Western and Eastern Cordilleras and drastically decreases below the Subandean Zone (Fukao *et al.*, 1989).

The northern part of the Chilean-Bolivian segment is more complex. To the W, the 50 km wide Coastal Cordillera is made of chiefly Jurassic-Early Cretaceous magmatic arc rocks, cut by the major N-S trending Atacama wrench fault system. The Longitudinal Valley, underlain by Middle Cretaceous arc rocks separates it from the Precordillera. The 3000 m high Precordillera is made of latest Cretaceous-early Paleogene volcanic and sedimentary rocks overlying Jurassic-Early Cretaceous back-arc sedimentary rocks, and overlain by Neogene-Quaternary tuff and piedmont deposits. To the N of 21°S, it merges with the Western Cordillera, whereas to the S of 21°S, it is separated from the latter by the Preandean Depression occupied by the Salar de Atacama. As a whole, these four desert units are 70 to 250 km wide.

The Western Cordillera is made of Neogene-Quaternary arc rocks, and comprises active volcanoes culminating at more than 6000 m. Isolated outcrops show that Tertiary arc rocks directly overlie Precambrian basement, at least locally. The Western Cordillera forms the Chilean-Bolivian border and limits to the W the 200 km wide, flat Bolivian Altiplano, in-filled by thousands of metres of Tertiary to Quaternary rocks. The latter exhibits an average altitude of more than 3900 m and contains endorheic lakes evolving locally into wide *salars*. The Eastern Cordillera is made of Paleozoic folded metasediments which are thrust to the SW or W onto the Altiplano border, and to the NE or E onto the Subandean Zone. At 19°S, crustal thickness exceeds 60 km and reaches 70 to 75 km below the Western and Eastern Cordilleras (Beck *et al.*, 1996; Fig. 2). The Subandean Zone is made of faulted and folded sedimentary rocks of Paleozoic to Miocene age, which are thrust to the NE or E onto the eastern lowlands. The latter is a large alluvial plain where products of the erosion of the Andes are presently being deposited.

Plate tectonic setting

As first proposed by Wegener early in this century, the Andean Orogeny is related to the convergence between the Pacific Ocean Plate and the South American Continent, the latter being pushed westwards by the opening of the South Atlantic Ocean. We know nowadays that not only South America, but also the Pacific oceanic Plate moves in an absolute reference frame. South America migrates roughly westwards at a rate of 3 cm/y. Between 2°S and 23°S, the Nazca Oceanic Plate migrates to the E to ENE, and the convergence rate between the Nazca and South American plates is around 8 cm/y, although it seems to be lower at 0° (Fig. 3). The whole Andean margin is, therefore, submitted to a roughly E-W convergence between the subducting Nazca Plate and the South American continental margin. This phenomenon is long regarded as the major cause of the Andean Orogeny (Rutland, 1971; James, 1971).

The NE trending Grijalva Fracture Zone (GFZ) located between 3°S and 5°S (Fig. 3) separates the Nazca Oceanic Plate into two portions. To the N of the GFZ, the currently subducting oceanic plate is 23 to 10 Ma old (Miocene) and the depth of the oceanic bottom is 2800 to 3500 m below sea level. Most of the Ecuadorian -Colombian trench is less than 4000 m deep. The Nazca Plate is overlain by the E-W trending Carnegie aseismic ridge, which is presently subducting offshore Ecuador between 0° and 2°S. The Carnegie Ridge is regarded as formed by the Galápagos hotspot, situated about 1000 km W of the Ecuadorian coast.

South of the GFZ, the subducting oceanic plate is 50 to 30 Ma old (Eocene - early Oligocene) and 4000 to 5600 m deep. The Peruvian-Chilean trench reaches depth of 6600 m (11°S, 17°S), 7500 m (19°S) and even 8055 m (23°S). The Nazca Oceanic Plate comprises two submarine highs. The NE-trending Nazca aseismic ridge, entered into the trench 3 to 10 Ma ago and is presently subducted at 15°S - 16°S. It presents summits culminating at -330 m, -841 m and -2059 m, from SW to NE, thus showing a NE plunge of its crestal line. To the S of 18°S, a N-S trending submarine ridge runs alongside the trench of northern Chile, and culminates locally at -1800 m.

Recent improvement in hypocenter relocation at the scale of South America has been done, giving a good definition of the Benioff-Wadati Zone geometry (Engdahl *et al.*, 1998). It shows along-strike segmentation of the Andes but without tears even around the latitude 15°S where the slab has a severe contortion due to changes in its dip. And only a very small portion of the observed seismicity originates within the Andean crust.

From N to S, the Nazca Plate subducts at an angle of 19° - 30° to a depth which varies from 250 km to 650 km. N and S of the Bolivian Orocline, at latitudes 2°S - 15°S and 26°S - 33°S, seismic activity indicates that the Nazca Plate is being subducted nearly horizontal at a depth of about 100 km for a distance of over 300 km before descending steeply into the mantle. There are no volcanic arcs because of the lack of asthenosphere wedge between the subducting flat slab and the upper continental plate (Barazangi and Isacks, 1976). These horizontal segments partially correlate with the oceanic ridges (Nazca Ridge, Juan Fernandez Ridge) or totally

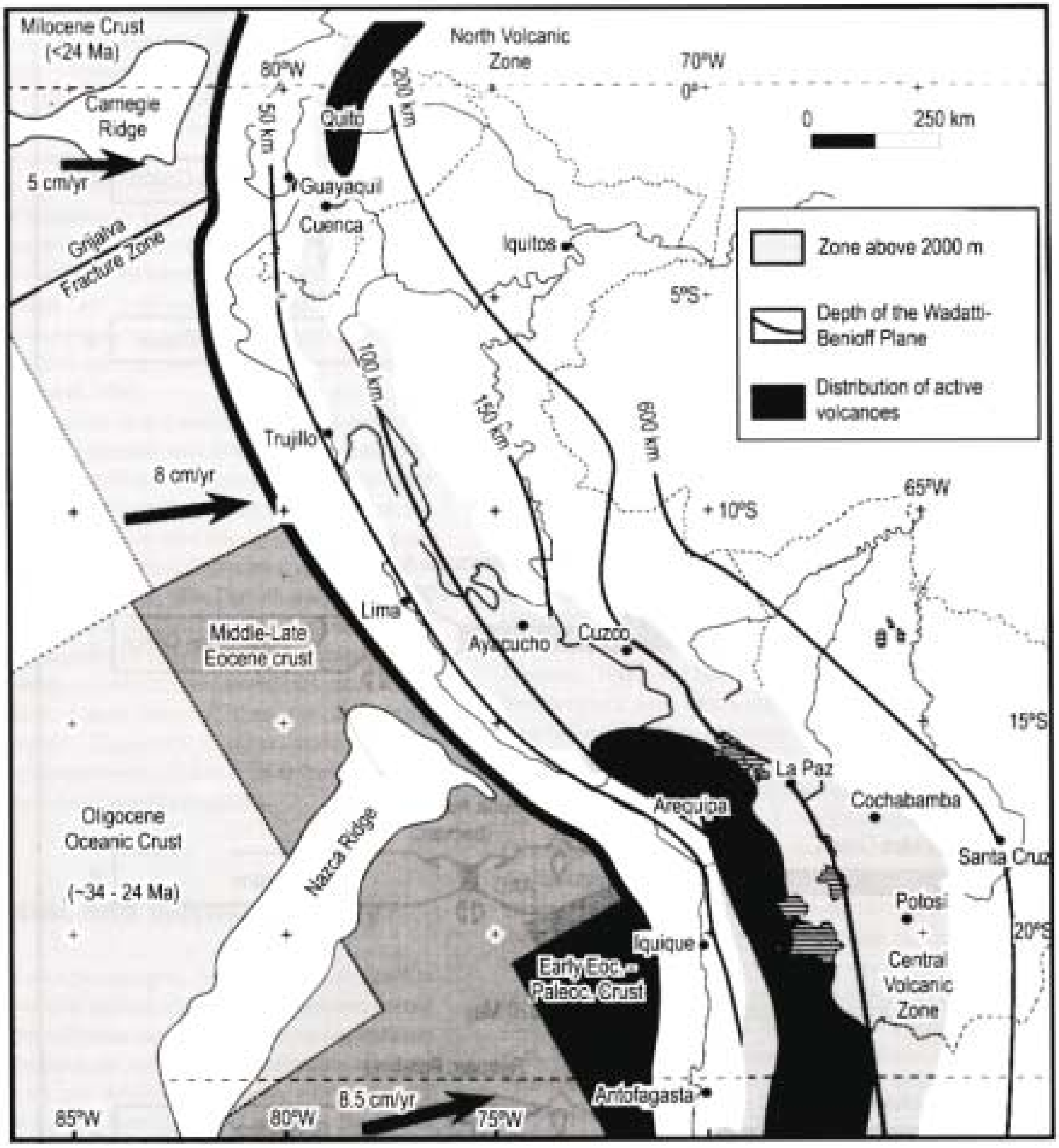


FIGURE 3 - Plate tectonic setting of the Andes of Ecuador, Peru, Bolivia and northern Chile.

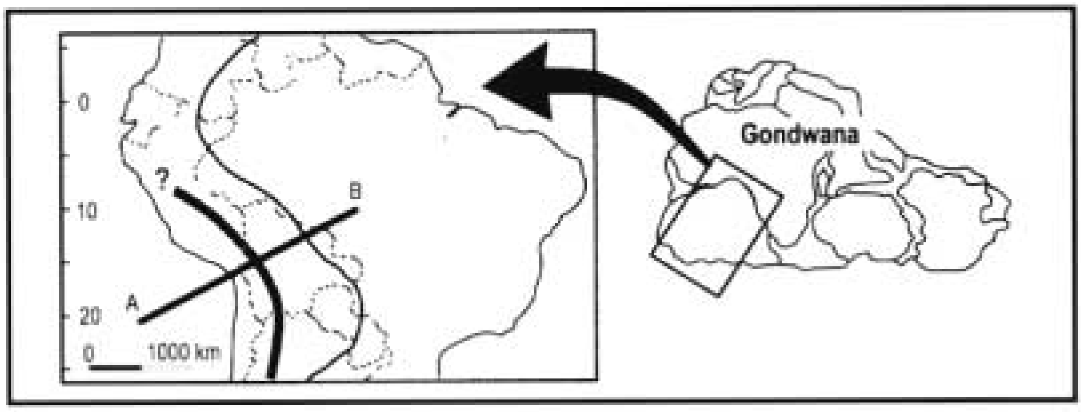


FIGURE 4 - Location of the Central Andes within the Gondwana. Thick grey line - boundary zone between the Arequipa-Antofalla Craton and the Amazonian Craton. Dashed line - eastern boundary of the inferred maximum extent of Paleozoic sedimentation in the "Peru-Bolivia" backarc/retroarc basin. A-B - location of the schematic geodynamic cross-sections of Fig. 5. (all are approximate).

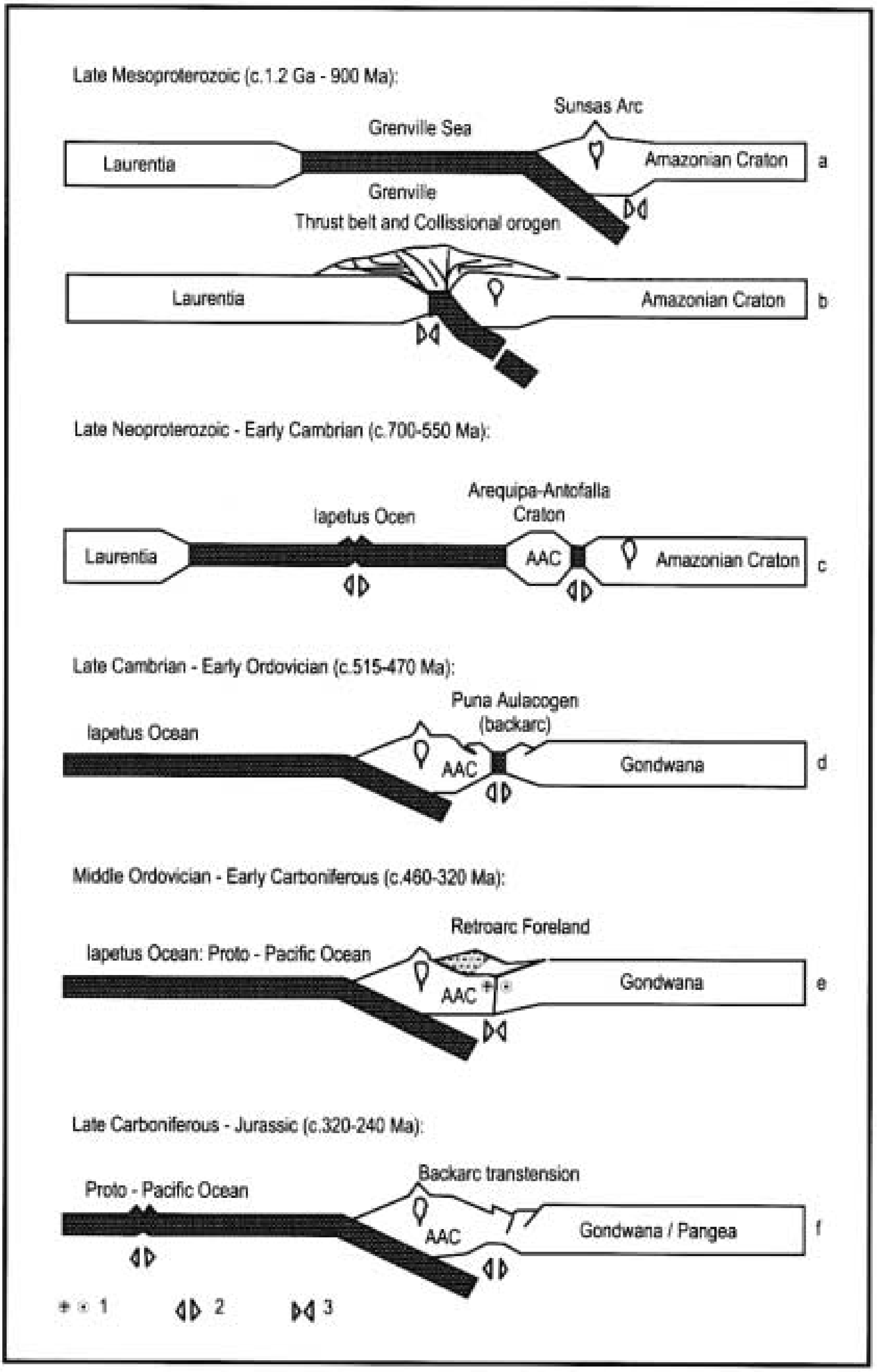


FIGURE 5 - Simplified conceptual model for the pre-Andean geodynamic evolution of the Central Andes (5° - 27°S) as proposed in the text. Approximate location of cross sections in Fig. 4. Overall regional stress field: 1, transcurrent; 2, extensional; 3, compressional.

(Inca Plateau?) subducted (Gutsher *et al.*, 1999). In addition, it is accompanied by important shallow earthquakes (< 20 depth) related to crustal shortening.

The contact zone between the Nazca and South American plates concentrates only a part of the convergence motion as shown by GPS measurements (Kendrick *et al.*, 1999). In the case of the Antofagasta 1995 earthquake, the last great Chilean event (M_w = 8.1), a locked portion of the plate interface was ruptured. The observed slip release (between 2 and 7 m) was less than the expected plate motion accumulation (over 8 m) and suggests that the resultant motion causes permanent deformation through uplift of the Altiplano and crustal shortening in eastern Andes (Norabuena *et al.*, 1998).

Low magnitude earthquakes recorded by local seismic arrays are revealing small-scale differences in seismicity of the Benioff-Wadati Zone and the continental crust (Comte *et al.*, 1999), and should be related with some specific tectonic processes. Body-wave velocity models at the scale of the South American Plate (Bijwaard *et al.*, 1998) and at smaller scale (Comte *et al.*, 1994; Dorbath *et al.*, 1996; Myers *et al.*, 1998) show a medium lateral heterogeneity where the descending oceanic Nazca Plate can be observed over 600 km depth as zones of faster seismic velocities due to the colder, denser material than the underlying asthenosphere. The crustal thickness underneath the Altiplano Plateau is of 60 - 75 km, twice as thick as normal, but its exact origin remains enigmatic.

Volcanic and seismic activity

Like all active margins, the Andes are submitted to intense volcanic activity, which presents, however, several peculiarities. Distribution of the active volcanoes indicates that the volcanic arc activity is not continuous along the Andean margin, defining a Northern, a Central and a Southern Volcanic Zone (NVZ, CVZ and SVZ; Fig. 3), that have been interpreted as related to the dip of the subduction plane. As a matter of fact, the Northern and Central Volcanic Zones overlie segments with relatively steep Wadatti-Benioff Zones, whereas the volcanic gap of southern Ecuador and northern-central Peru corresponds to a zone of flat subduction. In this latter case, the lack of an asthenosphere wedge between the subducting slab and the upper continental plate would prevent arc magma generation (Barazangi and Isacks, 1976). Moreover, the NVZ is characterized by basaltic andesite, whereas andesite, dacite lava and ignimbrite dominate the CVZ. The latter feature is interpreted as due to fractional crystallization during the ascent of the andesite magma through a thickened continental crust.

In addition, this part of the world has already generated great earthquakes of magnitude over 8.2 followed in general by destructive tsunamis. They occurred repeatedly with a certain periodicity of more than 100 years as the greatest instrumentally ever-recorded earthquake located in southern Chile (37°S - 46.5°S). This earthquake occurred on May, 22 1960 (M_w = 9.4, depth = 25 km) and was generated by a 20m slip upon an area of about 1000 x 200 km along the contact zone between the two plates. Over magnitude 8.7, the southern Peru on August 14, 1868 (M = 9.0), the northern Chile on May 9, 1877 (M_w = 9.0) and the Ecuador-Colombia on January 31, 1906 (M_w = 8.8) great earthquakes and their associated tsunamis caused a lot of damage. On the other hand, on June 9, 1994 the largest deep earthquake (M_w = 8.2, depth = 630 km) in recorded history occurred beneath the Andes providing more constraints on the dynamics and geometry of the Nazca Plate at that depth.

Pre-Andean Evolution of the North-Central Andes Domain

This section attempts to provide a coherent overview of the paleogeographic and geodynamic evolution of the Central Andes Domain during the Paleozoic, along with *some comments on its Proterozoic basement*. The synthesis has been compiled from different sources, and also includes some new concepts and interpretations. The area here considered as Central Andes is roughly 5°-27°S, covering part of Peru, Bolivia, northern Chile and northwestern Argentina (Fig. 4), and therefore excludes the Amotapes of northwestern Peru and southern Ecuador, and the Precordillera of western Argentina.

In contrast with the rather complex Phanerozoic accretionary history of the South American margin in the Northern and Southern Andes (Colombia-Ecuador and Chile-Argentina, respectively), the evolution of the Central Andes appears to be somewhat simpler, as there is no evidence for the accretion of allochthonous terranes during the Phanerozoic. Recent studies suggest that the crustal basement in most of the Central Andean area formed part of the Grenville Orogen, as a result of the collision between Laurentia and Amazonia in the Mesoproterozoic (Wasteneys *et al.*, 1995; Sadowsky and Bettencourt, 1996; Tosdal, 1996). Paleozoic rocks of the Central Andes record the break-up of the Neoproterozoic Proto-Pangea (Rodinia) in the latest Neoproterozoic-Cambrian to form a passive margin along western Gondwana (Bond *et al.*, 1984; Powell *et al.*, 1993), and its later evolution as an active margin during most of the Paleozoic and until present times (Sempere, 1995). The continuous superposition of magmatic, tectonic and sedimentary events has led to complex lateral variations, both in cross section and along strike, and has originated a wide range of settings for the development of mineral deposits (Fontboté *et al.*, 1990; Schneider, 1990; Fornari and Hérail, 1991) and hydrocarbon generation and accumulation (Gohrbandt, 1992; Moretti *et al.*, 1995; Tankard *et al.*, 1995) related with Paleozoic rocks.

Proterozoic basement

Due to the protracted superposition of orogenic events in the Central Andes, the interpretation of the Proterozoic evolution of the region is inevitably very fragmentary, and should be considered as a mere working hypothesis. The modern basement of the Central Andes consists of two crustal blocks with different origins: the Arequipa-Antofalla Craton

(Ramos *et al.*, 1986) and the Amazonian Craton (Teixeira *et al.*, 1989) or Central Brazil Shield (figs. 1 and 2). The Arequipa-Antofalla Craton is a Proterozoic terrane interpreted to have originated as the tip of a pre-Grenville Laurentian promontory (comprising Labrador, Greenland and Scotland) that was incorporated into the Grenville Orogen (Dalziel, 1994; Wasteneys *et al.*, 1995). Pb isotope composition seems to contradict this model, indicating instead closer ties with the Amazonian Craton (Tosdal, 1996). The reconstruction of the remnants of the Grenville Orogen in South America (Sadowski and Bettencourt, 1996) indicates that the Central Andes corresponds to an area intermediate between the magmatic arc (represented by the Sunsas igneous province, in eastern Bolivia and western Brazil) and the thrust belt (southeastern Canada) of the Grenville Orogen (Fig. 5b), and explains the similar trends identified between the Proterozoic outcrops along the Andes, and those of the Brazilian Shield (Litherland *et al.*, 1985, 1989).

Paleoproterozoic ages indicated by Rb/Sr whole-rock isochrons and bulk U/Pb zircon geochronology represent the pre-Grenville Laurentian-Amazonian protolith, and Mesoproterozoic ages of granulite-facies metamorphism indicated by U/Pb single-grain zircon geochronology represent the main collisional events of the Grenville Orogen (Wasteneys *et al.*, 1995; Sadowski and Bettencourt, 1996; Tosdal, 1996). Rifting during break-up of Rodinia in the Neoproterozoic-Cambrian led to separation of Laurentia from Amazonia (Fig. 5-c), leaving behind the parautochthonous Arequipa-Antofalla Craton attached to Amazonia (Central Brazil Shield). The boundary zone between the two crustal blocks, which is located beneath the Eastern Altiplano and Eastern Cordillera, thus constitutes a paleosuture and crustal weakness zone inherited from the Mesoproterozoic evolution of the Grenville Orogen (Figs. 1 and 2). This zone remained active during the Paleozoic, and ever since, with variable behaviour depending on the regional state of stresses (Ramos, 1988; Dorbath *et al.*, 1993; Forsythe *et al.*, 1993).

Tectonomagmatic episodes

Tectonic and magmatic events took place in the Central Andes in a rather continuous manner during the whole Paleozoic, shifting their foci and areal extent with time as a result of plate interactions, variable geometry of the plates involved, and the resulting stress regimes. These variable conditions led to apparently different styles of evolution depending on the local area under study. However, an overall tendency for the whole region may be discerned and simplified as follows. Apart from the aforementioned Mesoproterozoic (1.2 Ga - 900 Ma) ages resulting from the Grenville granulite-facies metamorphism, and the Paleoproterozoic (2.0 -1.9 Ga) ages obtained from its metagranitoid protolith, the crystalline basement underlying the Central Andes also presents Neoproterozoic to Middle Cambrian (600 - 520 Ma) ages of igneous and metamorphic events. These events are commonly assigned to the last phases of the Brasiliano Orogeny, which are referred to as Pampean Orogeny in the southern Central Andes (Rapela *et al.*, 1998). Break-up of Rodinia and rifting of eastern Laurentia from Gondwana in the Neoproterozoic and Early Cambrian led to the development of passive margins on both sides of the Southern Iapetus Ocean (Fig. 5-c).

Along the pre-Andean margin of Gondwana S of 27°S, a westward shift of the spreading centre is interpreted to have left oceanic crust between a detached continental block (Pampean Terrane) and the Gondwana margin (Rapela *et al.*, 1998). Later eastward subduction and closure of this remnant sea during the Early Cambrian led to the collision of the Pampean Terrane in the Middle Cambrian. To the N of 27°S, a similar history is also probable, with the Arequipa-Antofalla Craton also partially rifting and then later colliding along its southeastern boundary with the Amazonian Craton. Meanwhile, the western margin ofboth the Pampean Terrane and the Arequipa-Antofalla Craton remained passive until the Late Cambrian (Fig. 5-c).

Beginning in the Late Cambrian or earliest Ordovician, this western passive margin became an active continental margin (Fig. 5-d). The San Nicolás Batholith in southwestern Peru is interpreted as the root of the magmatic arc resulting from eastward subduction of oceanic crust along the active margin of Gondwana during the Paleozoic (Mukasa and Henry, 1990). Ordovician-Devonian ages obtained for lower intercepts in U/Pb geochronology of basement rocks along the western Arequipa-Antofalla Craton reflect thermal overprinting and Pb-loss coinciding with peaks of this Paleozoic magmatic activity (Shackleton *et al.*, 1979; Damm *et al.*, 1990; 1994; Mukasa and Henry, 1990; Tosdal, 1996). Different rates of plate activity and sense of migration of the magmatic arc developed depending on regional stresses and inhomogeneities, and basin development also changed in accordance with these plate interactions.

The Paleozoic development of this active margin is characterized by back arc extensional conditions during the early phase (latest Cambrian-Early Ordovician) and late phase (Late Carboniferous-Permian), resulting in a strongly subsiding thinned crust, with partial rifting and syn-sedimentary basic volcanism (Fig. 5d). In contrast, a compressional regime (retro-arc foreland) characterized the intermediate phase (Middle Ordovician-Early Carboniferous; Fig. 5e). An apparent lack of evidence for Silurian and Devonian tectonomagmatic activity in the southwestern Central Andes has been interpreted as evidence for a passive margin resulting from rifting off of part of the Arequipa-Antofalla Craton (Bahlburg and Hervé, 1997). However, this interpretation is difficult to reconcile with the evidence for a coeval active plate margin in southern Peru (down to 17°S) and northern Chile and Argentina (up to 27°S), as well as with the Silurian age of igneous and metamorphic events in the same region (Damm *et al.*, 1990, 1991, 1994).

With regard to Late Paleozoic tectonomagmatic events in the Central Andes, these have been traditionally assigned to a "Eohercynian" Orogeny (Mégard *et al.*, 1971; Bard *et al.*, 1974; Dalmayrac *et al.*, 1980), including Late Devonian-Carboniferous K/Ar and Rb/Sr ages, and regional stratigraphic and petrographic evidence. However, only one U/Pb zircon age is reported (330 ± 10 Ma; Carlier *et al.*, 1982). Other U/Pb zircon dates on granitoid along the NW trending segment of the Eastern Cordillera of Peru and

Bolivia establish their time of emplacement as Permian or younger (McBride *et al.*, 1983; Heinrich *et al.*, 1988; Farrar *et al.*, 1990; Kontak *et al.*, 1990).

This more recent evidence questions the relation of the granitoid plutons with widespread Late Devonian-Early Carboniferous "Eohercynian" magmatism and orogenesis, as previously interpreted. Nevertheless, there is evidence for local transpressional uplift and deformation of the Eastern Cordillera in the latest Devonian and Early Carboniferous, which separated an Altiplano basin from a Subandean-Chaco basin (Mégard *et al.*, 1971; Dalmayrac *et al.*, 1980; Kley and Reinhardt, 1994; Sempere, 1995; Díaz-Martínez, 1996). Maximum burial depths (locally exceeding 10 km) were attained in different areas of the Central Andean Paleozoic basin at different times during the Late Paleozoic (Late Devonian-Permian interval). As a result of this deep burial, and probably in conjunction with transcurrent stresses along the aforementioned suture zone (Sempere, 1995), the stratigraphically lower units (Ordovician-Silurian) reached very low-grade to low-grade metamorphism, as indicated by vitrinite reflectance and illite crystallinity (Kley and Reinhardt, 1994). This thermal overprint resulted in the reset of K/Ar system, thus explaining the Carboniferous-Early Permian ages obtained by some authors (Dalmayrac *et al.*, 1980; Paton, 1990).

At the same time, erosion related with the mid-Carboniferous global regression resulted in a disconformity or low-angle unconformity throughout the Central Andes, with Upper Carboniferous and Permian units directly overlying Devonian, Silurian or Ordovician units (Kley and Reinhardt, 1994; Isaacson and Díaz-Martínez, 1995; Díaz-Martínez, 1996, 1998b). A similar unconformity is observed in the southern Central Andes (northern Chile and northwestern Argentina), where the development of an active margin with related fore-arc, intra-arc and back-arc basins took place during the Late Carboniferous and Permian (Breitkreuz *et al.*, 1988, 1989; Bahlburg and Breitkreuz, 1991; Breitkreuz and Zeil, 1994). Thus, the reinterpretation of the evidence indicates that the alleged "Eohercynian" Orogeny may be no more than the conjunction of different processes and events taking place during the Late Paleozoic throughout the Central Andean region, but not a single tectonomagmatic event or belt localized in space and time.

The sedimentary record in Peru and western Bolivia presents evidence of marginal arc volcanism beginning in the late Early Carboniferous (Ambo Group, 6°S - 17°S; Dalmayrac *et al.*, 1980; Díaz-Martínez, 1995, 1998b). This subduction-related magmatic activity propagated to the S (16°S - 24°S) along the active margin during the Late Carboniferous and into the Permian (Carlier *et al.*, 1982; Bell, 1987; Breitkreuz *et al.*, 1988, 1989; Bahlburg and Breitkreuz, 1991; Breitkreuz and Zeil, 1994; Sempere, 1995; Isaacson and Díaz-Martínez, 1995; Bahlburg and Hervé, 1997). Farther to the S (20°S - 42°S), thick volcano-sedimentary successions developed in the Permian and Triassic (Choiyoi Group), with associated high-level intrusions (Kay *et al.*, 1989).

Further evidence for Late Paleozoic magmatic activity is also found in the Eastern Cordillera of Peru and Bolivia. This activity began earlier in the Peruvian sector, with Late Permian-Triassic ages (Kontak *et al.*, 1990), and migrated towards the S, developing in Bolivia during the Middle Triassic to Early Jurassic (McBride *et al.*, 1983; Heinrich *et al.*, 1988). The activity was not related to subduction processes, but instead, consists of alkaline volcanism and plutonism which are interpreted as a result of partial rifting and transtension along the suture zone between the Arequipa-Antofalla Craton and the Amazonian Craton (Fig. 5f), due to regional stresses during Pangea break-up (Kontak *et al.*, 1990; Atherton and Petford, 1991).

Tectono-sedimentary cycles

Most of the Late Cambrian to Devonian sedimentation in the Central Andes took place along a wide and elongated epicontinental marine basin broadly parallel to the margin of Gondwana (Fig. 4). The main source area for this basin was situated to the W and SW, although this source area was not a rifted-off former "Pacific continent" (as considered by Bahlburg and Hervé, 1997). Instead, it is considered that the source area was the active margin of Gondwana itself (Isaacson and Díaz-Martínez, 1995; Sempere, 1995), influenced by its discontinuous activity, specially the uplift and forward migration of the fold-thrust belt. Deepening of the basin and increased subsidence rates broadly coincide in time with the aforementioned Ordovician-Devonian peaks of magmatic activity, and demonstrate the syn-tectonic character of deposition, in close relation with tectonic piling and uplift along this fold-thrust belt (Sempere, 1995; Díaz-Martínez *et al.*, 1996).

The Paleozoic marginal orogen which fed the basin is today completely dismantled, with its roots cropping out in the Arequipa Massif and other smaller outcrops scattered throughout the Arequipa-Antofalla Craton. Extreme burial depths locally exceeding 10 km were achieved during the Late Paleozoic, leading to the reset of isotope ages during their maxima, and previous to the widespread erosion of depocenter areas. Paleozoic basin fills were later variably eroded, and the resulting stratigraphic gaps are wider at the margins of the basin, especially towards the western uplifted areas. The original width of the Paleozoic orogen seems narrower today due to tectonic erosion along the western margin of the Arequipa-Antofalla Craton (Stern, 1991), as well as compressional shortening due to Cenozoic folding and thrust-piling. Both processes (tectonic erosion and dismantling of the orogen and basin fill) continued during Mesozoic and Cenozoic subduction, uplift and erosion in the Central Andes, and resulted in the scarcity of evidence for the more marginal (fore arc and intra arc) and older basins.

Paleozoic sedimentation in the Central Andes may be subdivided into three main tectono-sedimentary cycles (limited by major unconformities): latest Proterozoic-Middle Cambrian, Late Cambrian-Early Carboniferous, and Late Carboniferous-Early Triassic. The first tectono-sedimentary cycle equates to the Pampean Cycle of Aceñolaza *et al.* (1990). The second tectono-sedimentary cycle equates to the Tacsarian and Cordilleran cycles of Suárez-Soruco (1989), and to the Tacsara, Chuquisaca and Villamontes supersequences of Sempere (1995). The third tectono-sedimentary cycle equates to the Cuevo supersequence of Sempere (1995). Overall, Paleozoic sedimentation was rather continuous, although remarkable changes in subsidence rates took place during certain time intervals. Sedimentary evidence for increased tectonic instability and uplift occur near the Ordovician-Silurian and Devonian-Carboniferous boundaries (Díaz-Martínez *et al.*, 1996), which have been traditionally used for further subdivision of the sedimentary pile (Suárez-Soruco, 1989; Gohrbandt, 1992; Sempere, 1995). However, as mentioned above, tectonomagmatic processes along the margin have been rather continuous and synchronous with the development of basins. Hence, it does not seem appropriate to maintain the traditional subdivisions of the Paleozoic sedimentary record based on discrete orogenic events.

Paleogeography and paleoclimates

Deposition within a large marine basin (Peru-Bolivia Basin) located adjacent to a marginal orogen prevailed during most of the Paleozoic (Figs. 1 and 2). The overall and progressive increase of rigidity and thickness of the crust beneath the Central Andes in the Paleozoic, led to a gradual increase of terrestrial (subaerial and lacustrine) sedimentation in the Late Paleozoic. Fluvial deposits are very rare until the Devonian, and become frequent in the Carboniferous. Eolian deposits begin to be present in the Late Carboniferous and are frequent until the Jurassic. The areal extent of marine deposition in the Late Permian and Triassic was very limited.

The Central Andes underwent important latitudinal movements during its drift as part of Gondwana's western margin in the Paleozoic. Despite the scarcity of confident paleomagnetic data, several models have been proposed for the Paleozoic paleogeographic evolution of the region. The overall trend consists of a subtropical (30°) latitudinal position during the Early Cambrian, and a shift towards higher latitudes in the Middle and Late Cambrian (Courjault-Radé *et al.*, 1992). During the Ordovician, Silurian and Devonian, the Central Andes remained at mid to high latitudes, with variable shifts. A gradual shift towards lower latitudes took place during the Early Carboniferous, and the area has remained at tropical latitudes ever since the Late Carboniferous (Díaz-Martínez *et al.*, 1993; Isaacson and Díaz-Martínez, 1995; Sempere, 1995; López-Gamundí and Breitkreuz, 1997). Paleozoic tropical carbonates and evaporites in the Central Andes are present only within Early-Middle Cambrian deposits and Late Carboniferous-Permian deposits. Thin carbonate-bearing units are present in the Ordovician, Silurian and Devonian systems containing cool-water faunal associations. Plant remains are frequent in the sedimentary record beginning in the Middle Silurian, and coal development is observed in the late Early Carboniferous (Ambo Group) and in the Late Carboniferous-Early Permian (Copacabana Formation). Glacially influenced deposits are preserved within latest Ordovician-Early Silurian units, and within Late Devonian-Early Carboniferous units. However, true tillites and glacially striated pavements are only very rarely found (Starck, 1995), and most of the evidence consists of glacially faceted and striated clasts, both as dropstones and within resedimented units (Díaz-Martínez *et al.*, 1999). These glacially-related deposits are interpreted as a result of glaciation of tectonically uplifted highlands, and therefore unrelated with the large continental icecaps in Gondwana.

Inherited pre-Andean structures

Structures inherited from the pre-Andean geodynamic evolution of the Central Andes have exerted an important influence on its later development. The crust beneath the Central Andes originally formed as part of Proterozoic orogens. The trend of the structures developed during the formation of these orogens greatly conditioned later Paleozoic basin formation and crustal weakness zones, and therefore influenced the geometry and distribution of tectonism and deposition. The boundary zone between the Arequipa-Antofalla Craton and the Amazonian Craton is the principal of these features (Fig. 4), probably inherited as a suture zone from the collision between Laurentia and the Amazonian Craton as part of the Grenville Orogen (Fig. 5). This crustal weakness zone was the location of (a) rifting in the Neoproterozoic and Cambrian, (b) back arc and foreland successor basin formation from the Ordovician to Carboniferous, (c) syn-sedimentary magmatism in the Ordovician and Silurian, (d) transpressive stresses originating local uplifts in the Devonian and Early Carboniferous, and (e) syn-sedimentary magmatism, and transtensional stresses originating semigrabens and grabens in the Late Carboniferous to Jurassic. In turn, the resulting Paleozoic features also influenced later events. For instance, (a) Paleozoic basin geometry and facies distribution greatly conditioned the location of decollement levels and lateral variations in the propagation of thrusts during Cenozoic tectonism (Baby *et al.*, 1989, 1992; Sempere *et al.*, 1991), and (b) listric faults originated during Late Paleozoic-Early Mesozoic extensional conditions were reversed during the Cenozoic. The same boundary zone was active with magmatism and strong tectonism during the Cenozoic, resulting in the formation of the Eastern Cordillera and Subandean fold-thrust belts.

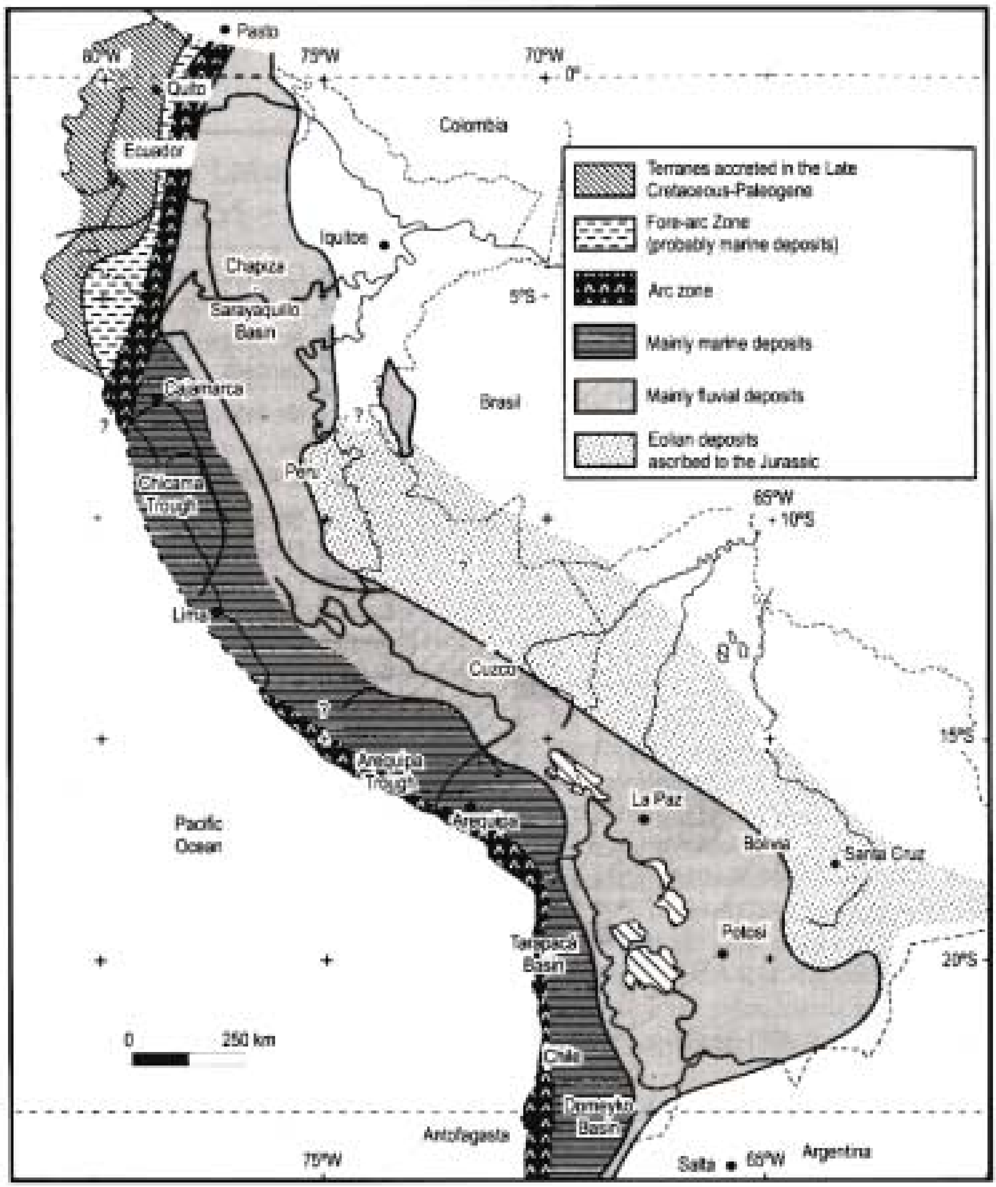


FIGURE 6 - Paleogeographic framework of the Andes of Ecuador, Peru, Bolivia and northern Chile for the Jurassic. (not all features showed are of the same age).

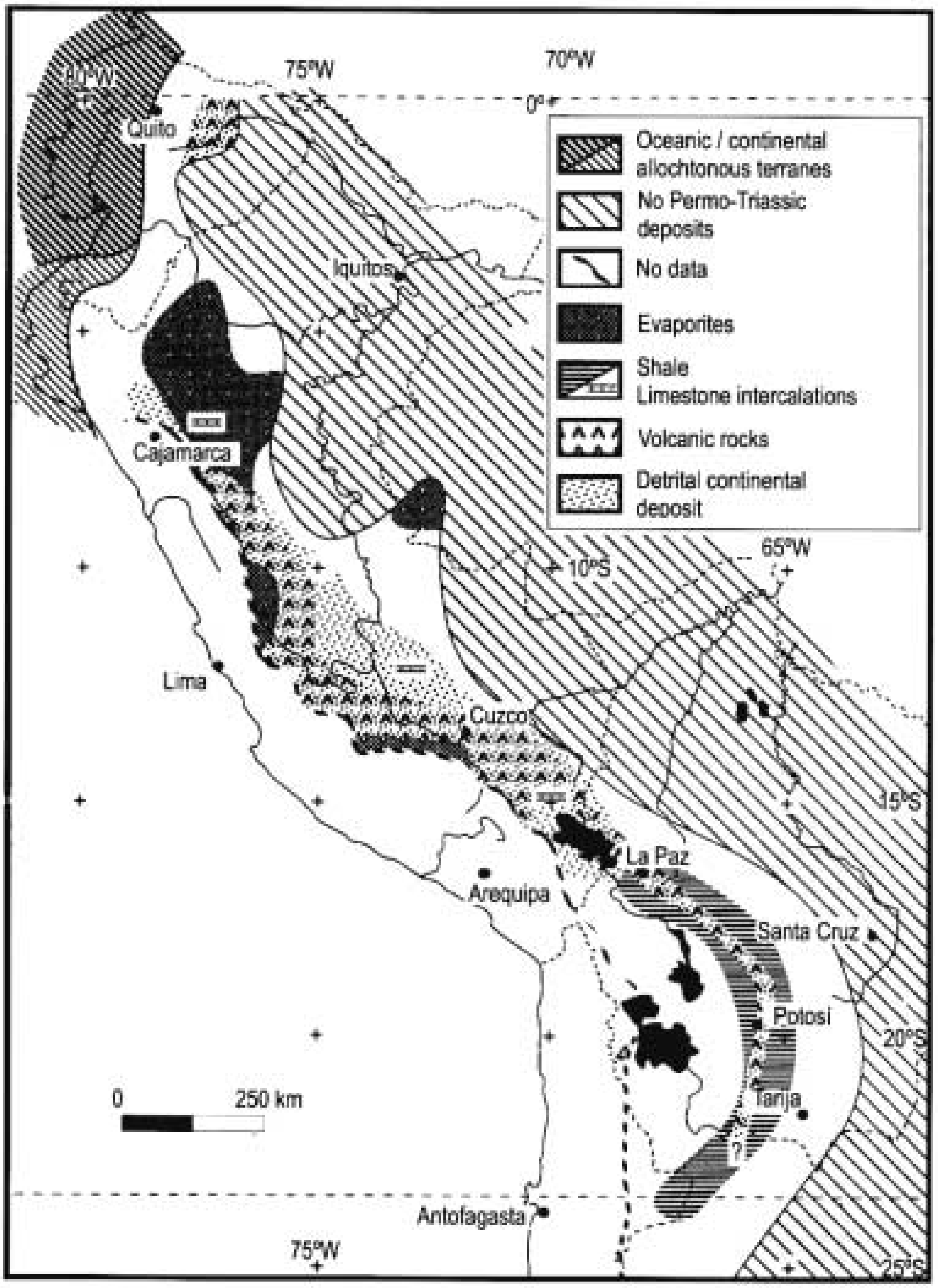


FIGURE 7 - Paleogeographic sketch of Ecuador, Peru, Bolivia and northern Chile for the Late Permian-Triassic period (after Dalmayrac et al., 1980).

Pre-Orogenic (Late Permian-middle Oligocene) Evolution of the North-Central Andes

This period may be divided into two main sub-periods, of Late Permian-Late Jurassic and latest Jurassic-Paleogene age, corresponding to different geodynamic, paleogeographic and tectonic settings.

Late Permian to Late Kimmeridgian (255 - 140 Ma)

Prior to the Sinemurian, the magmatic arc is not well-defined. During Jurassic, the paleogeographic framework exhibits the classical three-fold division of an active margin (Fig. 6). The fore-arc realm, is nearly unknown, because of subsequent tectonic erosion and/or deformation. It was located in the present-day Eastern Cordillera of Ecuador (Cordillera Real), whereas it probably lies offshore Peru and Chile. The magmatic arc controlled the paleogeographic pattern. It trends NNE in the northern segment; runs along the present-day Subandean Zone of Ecuador and crosscut obliquely the northern Peruvian margin. Farther S, it appears locally on the SE trending coast of southern Peru, and extends into northern Chile, where its present-day orientation is roughly N-S. The back-arc domain covers what are now the Oriente Basin of Ecuador, the Western Cordillera of Peru and most of the coast of northernmost Chile. Distal back-arc areas covering eastern Peru and Bolivia (Fig. 6) bordered it to the E.

Late Permian - Middle Sinemurian (255 - 195 Ma)

This period is characterized by Late Permian-Triassic extensional tectonics along NW-SE trending grabens, and by Late Triassic-Early Liassic marine transgressions. Coeval alkaline volcanism and intrusions due to partial melting of the lower crust were locally associated with deep-seated metamorphism.

Late Permian - Early Norian (255 - 215 Ma)

Very little is known about the westernmost areas of the margin. Coastal areas are, however, affected by significant although poorly understood deformation and metamorphism. In the Eastern Cordillera of Ecuador, type I granitoid (Tres Lagunas) yielded ages ranging from 257 (Sm/Nd) to 200 - 189 Ma (Rb/Sr, Aspden and Litherland, 1992; Litherland *et al.*, 1994), and in southwestern Ecuador orthogneiss yielded metamorphic ages of 234 to 198 Ma (K/Ar, Feininger and Silberman, 1982; Rb/Sr, Aspden *et al.*, 1995). Intrusion and metamorphism are interpreted as due to a strong thermal event of Late Triassic age, resulting in partial crustal melting and high temperature metamorphism. They would be associated with significant dextral movements related to the Tethyan oblique rifting (Litherland *et al.*,, 1994).

In northern and central Peru, no pre-Late Jurassic arc volcanic rocks are known. In the coast of southern Peru (Arequipa Massif), metamorphic and intrusive rocks yielded K/Ar ages ranging from 213 to 187 Ma (Stewart *et al.*, 1974; Romeuf *et al.*, 1993). In southern Peru, a Triassic-Liassic age is assumed by Boily *et al.* (1984) for the arc volcanism, since it is post-dated by, and locally intercalated with, Sinemurian marine deposits (Vargas, 1970; Jaillard *et al.*, 1990). However, one of the plutons intruding these volcanic rocks yielded a 211 Ma age, thus indicating a pre-Liassic minimum age (Romeuf *et al.*, 1993).

In northern Chile (26 - 31°S), basic to granitic, I-type intrusions yielded ages between 231 and 201 Ma (Berg and Baumann, 1985; Irwin *et al.*, 1988; Gana, 1991), and are roughly coeval with syn-metamorphic deformation dated at 220 - 201 Ma (Irwin *et al.*, 1987). Farther to the E, in Peru, thick sequences of fine- to coarse-grained, red sandstone and conglomerate beds (Mitu Group) were deposited in a subaerial environment, and filled fault controlled grabens (Mégard, 1978; Kontak *et al.*, 1985; Carlotto, 1998). The associated volcanic rocks are alkaline basalt with subordinate tholeiitic basalt, and acidic, dacitic to rhyolitic pyroclastites (Fig. 7). Shoshonitic and peralkaline suites (Kontak *et al.*, 1985) as well as comendite (Noble *et al.*, 1978) are present, indicating an intraplate extensional tectonic regime. Because they overlie fusulinid-bearing limestone of Early Permian age, the volcanic and sedimentary rocks of the Mitu Group have been ascribed to the Late Permian-Triassic (Newell *et al.*, 1953; Laubacher, 1978). Farther to the N (Ecuador), this period corresponds to non-marine clastics and subordinate volcanics (Sacha Formation, Rivadeneira and Sánchez, 1991). Deposits similar to those of the Mitu Group have been recently identified in Bolivia, where they are found in major paleogeographic structures, which governed the Andean tectonics (Sempere *et al.*, 1999; Fig. 7). In northern Chile, deep transtensional rift basins were filled by thick sequences of subaerial basic, intermediate and acid volcanic rocks and subaerial siliciclastic and volcaniclastic rocks of Late Permian-Triassic age (Chong, 1976; Suarez and Bell, 1991; Flint *et al.*, 1993).

The Late Permian-Early Triassic age has been confirmed in southern Peru by 270 - 200 Ma ages obtained from intrusions and shoshonite, which, however, extent into the Early Jurassic (200 - 180 Ma, Kontak *et al.*, 1985; Clark *et al.*, 1990). In southern Peru, coeval biotite granodiorite and monzogranite derived from partial melting of the lower crust (Granitoid province) are crosscut by mafic dykes showing chemistry and mineralogy similar to that of the alkali basalt of the Mitu Group (Kontak *et al.*, 1985). Farther the SE, most of the granitic to granodioritic, peraluminous intrusions of the Bolivian Cordillera Real yielded ages from 225 to 195 Ma (Ávila-Salinas, 1990). These sedimentary, tectonic and magmatic phenomena were interpreted as resulting from a NW-SE trending rifting of Late Permian-Triassic age, related to the break-up of Gondwana and/or to the Tethyan rifting (Vivier *et al.*, 1976; Kontak *et al.*, 1985; Jaillard *et al.*, 1990; Flint *et al.*, 1993; Sempere *et al.*, 1998). No clear evidences of subduction-related volcanism have been found as yet in these areas.

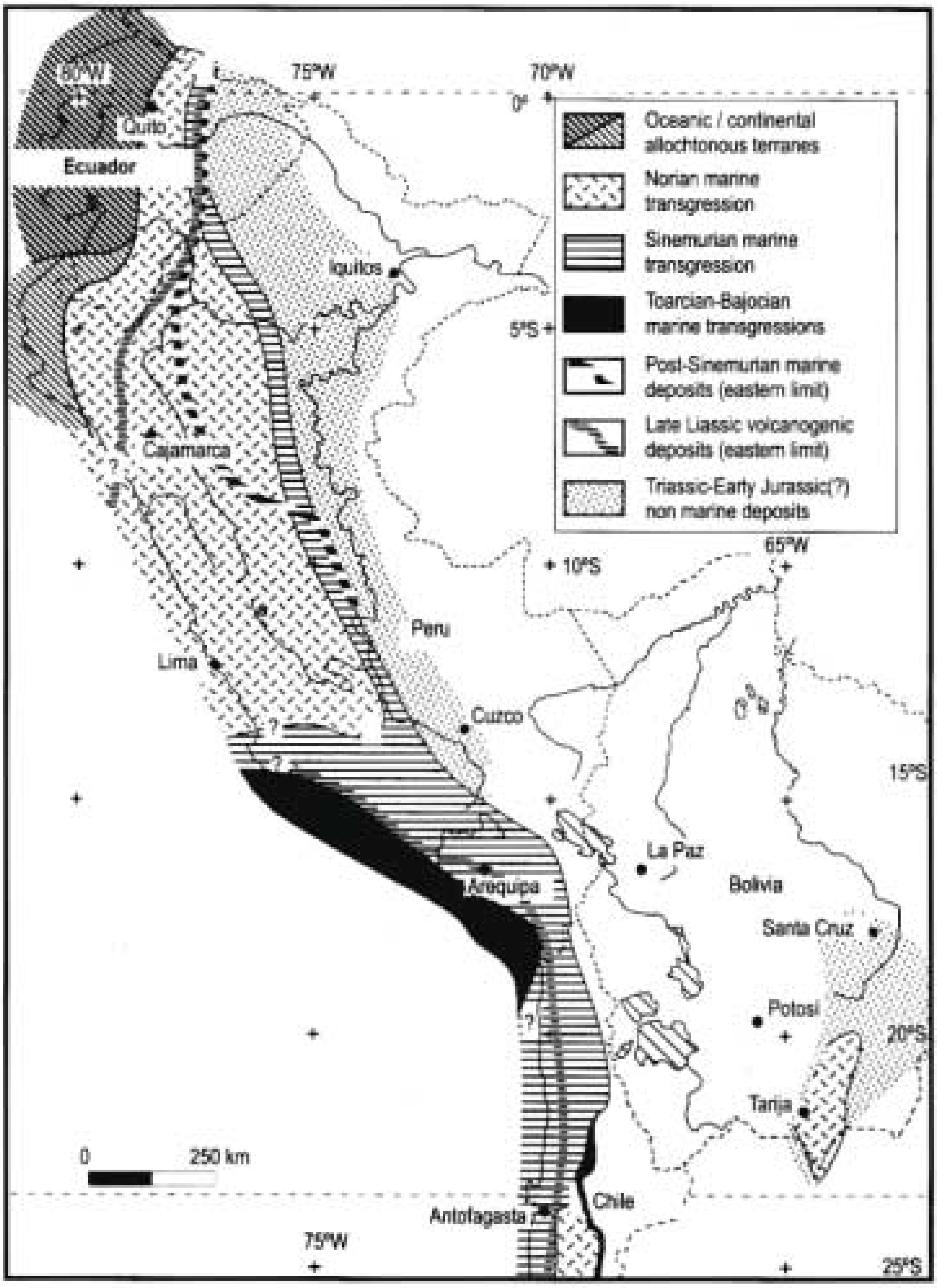
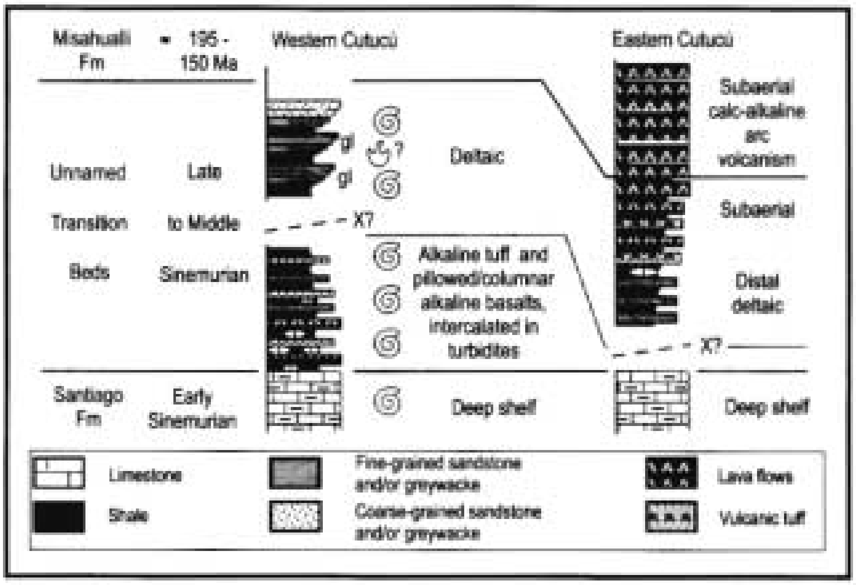


FIGURE 8 - Paleogeographic sketch of the Late Triassic and Early Jurassic marine transgressions.

FIGURE 9 - Transition between the marine shelf limestones (Santiago Fm) intercalated with alkaline lava flows and the subaerial continental arc volcanism (Misahualli Fm) in the southern Subandean zone of Ecuador (Cutucú Cordillera), during the Sinemurian (after Romeuf et al., 1997).



Late(?) Norian - Middle Sinemurian (215 - 195 Ma)

Two main marine transgressions and a relative tectonic quiescence characterize this time-span. The Late Norian marine transgression reached the western part of the Andean margin from southern Colombia to northern Chile (Fig. 8). Norian deposits usually overlie Permian-Triassic subaerial volcanic and sedimentary rocks, locally in angular unconformity, or may overlie the Paleozoic basement. In the central and northern parts of the basin (Ecuador, Peru), Norian deposits consist of shallow-marine shelf limestone and dolomite (Mégard, 1978; Pardo and Sanz, 1979; Loughman and Hallam, 1982; Prinz, 1985; Rivadeneira and Sánchez, 1991; Rosas *et al.*, 1997), whereas in the eastern (eastern Peru, Mégard, 1978) and southern areas (southern Bolivia, northern Chile, Gröschke *et al.*, 1988; Suárez and Dalenz, 1993; Ardill *et al.*, 1998), they begin with a transgressive clastic, locally evaporitic sequence, evolving toward marine deposits.

A stratigraphic hiatus of latest Norian-Rhetian age occurred locally in Peru (Mégard, 1978; Loughman and Hallam, 1982; Prinz, 1985), which may explain local disconformities (Pardo and Sanz, 1979). The second major marine transgression is of Late Hettangian to Early Sinemurian age (Fig. 8). In Peru, these transgressive deposits onlap onto the Permian-Triassic volcanogenic rocks (Mégard, 1978; Loughman and Hallam, 1982). In the central part of the basin, they consist of dark limestone units rich in organic material to bituminous shale, the upper part of which is rich in phosphate (Mégard, 1978; Loughman and Hallam, 1982; Prinz, 1985; Rosas *et al.*, 1997). In many southern areas, the Sinemurian transgressive beds are the first marine deposits to be recorded. They may locally onlap onto the Paleozoic basement (Antofagasta, Muñoz *et al.*, 1988; Baeza and Quinzio, 1991), or overlie undated (Triassic?) arc volcanic rocks (southern Peru, Vargas, 1970; Vicente *et al.*, 1982; Iquique and Arica, Muñoz and Charrier, 1993).

In the back-arc areas of central Peru, scattered lava flows interbedded in Early Liassic marine sediments display alkaline chemistry, indicating that extensional tectonic regime still prevailed (Rosas and Fontboté, 1995; Romeuf *et al.*, 1997; Figs. 9 and 12). In summary, the Late Triassic-Early Jurassic period is marked by a mild extensional regime, interpreted as related to the break-up of Gondwana and to the Tethyan rifting. Large-scale marine transgressions occurred in the Late Norian and Early Sinemurian, respectively. Subduction of a paleo-Pacific Plate beneath this part of the South American Plate has not been proved as yet.

Late Sinemurian - Kimmeridgian (195 - 150 Ma)

This interval is marked by active subduction-related magmatism, sinistral wrenching movements along NW-SE faults, and tectonically controlled sedimentation. Three main domains are distinguished: (1) the NNE trending Ecuadorian-northern Peruvian segment is characterized by a NNE-trending magmatic arc bordered to the E by a back-arc subaerial basin; (2) in northern and central Peru, the carbonate shelf sedimentation continued and was interrupted by a Middle Jurassic tectonic event; (3) in southern Peru and northern Chile, the western magmatic arc was flanked to the E by a marine, subsiding back-arc basin, bordered by emergent eastern areas (Fig. 8).

In Colombia (Mojica and Dorado, 1987), southern Ecuador (Romeuf *et al.*, 1997) and northwestern Peru (Pardo and Sanz, 1979), Middle to Late Sinemurian marine deposits are intercalated with alkaline basaltic flows, whereas subduction-related volcanic arc rocks appear in the immediately overlying deposits, which evolve toward subaerial environments. The Late Sinemurian age of the beginning of the magmatic arc activity is confirmed by dates obtained from associated arc-related intrusions (195 - 190 Ma, Aspden *et al.*, 1987; Litherland *et al.*, 1994). The activity of this NNE trending continental magmatic arc (Misahualli and Colán formations) continued on during the Middle Jurassic, and eventually ceased by latest Jurassic times (150 - 140 Ma, Aspden *et al.*, 1987; Litherland *et al.*, 1994, Romeuf *et al.*, 1995). To the W of the magmatic arc, marine volcaniclastic sediments, dated as late Middle to early Late Jurassic in southern Ecuador and northern Peru (Mourier, 1988), are interpreted as fore-arc deposits.

In southwestern Peru, the disconformable marine transgression is dated as Early Sinemurian, Early Toarcian, Late Toarcian (commonly), or even Late Bajocian-Bathonian, according to the regions, suggesting either a contrasted and complex paleogeography, or the juxtaposition of distinct units due to subsequent displacements (Kurth *et al.*, 1996). Most of the successions display a sedimentary hiatus of Late Bajocian-Bathonian age (Vicente, 1981, 1989). Volcanic arc rocks formerly ascribed to the Triassic-Liassic (Chocolate Formation) yielded radiometric dates of 177 - 157 Ma and Aalenian-Bathonian paleontological ages (Roperch and Carlier, 1992; Romeuf *et al.*, 1995), thus suggesting that they are coeval with the magmatic arc of Ecuador, northern Peru and northern Chile. These volcanic rocks are overlain by volcaniclastic and siliciclastic marine shelf deposits of Aalenian-Callovian age (Upper Río Grande and Guaneros formations, Rüegg, 1956; Aguirre and Offler, 1985; Romeuf *et al.*, 1995), thus indicating a progressive and significant decrease of the arc activity during the Middle Jurassic. No volcanic rocks of Late Jurassic age known.

In the coast of northern Chile, outpouring of a thick accumulation of subduction-related dacite, andesite, basalt and tuff (La Negra Formation, Fig. 10) began also during the Late Sinemurian-Pliensbachian (Muñoz *et al.*, 1988; Baeza and Quinzio, 1991; Muñoz and Charrier, 1993). They follow basic intrusions related to probably transtensional sinistral movements (Pichowiak *et al.*, 1990). In northernmost Chile, this volcanic series is overlain by marine sediments of early Middle Jurassic to Late Jurassic age, according to the areas (Muñoz and Charrier, 1993; Kossler, 1997), but volcanic activity is assumed to have lasted until the Late Jurassic (Charrier and Muñoz, 1994) or even the Early Cretaceous (Mpodozis and Ramos, 1989; Scheuber *et al.*, 1994), farther S. Extensional or transtensional regime prevailed during the Jurassic, favouring the formation and play of the Atacama wrench fault system (Brown *et al.*, 1993).

East of the magmatic arc, the Jurassic back-arc basin received poorly dated, mainly argillaceous and volcaniclastic subaerial sedimentation (Chapiza and Sarayaquillo formations; Tschopp, 1953; Seminario and Guizado, 1976; Rivadeneira and Sánchez, 1991; Fig. 6). These sedimentary rocks rest on the Liassic carbonate beds to the W, or on the Paleozoic basement to the E. In the latter case, they may include unfossiliferous Late Triassic-Liassic beds. They comprise a lower sequence of shale, fine-grained sandstone, dolomite and evaporite of sabkha environment, and an upper sequence consisting of coarse-grained sandstone and conglomerate of fluvial environment (Tschopp, 1953; Mégard, 1978). However, Jurassic marine platform limestone units are locally known in easternmost Ecuador (Petroproducción, unpublished data).

In the back-arc areas of Peru, platform carbonate deposition, started in the Late Triassic, went on until the early Middle Jurassic (Figs. 11 and 12). This carbonate shelf was bordered to the E by continental back-arc deposits consisting in fluvial red beds to the N, and locally significant accumulations of eolian sandstone to the S (Sempere, 1995). No information is available on the western part of central Peru. On the Peruvian Platform, the bituminous facies and phosphate-rich carbonate beds of the Sinemurian transgression (Aramachay Formation) are overlain by stratified oolitic and skeletal limestone presenting local cross-stratified calcarenite beds of Late Sinemurian to Late Toarcian, maybe Aalenian age (Condorsinga Formation; Mégard, 1978; Westermann *et al.*, 1980; Loughman and Hallam, 1982; Prinz, 1985). These Liassic deposits are overlain by sandy-argillaceous limestone of Late Aalenian-Late Bajocian age (Chunumayo Formation, Mégard, 1978; Westermann *et al.*, 1980). The carbonate series ends locally (Huancayo) with unconformable fluvial sandstone (Cercapuquio Formation) overlain by laminated carbonate, shale and evaporite beds of tidal flat environment, ascribed to the Bajocian (Chaucha Formation; Mégard, 1978; Moulin, 1989; Fig. 11). However, in many areas, the upper part of the platform carbonate sequence is eroded (Prinz, 1985), due to the Bathonian tectonic event.

In the back-arc areas of northern Chile, the Sinemurian transgressive beds are overlain by open marine deposits (Chong, 1976; Vicente *et al.*, 1982; Muñoz *et al.*, 1988; Ardill *et al.*, 1998). In southern Peru (Arequipa Trough), this succession comprises a Pliensbachian to Bajocian diachronic transgression, Bathonian turbidite beds, Callovian black shale and Oxfordian-Kimmeridgian shelf sandstone (Vicente, 1981, 1989; Vicente *et al.*, 1982; Fig. 13). In northernmost Chile, the turbidite beds progressively disappear and carbonate interbeds appear in the succession (Muñoz *et al.*, 1988). Farther S (Domeyko Cordillera), the succession is dominated by offshore shale deposits locally rich in organic matter (Chong, 1976; Ardill *et al.*, 1998). These marine back-arc basins are bordered to the E (southern Peru, Bolivia) by emergent areas which received locally thick, coarse-grained, quartz-rich eolian sandstone ascribed to the Early to Middle Jurassic (Ravelo and Ichoa formations; Oller and Sempere, 1990; Carlotto, 1998; Sempere *et al.*, 1998).

Scattered alkaline lava flows are known during this interval in the back-arc areas. Alkaline basalt flows are interbedded in the Sinemurian limestone of central Peru (Rosas *et al.*, 1997), scarce sills and intracontinental tholeiitic flows dated at 185 - 170 Ma occur in Bolivia (late Early to early Middle Jurassic, Soler and Sempere, 1993; Sempere *et al.*, 1998), and volcanic flows interbedded in the Bajocian to Callovian deposits of the Domeyko Basin are regarded as related to transtensional movements (Ardill *et al.*, 1998; Fig. 14). On the other hand, the mid-Jurassic Arequipa Trough has been interpreted as a pull-apart basin due to sinistral movements (Vicente *et al.*, 1982). Therefore, this period seems to be dominated by a mild extension, probably due to a sinistral transtensional regime (see also Scheuber *et al.*, 1994).

However, the southern part of the area (10°S - 25°S) is characterized by a sedimentary hiatus of Late Bajocian-Bathonian age (Mégard, 1978; Vicente, 1989; Ardill *et al.*, 1998). This "Bathonian phase" is classically regarded as responsible for the disconformity that separates the lower and upper red beds of the Peruvian Eastern Basin (Mégard, 1978). It is coeval with a rapid exhumation (3000 m) of the Eastern Cordillera of central Peru (Laubacher and Naeser, 1994), with metamorphism of the arc zone of northern Chile (170 - 150 Ma, Lucassen and Thirlwall, 1998) and with intense folding and reverse faulting of the arc zone of north central Chile (163 - 140 Ma, Irwin *et al.*, 1997). Therefore, a poorly documented compressional deformation seems to have occurred in the Late Bajocian-Bathonian.

In summary, between 195 and 140 Ma, the NNE trending Colombian-Ecuadorian and the N-S trending Chilean segments are characterized by abundant arc magmatism, unstable tectonic regime, subaerial back-arc sedimentation and locally marine fore-arc deposits. At the same time, transtensional, probably sinistral tectonic regime prevailed along the NW-SE trending Peruvian segment, which resulted in localized pull-apart basins and in the alternation of marine sedimentation and emergent periods. This situation has been interpreted as the result of a southeastward convergence of the paleo-Pacific Plate beneath South America (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995).

Kimmeridgian to late Paleocene (140 - 57 Ma)

During the Kimmeridgian-Paleocene interval, the paleogeographic framework was still controlled by the fore-arc, arc and back-arc zones. However, the trend of the arc zone became NE in Peru and N-S in Chile, and the back-arc area presented a three-fold subdivision inherited from the Jurassic structures. No magmatic arc is known at this time in Ecuador, which constituted a back-arc domain.

The Kimmeridgian-early Late Cretaceous fore-arc zone is virtually unknown. During the latest Cretaceous, fore-arc zones appear in northernmost Peru-southwesternmost Ecuador and southern Peru-northern Chile, due to the eastward shift of the magmatic arc from 80 Ma. The magmatic arc crops out widely in coastal Peru and northern Chile. The northern extension of the NW trending Peruvian arc crosscuts obliquely the southernmost part of Ecuador, since the subduction zone probably extended northwestwards into the oceanic realm. However, insights into the geodynamic evolution of the East-Pacific oceanic realm are provided by the analysis of the stratigraphic content and geochemical

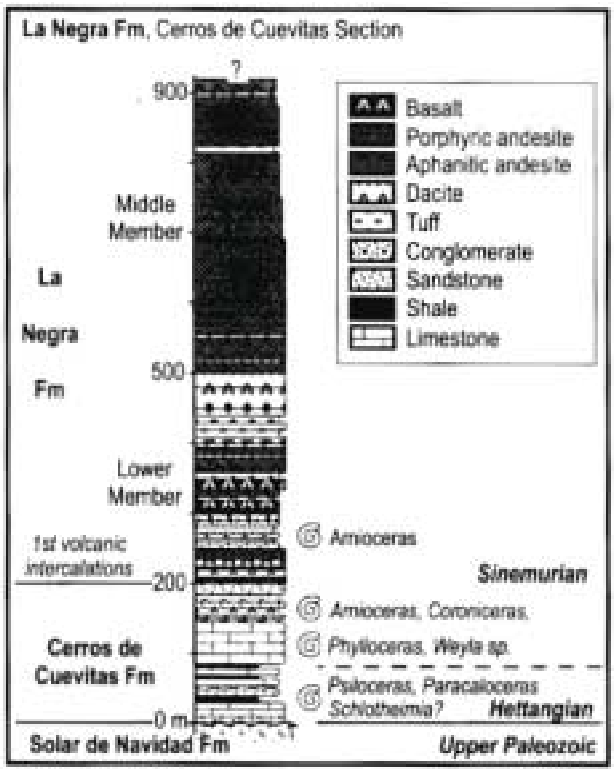


FIGURE 10 - Section of the La Negra Formation in the Jurassic arc zone of northern Chile (after Muñoz et al., 1988).

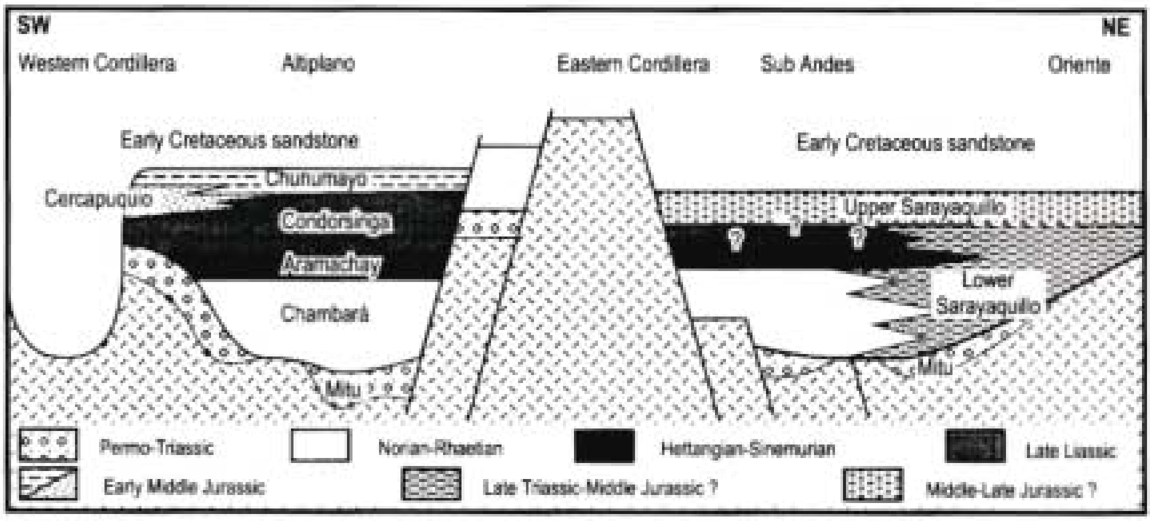


FIGURE 11 - Paleogeographic sketch of the Pucara Group (Late Triassic-Early Middle Jurassic) of Central Peru (after Loughmann and Hallam, 1982). No scale.

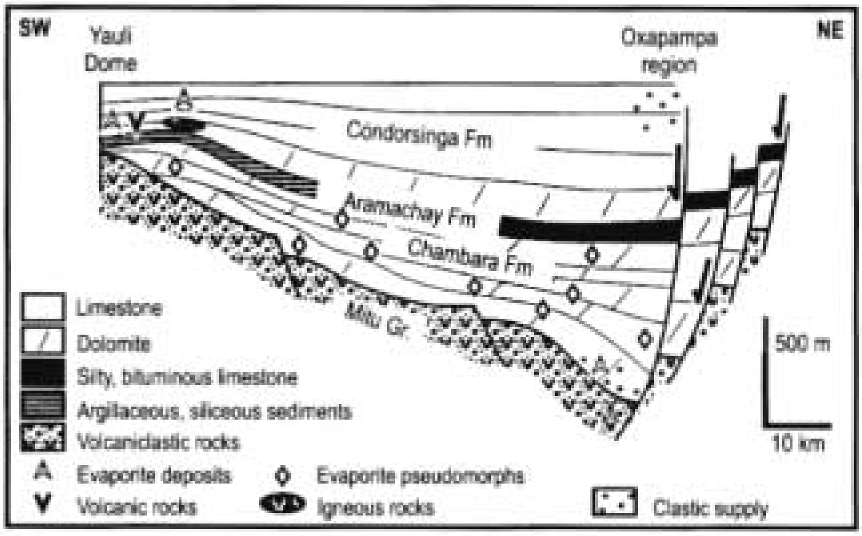


FIGURE 12 - Paleogeographic sketch of the Pucara Group (Late Triassic-Early Middle Jurassic) of Central Peru (after Rosas and Fontboté, 1995). Compare with Fig. 11.

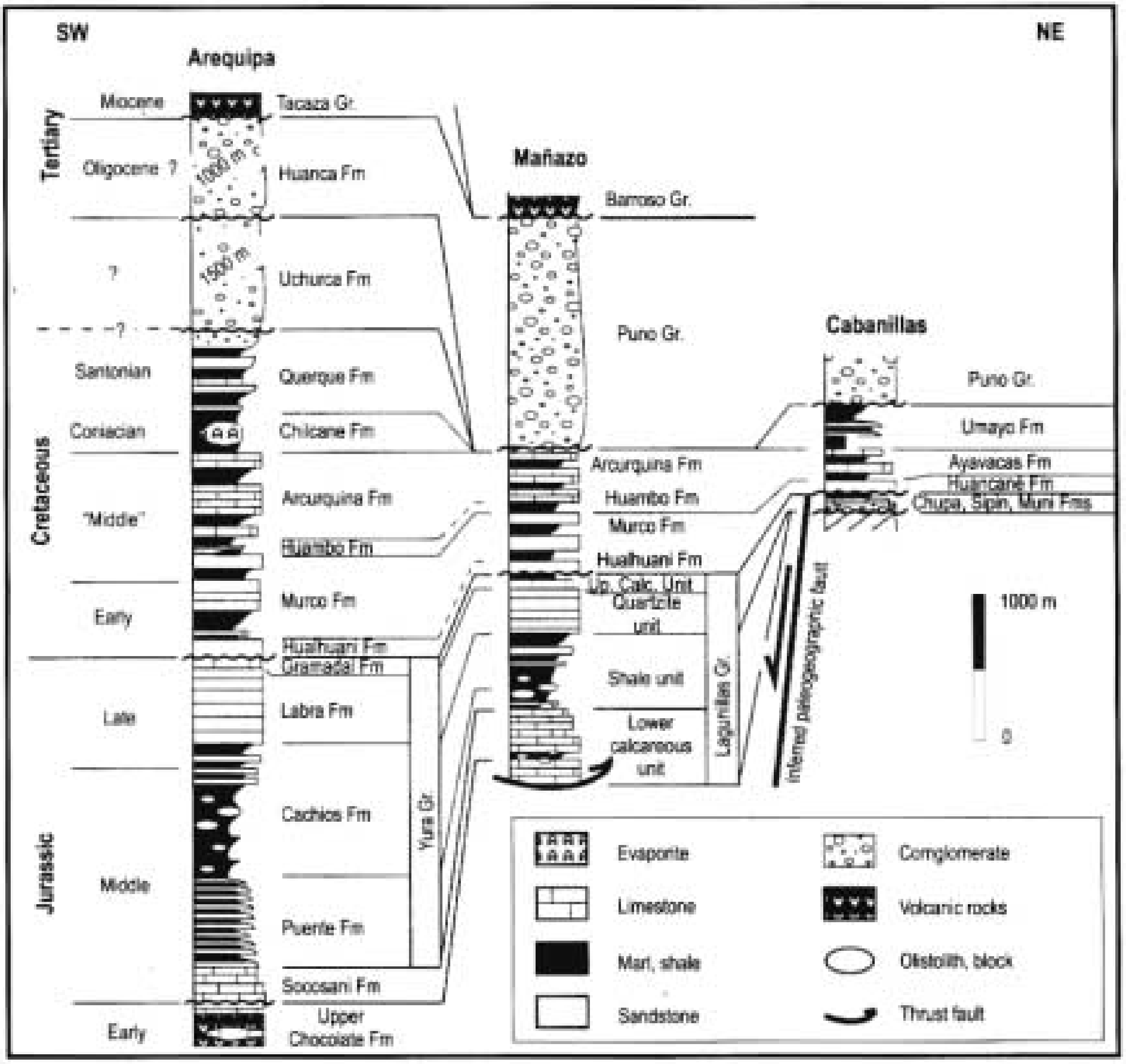


FIGURE 13 - Sections of the Jurassic-Tertiary ssuccessions accross the Arequipa Basin of southwestern Peru (after Jaillard and Santander, 1992).

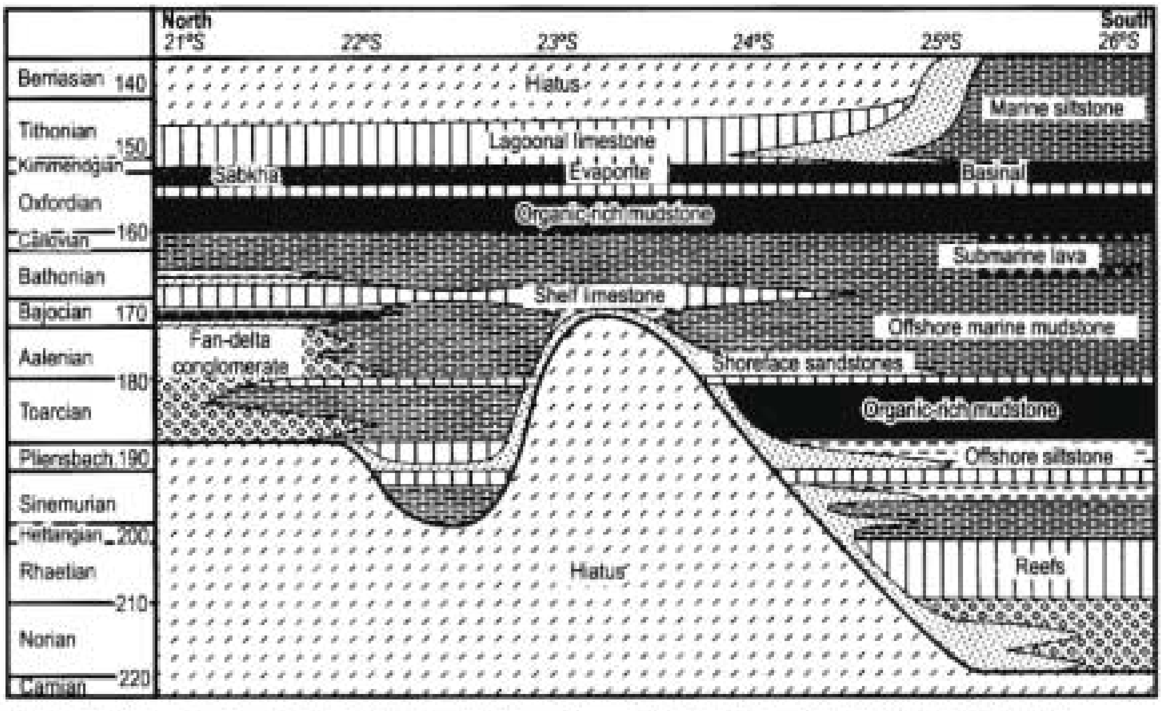


FIGURE 14 - Chronostratigraphic chart of the Jurassic back-arc Domeyko Basin (northern Chile) (after Ardill et al., 1998).

signatures of the Cretaceous oceanic terranes subsequently accreted to the Ecuadorian margin.

In Peru, the back-arc area comprised a western, mobile and subsiding basin (West-Peruvian Trough of Benavídes, 1956, present-day Western Cordillera), an axial positive threshold (Marañon Geanticline, Cuzco-Puno Swell, and present-day Eastern Cordillera) and an eastern, stable and moderately to little subsiding basin (present-day Subandean Zone and eastern basin, Fig. 16). The latter extends northwards into Ecuador (present-day Oriente Basin), and southwards into Bolivia (Potosí Basin). The Kimmeridgian-late Paleocene time-span (150 - 55 Ma) can be divided into three periods, characterized by distinct paleogeographic and tectonic settings, controlled by the plate kinematics evolution.

Kimmeridgian? - Aptian (150 - 110 Ma)

Latest Jurassic to earliest Cretaceous times are marked by a complete reorganization of the paleogeographic pattern and tectonic evolution of the Andean margin, interpreted as the result of a drastic change in the convergence direction, which triggered a magmatic arc re-organization and significant tectonic deformation (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995). These events were followed in the Early Cretaceous by an important diachronous marine transgression in the whole area.

Kimmeridgian? - Berriasian (150 - 135 Ma)

In Ecuador and northern Peru (and Colombia), the activity of the continental volcanic arc (Misahualli and Colán formations) ceased by the end of Kimmeridgian times (150 to 140 Ma ago, Aspden *et al.*, 1987; Mourier, 1988; Litherland *et al.*, 1994; Fig. 17). The deformed and eroded magmatic arc is then overlain by unconformable fluvio-marine sandstone, the diachronous base of which is dated as Valanginian (?) to Albian, from SW to NE (Villagómez *et al.*, 1996; Robert *et al.*, 1998). This major hiatus and unconformity suggests the occurrence of a significant latest Jurassic-Early Cretaceous tectonic event (Litherland *et al.*, 1994). No Kimmeridgian deposits are known in the fore-arc areas of Ecuador and northern Peru.

In southwestern Ecuador, the Raspas high pressure metamorphic complex yielded a 132 Ma K/Ar age (Feininger and Silberman, 1982) and 130 - 115 Ma Ar/Ar and Sm/Nd ages (Malfere, 1999). These are interpreted as cooling ages, subsequent to the accretion and HP metamorphism of an oceanic plateau and accretionary prism in the latest Jurassic-earliest Cretaceous (Gabriele *et al.*, 1999; Malfere *et al.*, 1999). This suture extends northwards along the western edge of the Eastern Cordillera of Ecuador (Peltetec suture, Litherland *et al.*, 1994; Aspden *et al.*, 1995) and to the W of the Central Cordillera of Colombia (Amaime Terrane, Aspden and McCourt, 1986; Toussaint and Restrepo, 1994). According to Litherland *et al.* (1994), this event corresponds to the accretion of a continental microplate (Chaucha Terrane) to the Ecuadorian margin.

In northwestern Peru, Early (?) Tithonian lagoonal deposits are abruptly overlain by deep shelf shale, and then by a thick series of coarse-grained volcaniclastic turbidite beds of Late Tithonian age, which reworked the Jurassic volcanic arc. This evolution points to the creation of a deep, N-S trending sedimentary basin interpreted as a pull-apart basin (Jaillard and Jacay, 1989), which extends southwards (Fig. 6). This succession follows (Jaillard and Jacay, 1989) with Late Tithonian black shale of shelf environment, Tithonian-Berriasian massive sandstone of nearshore environment (Tinajones Formation; Wilson, 1984), and disconformable massive clean sandstone of presumably Berriasian age (Chimú Formation, Benavídes, 1956; Jaillard and Jacay, 1989).

In the back-arc areas of Ecuador and northern Peru, the Late Jurassic back-arc red beds seem to grade upwards into coarser-grained red beds interbedded with basaltic to rhyolitic flows (Yaupi Mb, Jaillard, 1997), locally dated as earliest Cretaceous (Hall and Calle, 1982; Jaillard, 1997). This magmatic activity would be coeval with small-sized stocks, which intrude the Jurassic red beds and are disconformably capped by Early Cretaceous sandstone (Tschopp, 1953; Tafur, 1991).

In central and southern Peru, Kimmeridgian times were marked by the arrival of clastic deposits. Siliciclastic shelf sediments abruptly overlie the Callovian black shale of the Arequipa Trough (Vicente *et al.*, 1982; Fig. 13). These sediments are correlated with undated unconformable conglomerate of the Peruvian (Chupa Formation, Klink *et al.*, 1986; Jaillard, 1995) and Bolivian Altiplano (Condo Formation), which, however, might be younger. The Early Tithonian marine transgression is recorded in the western part of central and southern Peru by shallow marine limestone (Jaguay Formation, Rüegg, 1961; Gramadal Formation, Chávez, 1982; Batty and Jaillard, 1989), which appear to be overlain by Tithonian-Berriasian (?) black shale and sandstone (Oyón Formation, Mégard, 1978; Tiabaya outcrops, Geyer, 1982).

Volcanic arc activity is known in the Lima area (Atherton *et al.*, 1985; Alemán, 1996). It seems to have begun by Late Tithonian times, since the upper part of the arc series yielded ammonites of Late Tithonian age (Bulot, personal communication, 1998; formerly ascribed to the Berriasian; Wiedmann, 1981), and have continued during part of Berriasian times (Alemán, 1996). Disconformable clean massive sandstone is ascribed to the Berriasian in the western parts of central and southern Peru (Chimú, Goyllarisquizga and Hualhuani formations, Benavides, 1956; Batty and Jaillard, 1989). However, in spite of limited paleontological evidence, the base of these deposits seems to be diachronous, being much younger toward the E (Wilson, 1963; Jaillard, 1995; Robert *et al.*, 1998). In the eastern basins of Peru and Bolivia, no earliest Cretaceous deposits have been recognized so far, below the unconformable Early Cretaceous sandstone units.

In northernmost Chile, the Kimmeridgian phase is marked by a marine regression which a culminated with a local hiatus and an angular unconformity, by sinistral wrench movements along the N-S trending Atacama Fault System, and by subsidence of the back-arc areas (Bogdanic and Espinosa, 1994). In the back-arc basin of northernmost Chile, Oxfordian limestone and shale grade upwards into limestone and shale with interbeds of evaporite in the lower part, and sandstone intercalations in the upper part (Muñoz *et al.*, 1988). Farther S (Domeyko Basin), the Callovian marine black shale beds are overlain by evaporite beds deposited in basin to sabkha environments, and then by inner shelf limestone of Late Kimmeridgian-Early Tithonian age, which laterally grade southwards into marine siltstone (Chong, 1976; Ardill *et al.*, 1998; Fig. 14). Farther S (31°S), the Coastal Cordillera was deformed by W-verging open to tight folds between 140 and 126 Ma (Irwin *et al.*, 1987). The deformation is coeval with significant sinistral lateral displacements of fore-arc and arc slivers along the Atacama Fault System (26°S), dated between 145 and 125 Ma (Kurth *et al.*, 1996).

During the latest Jurassic-earliest Cretaceous, magmatic arc activity seems to have continued without changes of location. However, a magmatic gap seems to have occurred between 150 and 140 Ma (Hammerschmidt *et al.*, 1992; Fig. 15) and the magmatic activity appears to decrease significantly near the Jurassic-Cretaceous boundary (Mpodozis and Ramos, 1989; Scheuber *et al.*, 1994; Charrier and Muñoz, 1994).

The end of the NNE trending Ecuadorian-north-Peruvian magmatic arc by Tithonian times, and the beginning of the *activity of the NW trending Peruvian magmatic arc in the Late* Tithonian, expresses a drastic change in the convergence direction, which passed from nearly southwards to nearly northeastwards (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995). This plate kinematics reorganization correlates with geodynamic events in the southeastern Pacific and the Tethys Ocean. During the Late Jurassic, the southeastern Pacific accretion ridge would have been oriented roughly NE-SW and connected with the Tethyan accretionary system (Caribbean and Central Atlantic oceans, Duncan and Hargraves, 1984). The outpouring of a major plume along the *Pacific accretionary ridge in Tithonian times (Chatzky ridge)* would have disturbed the accretion direction; brought about the break-up of the Eastern paleo-Pacific Plate, and created a triple junction (Nakanishi *et al.*, 1989). Meanwhile, the Tethyan realm was marked by a significant slowdown in the spreading rates, allowing the newly created NW-SE trending Southeastern Pacific Ridge to impose a northeastward drift direction for the Eastern paleo-Pacific Plate. These events would have provoked, in the Tithonian (140 Ma), a change of convergence direction, according to the process proposed by Duncan and Hargraves (1984) for Early Cretaceous times.

This major geodynamic change may account for the accretion of the oceanic terrane of Ecuador and Colombia, the creation of the Chicama Basin of northern Peru, the widespread emergence and subsequent gap of Late Tithonian-Berriasian deposits, the unconformity of the Early Cretaceous deposits and the possible compressional deformation recorded in northern Peru. These events may be correlated with the Araucan phase of northern Chile (Stipanicic and Rodrigo, 1969; Scheuber *et al.*, 1994).

Berriasian-Aptian (140 - 110 Ma)

This period is marked by the widespread deposition of disconformable, diachronous quartz-sandstone units, by a relative tectonic quiescence; and along the Peruvian and Chilean margins, by the ongoing, although mild, volcanic arc activity. In Ecuador and northern Peru, very little is known about this period. In the back-arc areas of Ecuador, one K/Ar date and one palynological age suggest that subaerial red bed deposition continued during part of this period (Bristow and Hoffstetter, 1977; Baldock, 1982). The overlying transgressive disconformable sandstone beds are of Albian age (Jaillard, 1997). In the Cordillera Real of Ecuador, Litherland *et al.* (1994) mentioned numerous K/Ar resets in the Jurassic granite, interpreted as due to significant dextral movements related to the collision of displaced terranes. In the neighbouring Marañon Basin of northeastern Peru, a sedimentary hiatus seems to separate the Jurassic red beds and the disconformable, diachronous transgressive sandstone beds of Early Cretaceous age (Laurent, 1985).

Meanwhile, somewhere in the southeastern paleo-Pacific domain, a mantle plume was responsible for the formation of a large and thick Oceanic Plateau that was accreted to the Andean margin in Late Cretaceous times (Cosma *et al.*, 1998; Reynaud *et al.*, 1999), and crops out presently in the Western Cordillera of Ecuador where it has been dated as Barremian or Hauterivian (123 Ma, Lapierre *et al.*, 1999).

In Central and Southern Peru, scarce outcrops of volcanic rocks in the coastal and western areas suggest that volcanic arc activity continued at least locally until Aptian times (Bellido, 1956; Vidal *et al.*, 1990; Alemán, 1996). However, the occurrence of thick intercalations of quartz-rich sandstone and unusual volcanic-free shelf limestone in the Lima area (Rivera *et al.*, 1975; Alemán, 1996) suggests that, either volcanic activity was local and sporadic, or part of the coastal terranes have been subsequently displaced (May and Butler, 1985).

Farther E, in the back-arc areas, well-sorted, clean, quartz-sandstone beds were deposited from Late Berriasian times onwards (Benavides, 1956; Wilson, 1963). In northern Peru and Ecuador, these were dated as Late Berriasian-Early Valanginian in the eastern region, (Benavides, 1956; Rivera *et al.*, 1975; Bulot, personal communication, 1998), Berriasian-Barremian in the western part of the Eastern Basin of Peru (Tarazona, 1985), Aptian in the centre of the basin, and early Late Albian in the eastern parts of the Eastern Basin (Villagómez *et al.*, 1996; Robert *et al.*, 1998, Fig. 18). In central Peru, such a diachronism was suggested by Wilson (1963), and, although paleontological evidence is poor, a comparable diachronism may occur in southern Peru (Jaillard, 1995).

Paleocurrents indicate that the Guiana and Brazilian shields were the sources of the clastic supply. Paleoenvironments evolve from subaerial/fluvial to nearshore/shallow marine from E to W, and deposition is mainly controlled by eustatic variations (Moulin, 1989). The isopach map clearly indicates an eastern depocenter situated in northern Peru (present Marañon River), and western depocenters N of Lima and around Arequipa (Jaillard, 1994; Fig. 19). Scarce syn-sedimentary tectonic features suggest a mild extensional regime (Moulin, 1989).

In the back-arc area of northernmost Chile, Late Jurassic strata are overlain by fine-grained sandstone, siltstone and shale with occasional tuff and andesitic lava of

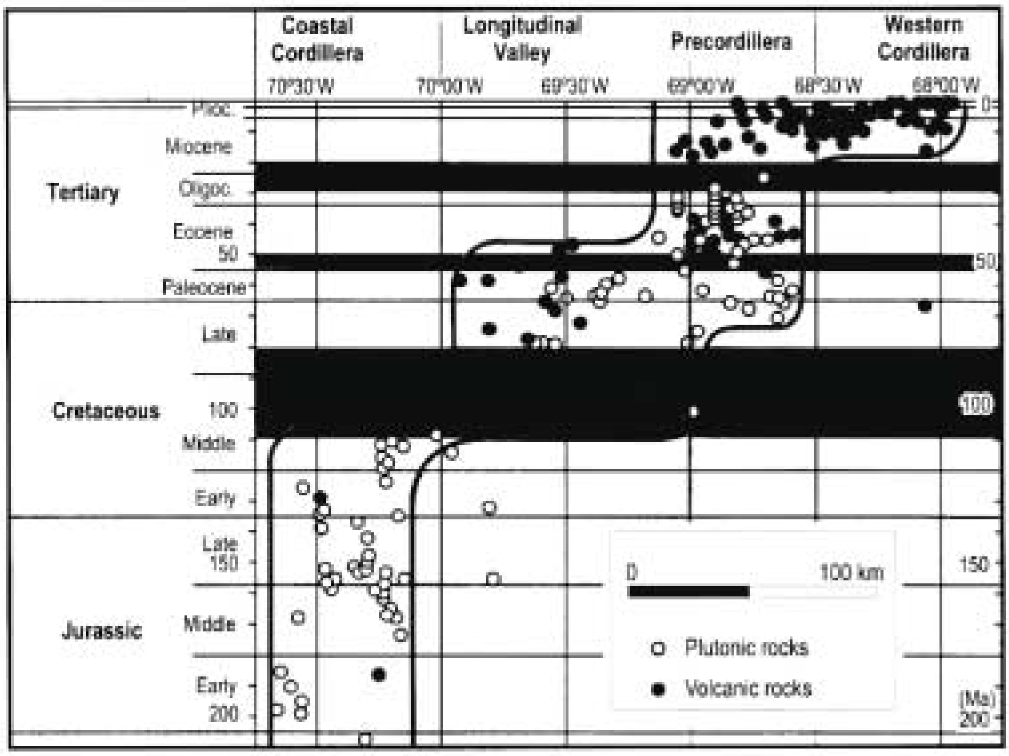


FIGURE 15 - Age, location and nature of the magmatic arc rocks in northern Chile (after Hammerschmidt et al., 1992).

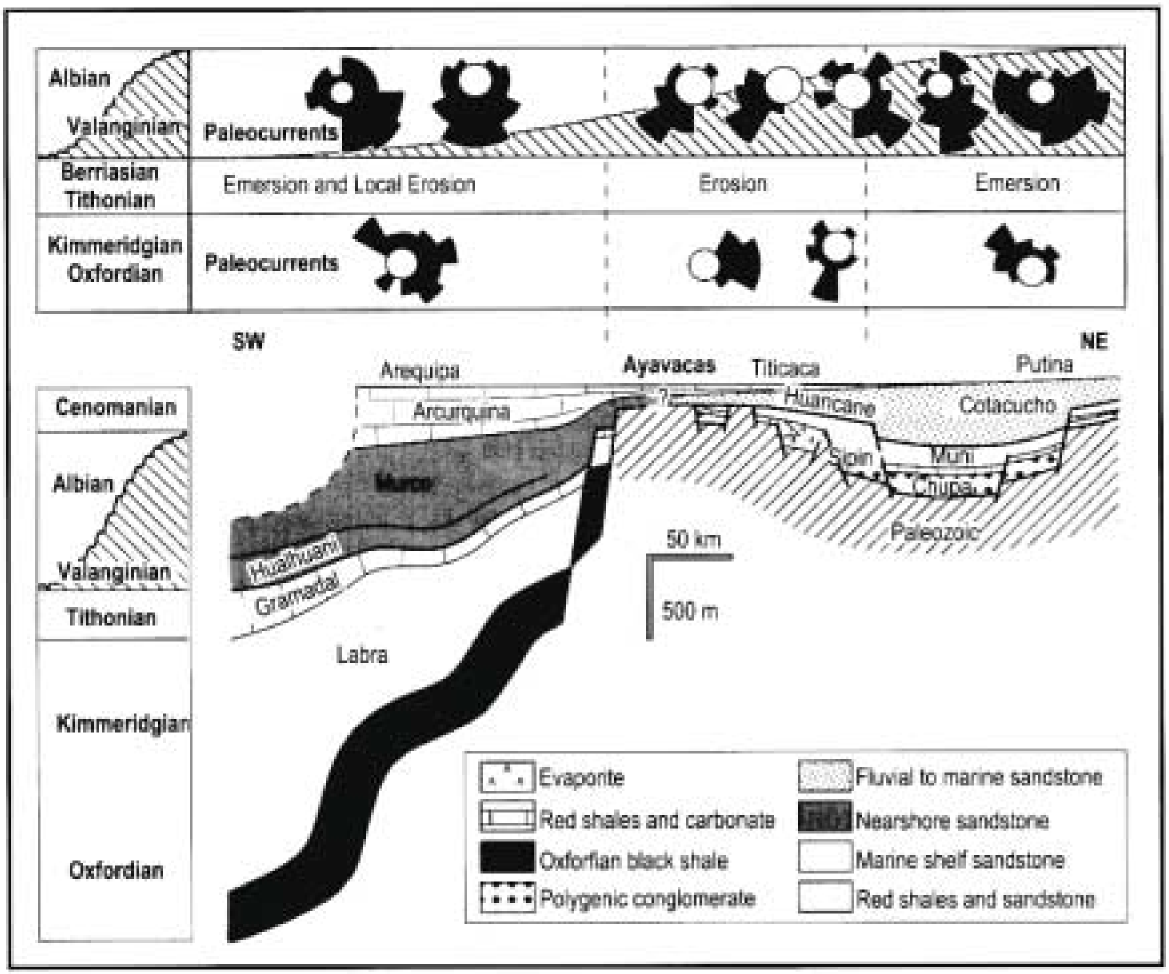


FIGURE 16 - Paleogeographic profile of southern Peru at the end of the Cenomanian (after Jaillard, 1994). Note the change in paleocurrent directions between Late Jurassic and Early to Middle Cretaceous times.

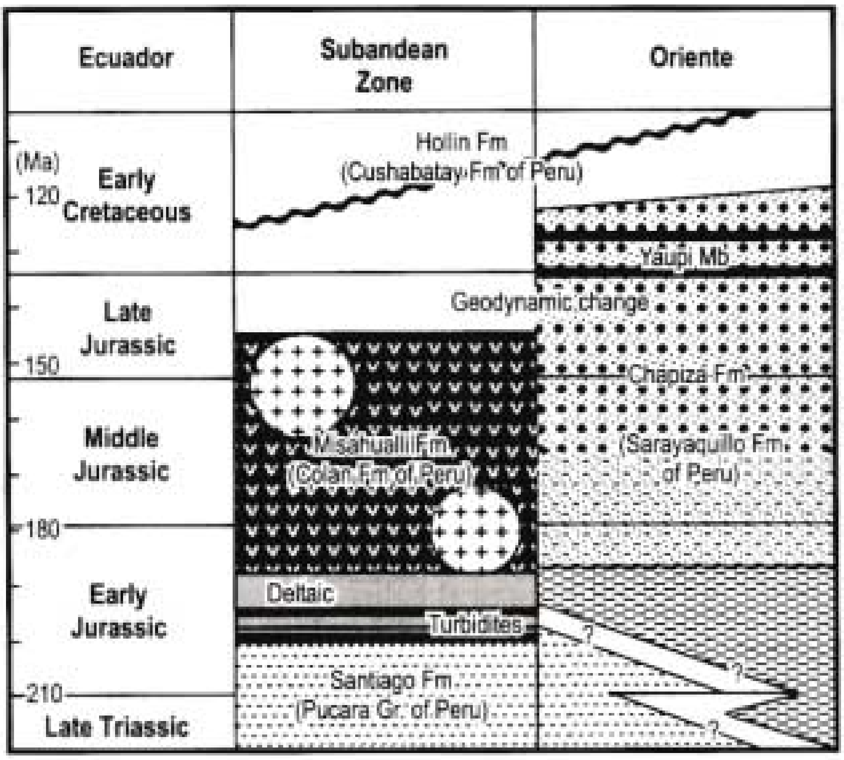


FIGURE 17 - Chronostratigraphic sketch of the Jurassic-Early Cretaceous series of eastern Ecuador (after Jaillard, 1997).

Kimmeridgian-Barremian age (Bogdanic and Espinosa, 1994; Scheuber *et al.*, 1994). Farther S, a limestone unit of Hauterivian-Barremian (and Aptian?) age crops out in the Coastal Cordillera (Bogdanic and Espinosa, 1994). The Early Cretaceous volcanic arc crops out farther S in the Coastal Cordillera. There, magmatic arc activity went on without significant changes in the location of the magmatic front; a maximum activity is recorded in the Aptian (120 - 110 Ma, Hammerschmidt *et al.*, 1992; Fig. 15). In Bolivia, undated continental conglomerate and sandstone are ascribed to the Early Cretaceous, since they conceal the "Araucan" angular unconformity (Sempere, 1994). They crop out in the centre of the Potosi Basin and are associated with alkaline basalt and basaltic andesite indicating extensional conditions (Soler and Sempere, 1993). Farther to the SE, in northern Argentina, sub-alkaline to alkaline granitic plutons (130 - 120 Ma) indicate that an extensional tectonic regime prevailed in the distal back-arc areas (Viramonte *et al.*, 1999).

The abrupt arrival in the Early Cretaceous of huge amounts of detrital quartz derived from the E in northern South America, may be interpreted as the result of both the large-scale westward tilt of the South American Plate due to the South Atlantic rifting, and a significant climatic change with increased precipitation, which allowed the detrital material to be transported for large distances (Jaillard, 1994).

Albian - Turonian (110 - 88 Ma)

This period was marked by a large-scale marine transgression; by important magmatic activity along the Chilean and Peruvian margins; and by the beginning of compressional deformation ("Late Albian Mochica phase", Mégard, 1984; 105 - 100 Ma). The only known outcrop corresponding to the fore-arc zone is represented by the eastern flank of the Amotape-Tahuin Massif of northern Peru-southern Ecuador. There, Paleozoic rocks are covered by undated transgressive siliciclastic conglomerate beds, overlain by massive shelf-limestone and anoxic laminated black limestone of Early to Middle Albian age (Olsson, 1944; Fischer, 1956; Reyes and Caldas, 1987; Jaillard *et al.*, 1999). This succession is comparable to that of the coeval series known from the back-arc areas of northern Peru.

The shelf carbonate sedimentation was rapidly overlain by basinal black shale interbedded with siliciclastic turbidite beds of Late Albian age, exhibiting slumping and bearing large-scale olistoliths (Copa Sombrero Group or Formation: Morris and Alemán, 1975; Reyes and Caldas, 1987; Jaillard *et al.*, 1999). These express an unstable tectonic setting, interpreted as the result of the creation of a pull-apart basin (Lancones-Celica Basin) related to the northward migration of the Amotape-Tahuin Paleozoic massif (Jaillard *et al.*, 1999). In the Cenomanian and Turonian, continuation of the turbiditic sedimentation (Morris and Alemán, 1975) suggests that the northward migration of the Amotape-Tahuin fore-arc sliver went on, with a possible maximum tectonic activity during the Cenomanian (Jaguay Negro Formation).

During Albian times, the arc zone of Peru and southernmost Ecuador was marked by the outpouring of huge volumes of subduction-related calc-alkaline lava (Casma Group, Celica, Copara and Matalaque formations; Atherton *et al.*, 1983; Soler 1991; Reynaud *et al.*, 1996; Fig. 20), which are locally interbedded with ammonite-bearing sediments indicating a Middle to Late Albian age (Myers 1975; Reyes and Caldas, 1987). These lava flows are associated with volcaniclastic turbidite beds deposited in strongly subsiding intra-arc basins (Atherton and Webb, 1989), interpreted as pull-apart basins related to dextral wrenching (Soler, 1991). Folds in the Albian volcanogenic pile are locally

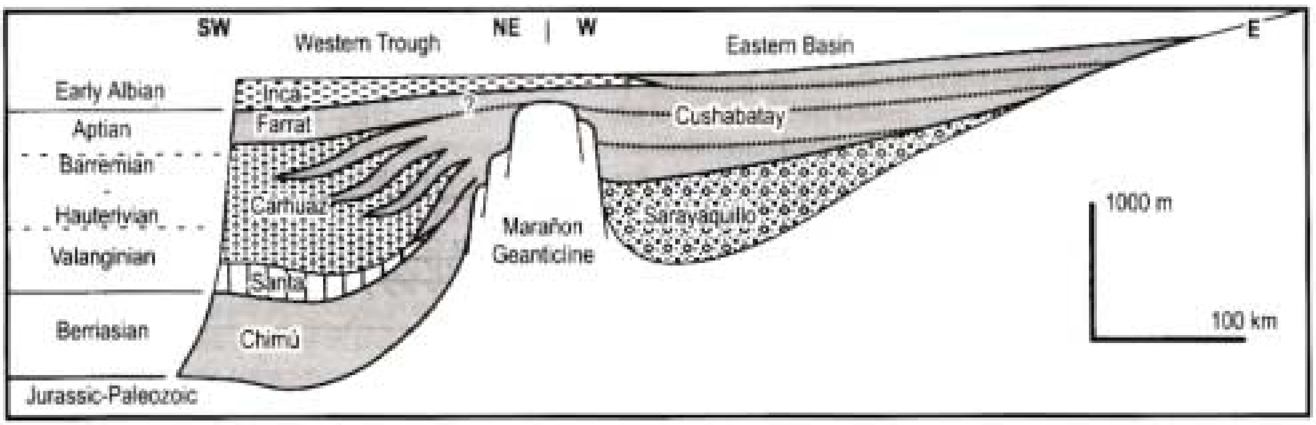


FIGURE 18 - Paleogeographic profile of southern Peru at the end of the Early Cretaceous.

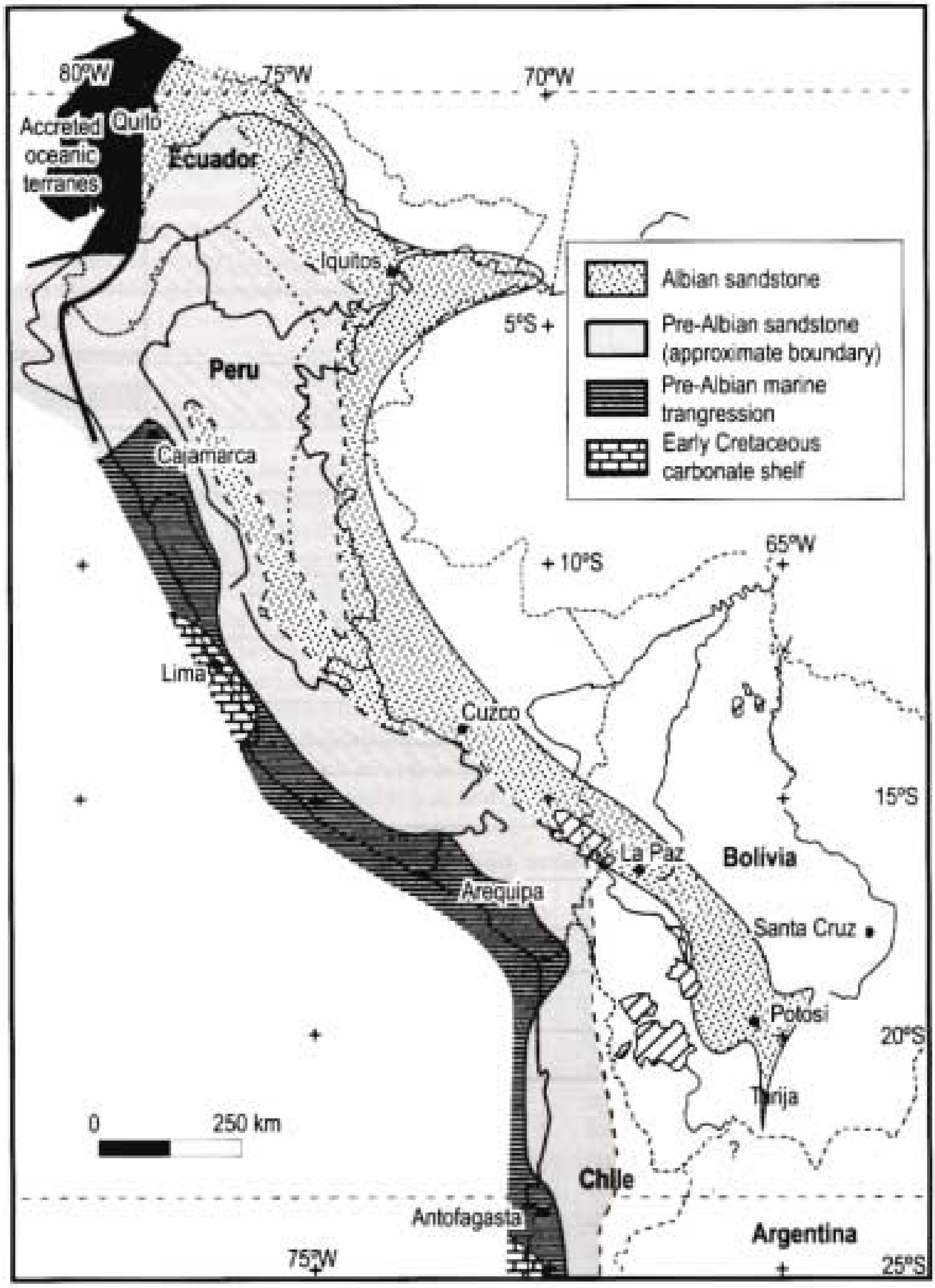


FIGURE 19 - Paleogeographic sketch of the Early Cretaceous transgression.

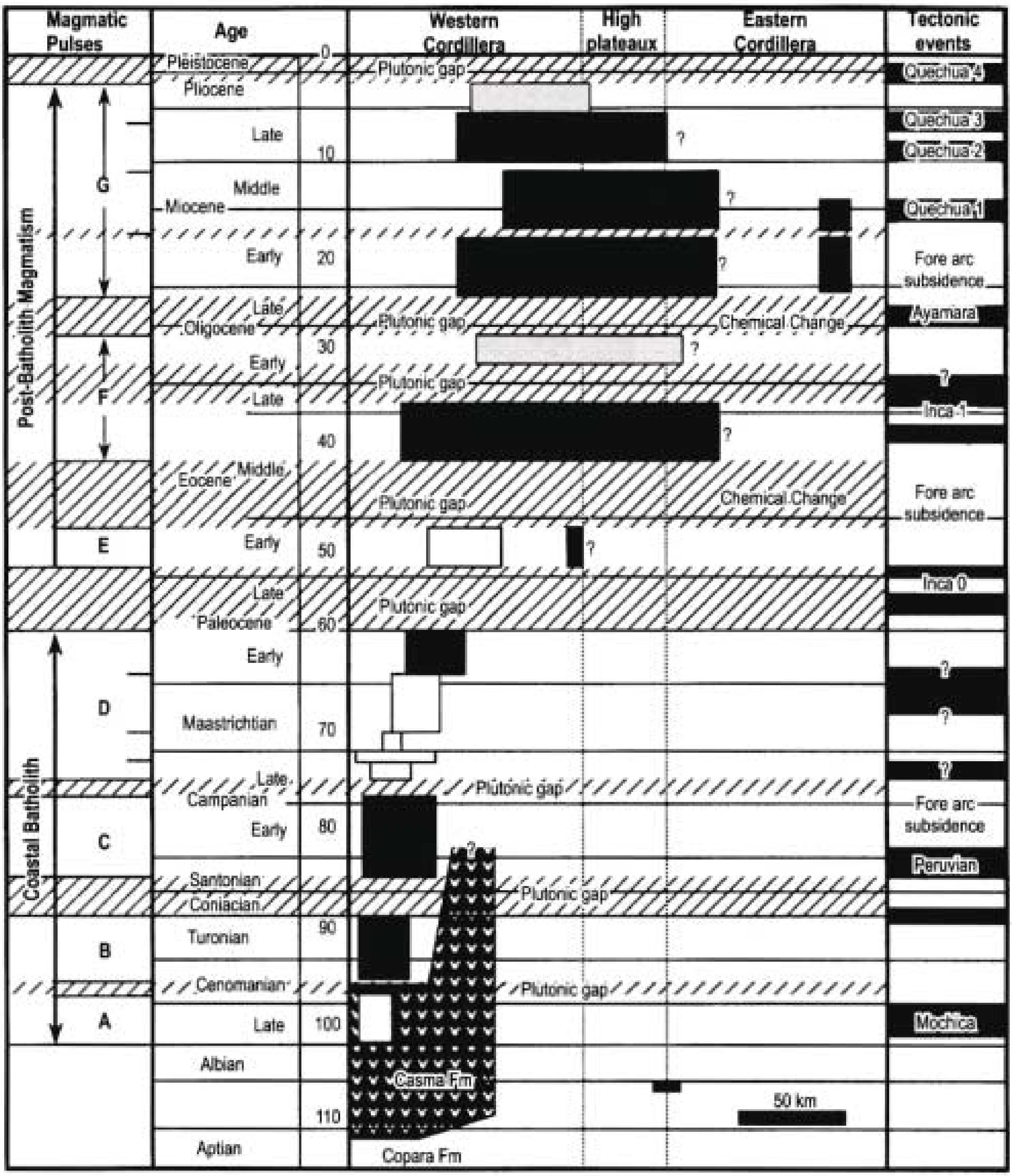


FIGURE 20 - Age, location and intensity of magmatic events in the magmatic arc of central Peru, and their relations to Andean tectonic events (after Soler, 1991). Light shaded areas: low magmatic activity, dark shaded areas: high magmatic activity.

cross-cut by basic to intermediate intrusions dated at 104 - 101 Ma (Wilson, 1975; Cobbing *et al.*, 1981; Bussel, 1983), thus indicating that compressional deformations began by Middle Albian times (Cobbing *et al.*, 1981).

Deformation was associated with significant dextral movements (Myers, 1975; Bussel and Pitcher, 1985). In the arc zones of Peru and Ecuador, the Late Albian tectonic compression is marked by local folding and faulting, by the end of marine sedimentation, by a general decrease of magmatic activity (Soler and Bonhomme, 1990), and by the replacement of volcanic effusions by plutonic intrusions, suggesting that the arc zones were significantly uplifted (Cobbing *et al.*, 1981; Soler, 1991). These calc-alkaline plutons, which intrude the volcanic arc, define the so-called Coastal Batholith of Peru (Pitcher, 1978; Cobbing *et al.*, 1981; Soler, 1991). Effusive magmatism ceased by earliest Cenomanian times, and incipient plutonic activity was rather low (Soler and Bonhomme, 1990), except in central-southern Peru where intrusions are dated at 101 - 94 Ma (Beckinsale *et al.*, 1985). The stability of the magmatic front in central Peru suggests that the Late Albian tectonic event did not change significantly the shape of the active margin (Soler and Bonhomme, 1990; Jaillard and Soler, 1996; Fig. 20). Most of the arc zone of Ecuador and Peru seems to have remained emergent during Cenomanian-Turonian times, since the Albian volcanic piles are usually unconformably capped by Santonian to Campanian transgressive sediments (Jaillard, 1995; Jaillard *et al.*, 1996). Plutonic activity was high during the Cenomanian (Beckinsale *et al.*, 1985; Mukasa, 1986, 94 - 90 Ma pulse of Soler and Bonhomme, 1990), but no intrusions of Turonian age are known (90 - 85 Ma plutonic gap, Soler, 1991).

In northern Chile, the locus of the magmatic arc significantly shifted eastward during Aptian-Albian times (Hammerschmidt *et al.*, 1992), and the Middle Cretaceous magmatic arc (115 - 90 Ma) is partly superimposed on the Jurassic arc (Charrier and Muñoz, 1994; Scheuber *et al.*, 1994; Fig. 15). In northernmost Chile, arc-related andesite, breccia, agglomerate, tuff, sandstone and conglomerate of Albian age overlie the Early Cretaceous deposits of the back-arc basin (Scheuber *et al.*, 1994). They yielded 115 - 104 Ma dates and are crosscut by 115 - 80 Ma intrusions (Bogdanic and Espinosa, 1994). An extensional or transtensional regime is assumed to have prevailed (Scheuber *et al.*, 1994). However, although no deformation of Albian age has been recognized, the significant eastward shift of the Middle Cretaceous arc (50 km) may result from a shortening event. Volcanic arc activity continued until early Late Cretaceous times (Scheuber *et al.*, 1994).

No arc-related magmatism is known on the continental margin N of 3°S (Ecuador). Therefore, the Peruvian subduction zone extended probably northwestwards into the oceanic domain by means of an intra-oceanic subduction zone, which gave way to the formation of island arcs of Albian to early Late Cretaceous age. This interpretation is supported by the occurrence, on the accreted oceanic terranes of Ecuador, of pre-Cenomanian island arc rocks (Las Orquídeas and Toachí units, Jaillard *et al.*, 1995; Benítez, 1995; Cosma *et al.*, 1998), overlain by volcaniclastic arc series of Cenomanian to Santonian age (Cayo and Pilatón formations, Faucher *et al.*, 1971; Kehrer and Van der Kaaden, 1979; Jaillard *et al.*, 1995). Note that, there too, volcanic activity seems to have ceased by Cenomanian times.

In the back-arc areas of Ecuador and Peru, the Aptian-Albian boundary is marked by scattered volcanic manifestations, varying from basaltic flows to rhyolithic tuff, intercalated within the first transgressive units. This bimodal volcanism as been locally determined as alkaline, indicating an intracontinental extension (Soler, 1989).

Earliest Albian times are marked by the beginning of a major large-scale marine transgression which reached its maximum extent by Turonian times (Figs. 13 and 21). The Lima area already received a marine sedimentation from Early Cretaceous times (Rivera *et al.*, 1975; Alemán, 1996). The Early Albian transgression first reached the western part of the back-arc areas (Benavides, 1956; Jaillard, 1995; Robert *et al.*, 1998), where it deposited red to yellow coloured silt and sandstone, with glauconitic and locally oolitic limestone in the upper part (Inca and Pariahuanca formations, Benavides, 1956; Wilson, 1963; Moulin, 1989).

The Albian transgression is then marked by three pulses of mid Early, early Middle and early Late Albian age (Robert *et al.*, 1998). The first pulse only reached the western part of the Eastern basins (Chulec Formation, Benavídes 1956; Wilson, 1963; Robert, in progress), whereas the third one reached locally the eastern border of the Eastern Basin of Ecuador (Basal Napo Shales) and northern Peru, where it may rest on Paleozoic rocks (Jaillard, 1997). This latter transgression probably reached also the axial swell of southern Peru and triggered the deposition of transgressive, partly marine sandstone (Huancané Formation, Carlotto *et al.*, 1995; Jaillard, 1995), which grade to the E into thicker deposits (lower part of Putina Group, Audebaud *et al.*, 1976; Jaillard, 1995). It could have reached also the Bolivian Potosí Basin where a few tens of metres of coarse-grained, locally conglomeratic sandstone are known (La Puerta Formation, Sempere, 1994). The early Middle and early Late Albian transgressive pulses are associated with widespread anoxic deposits in central and northern Peru and in Ecuador (Pariatambo, Basal Napo, Chonta formations, Villagómez *et al.*, 1996; Robert *et al.*, 1998). Late Albian times are then marked by the development of carbonate shelves in the western part (Yumagual Formation, part of Mujarrún, Jumasha and Arcurquina formations, Benavides, 1956, 1962; Wilson, 1963; Jaillard, 1987), and by the deposition of deltaic-fluvial sandstone in the eastern parts of the back-arc areas (Agua Caliente Formation, T sandstone, Putina Group), the progradation and retreat of which are mainly controlled by eustatic variations with a subordinate influence of tectonic events (Jaillard, 1994, 1997).

The effects of the Late Albian tectonic event are mild in the back-arc areas. The Late Albian shelf carbonate sedimentation exhibits slumps, syn-sedimentary faults and breccia, clastic dykes and differential subsidence, which together express an extensional regime (Audebaud, 1971; Jaillard, 1987, 1994). Farther to the E, progradation of deltaic systems may be regarded as the result of a slight uplift related to the Late Albian tectonic event (Jaillard, 1987, 1997). In northern Argentina, alkaline volcanic rocks (110 - 100 Ma) are interpreted as related to a rift episode (Viramonte *et al.*, 1999).

In the western part of the back-arc areas of Ecuador and Peru, the carbonate shelf sedimentation recorded significant eustatic transgression near the Albian-Cenomanian boundary, in Middle Cenomanian, early Late Cenomanian (widely characterized by *Neolobites vibrayeanus* (= *N. kummeli*), and Early Turonian times. Each transgression is marked by conspicuous marly levels which grade upwards into massive platform carbonate exhibiting frequently desiccation features, thus indicating a shallow marine environment (Jaillard, 1987, 1995, 1997; Fig. 22). In some southern parts of the Eastern Basin of Peru, the Albian-Cenomanian fluvio-marine sandstone beds (Oriente and Putina groups) are overlain by Early Turonian marine shale, illustrating the large extent of the Turonian transgression. In some areas, the upper part of the Turonian shelf limestone exhibits mild syn-sedimentary tectonic features announcing the tectonic event of the Turonian-Coniacian boundary (Jaillard, 1987, 1995, 1997).

Subsidence was intense in the western areas of northern Peru, and decreased drastically toward the NE and SE. The Albian-Turonian series reaches nearly 2000 m in the western part of northern Peru (Benavides, 1956; Wilson, 1963), about 300 m in the Oriente Basin of Ecuador (Jaillard, 1997), 600 m in southwestern Peru (Benavides, 1962; Jaillard, 1995), and 30 m in the Potosí Basin of Bolivia (Sempere, 1994). The axial swell continuously behaved as a positive area.

The Middle to Late Albian deformation was the first significant compressional deformation recorded in the Cretaceous evolution of the Andean margin, which affected mainly the fore-arc and arc zones. It coincided with a period of high convergence rate and with the opening of the South Atlantic Ocean at equatorial latitudes, which triggered the westward drift of the South American Plate and therefore, the trenchward motion of the overriding plate (Frutos, 1981; Jarrard, 1986; Soler and Bonhomme 1990; Jaillard and Soler, 1996). The strong dextral wrench component of this deformation (Bussel and Pitcher, 1985; Soler, 1991; Jaillard, 1994) resulted from the northeastward motion of the Farallón Plate, indicated also by the lack of any arc magmatism along the Colombian-Ecuadorian margin (Aspden *et al.*, 1987). Due to the oblique direction of the oceanic plate, convergence was accommodated by lateral displacement and wrenching deformation along the edge of the active margin, rather than by shortening of the whole margin. The resumption of volcanic activity along the Peruvian margin may be related to the beginning of the Middle Cretaceous period of high convergence rate (Soler, 1991).

Coniacian - late Paleocene (88 - 57 Ma)

This period is marked by a major change in the paleogeographic pattern, by the occurrence of compressional tectonic events, the intensity of which increased through time, and by the incipient eastward migration of the arc zone in Peru. A progressive but general marine regression, the arrival of fine-grained detrital deposits and the eastward shift of depocenters in the eastern basins marked sedimentation. Tectonic events are of Late Turonian-Early Coniacian (88 Ma), Santonian (85 Ma), Late Campanian? (80 - 75 Ma) and Late Maastrichtian age (70 - 65 Ma). Some of these events coincide with the accretion of oceanic terranes in Ecuador or northern Peru.

Turonian-Coniacian boundary event and Coniacian - Early Santonian evolution (88 - 85 Ma)

In the fore-arc Celica-Lancones Basin (northern Peru-southern Ecuador), the youngest fossil recovered from the turbidite series is of Early Conician age (Petersen, 1949; Jaillard *et al.*, 1999). A sedimentary gap occurred then during Late Coniacian and Santonian times. In the Talara fore-arc basin of northwestern Peru, no deposits are known between the Albian shelf carbonate and the Campanian transgressive marine deposits (González, 1976; Morales, 1993). No information is available on the other fore-arc zones. In the oceanic fragments accreted to Ecuador, the Turonian-Coniacian boundary roughly coincides with the beginning of the Cayo island-arc activity, as expressed by thick series of coarse-grained volcaniclastic turbidite beds (Cayo and Pilatón formations, Jaillard *et al.*, 1995; Benítez, 1995; Kehrer and Van der Kaaden, 1979).

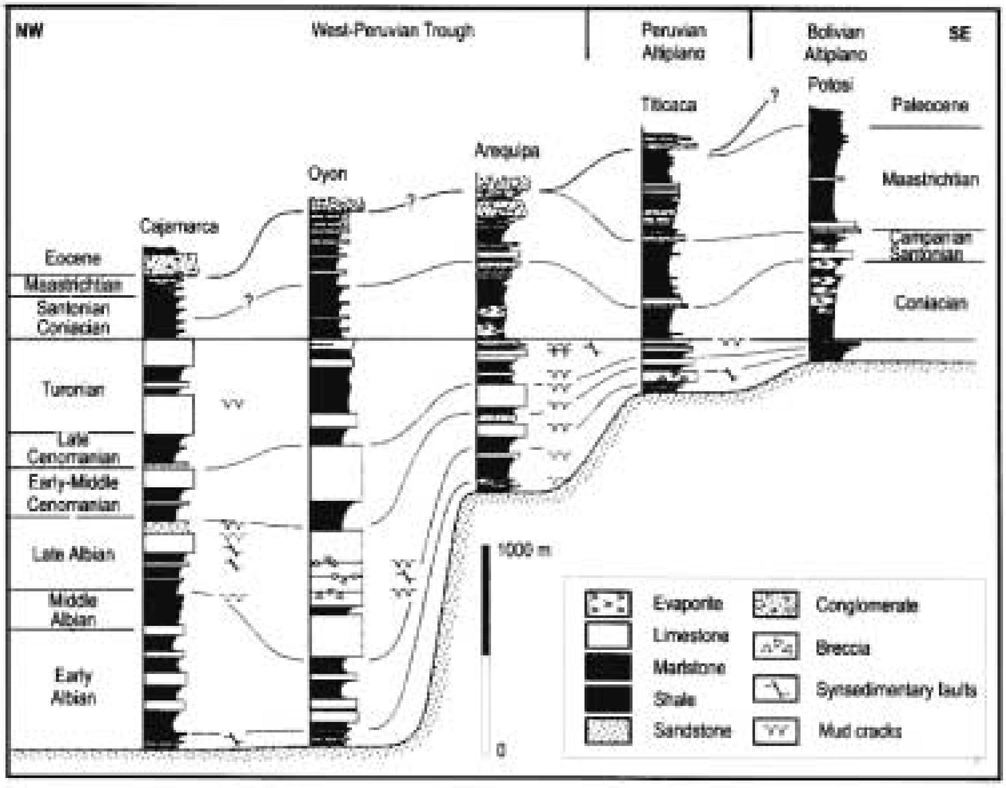
The Albian volcanic arc series of southern Ecuador are unconformably capped by Early Campanian transgressive marine deposits (Naranjo Formation, Jaillard, 1997). In southern Peru, Albian volcanic rocks are unconformably capped by undated shelf limestone of Senonian age (Omoye Formation, Vicente, 1981; Jaillard, 1994). In the arc zones, the effects of the Late Albian and Turonian-Coniacian deformation are, therefore, indistinguishable. However, the deformational event seems to have occurred before the Early Campanian transgression of southern Ecuador. In the Coastal Batholith of Peru, significant wrench movements associated with a variable compressional regime has been recognized during Turonian-Coniacian times (Bussel and Pitcher, 1985). Plutonic intrusions are very scarce, and volcanic activity is unknown (90 - 85 Ma magmatic gap, Soler, 1991; Fig. 20). However, although the magmatic front appears to be nearly stable, a slight eastward shift of some kilometres can be detected (Soler and Bonhomme, 1990; Jaillard and Soler, 1996). In northern Chile, the magmatic arc is marked by a well-expressed magmatic gap between 90 and 80 Ma (Hammerschmidt *et al.*, 1992; Scheuber *et al.*, 1994), which followed (Hammerschmidt *et al.*, 1992) or was associated with the eastward shift of the magmatic front (Scheuber *et al.*, 1994).

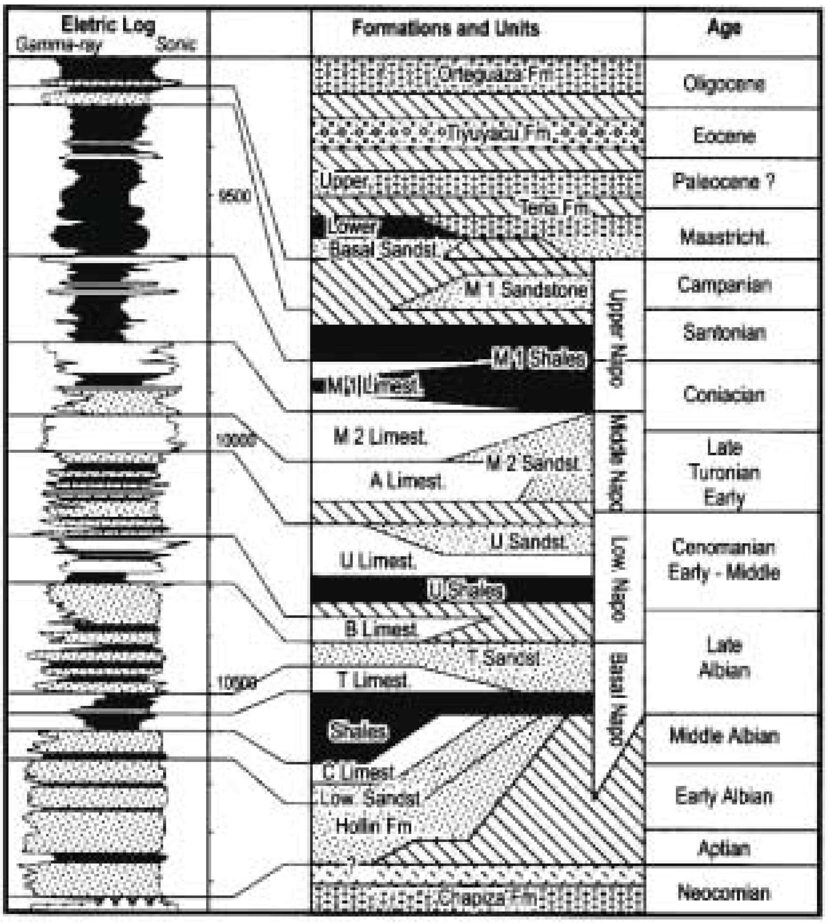
In the western part of the back-arc areas of Ecuador, Peru and Bolivia, the end of the carbonate platform sedimentation marks the Turonian-Coniacian boundary. In the N, it is replaced by ammonite-rich marine shale with limestone interbeds (Celendín, Upper Napo, Upper Chonta formations, Tschopp, 1953; Benavides, 1956; Wilson, 1963; Jaillard, 1987, 1997; Figs. 21 and 22), whereas in the S, the Turonian limestone units are overlain by red shale and silt with abundant evaporite layers (Chilcane, Aroifilla formations, Vicente, 1981; Sempere *et al.*, 1997; Figs. 13 and 21). In Ecuador and northern Peru, two main marine transgressions are recognized, of Early Coniacian and Late Coniacian-Early Santonian age, respectively. They determine two thickening-upward progradational sequences, of Coniacian and Early Santonian age, respectively (Jaillard, 1997). The appearance of detrital quartz in the Coniacian sequence indicates the creation of new source areas. No Late Santonian fauna has been found so far in these sequences.

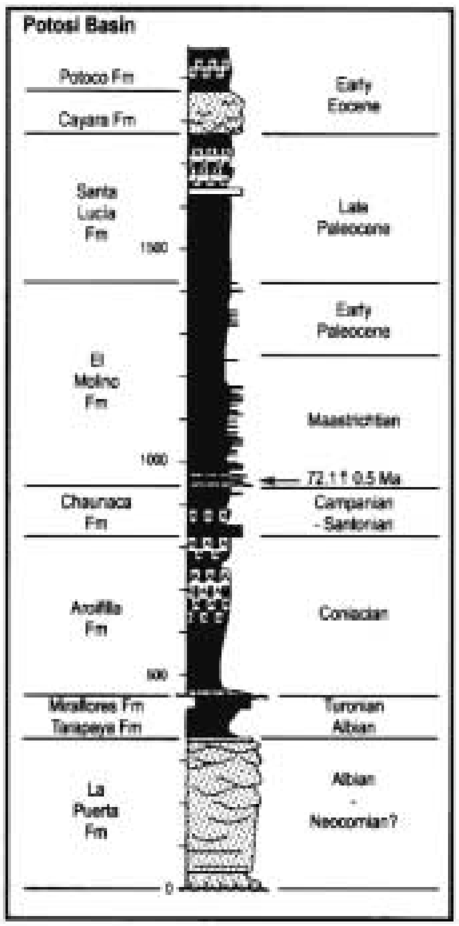
FIGURE 21 - Correlation of representative Albian-Eocene stratigraphic successions of Peru and Bolivia (after Jaillard and Soler, 1996).

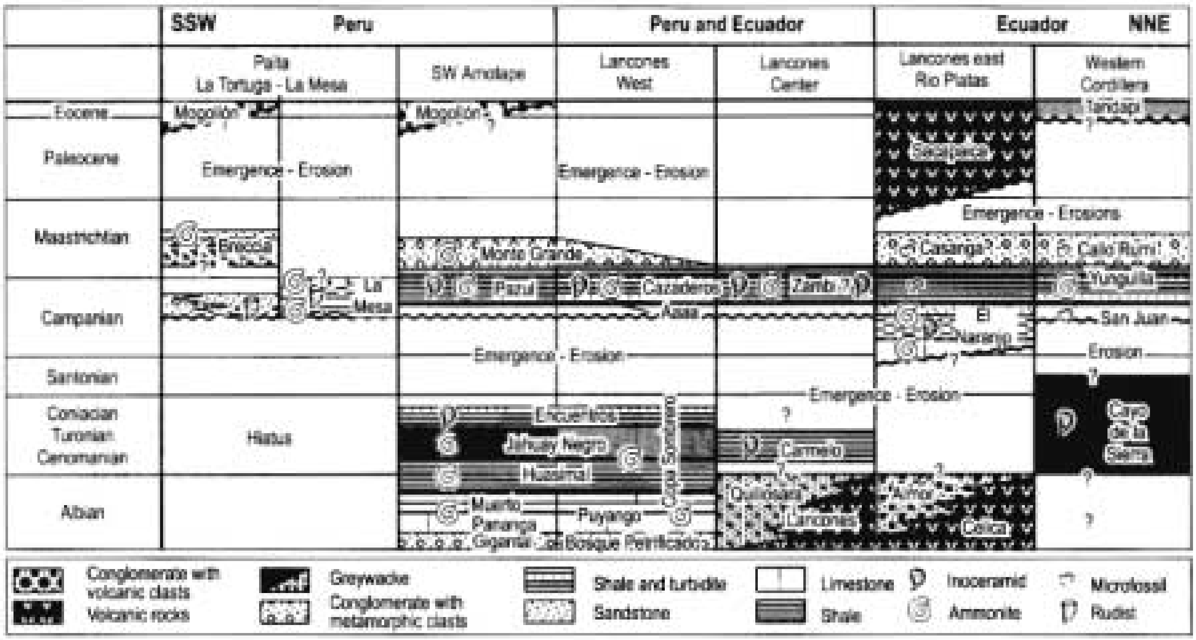
FIGURE 22 - Representative well log and chronostratigraphic chart of the Cretaceous succession of the Oriente Basin (eastern Ecuador; after Jaillard, 1997).

FIGURE 23 - Representative section of the Cretaceous series of the Potosi Basin (central Bolivia; after Sempere et al., 1997).









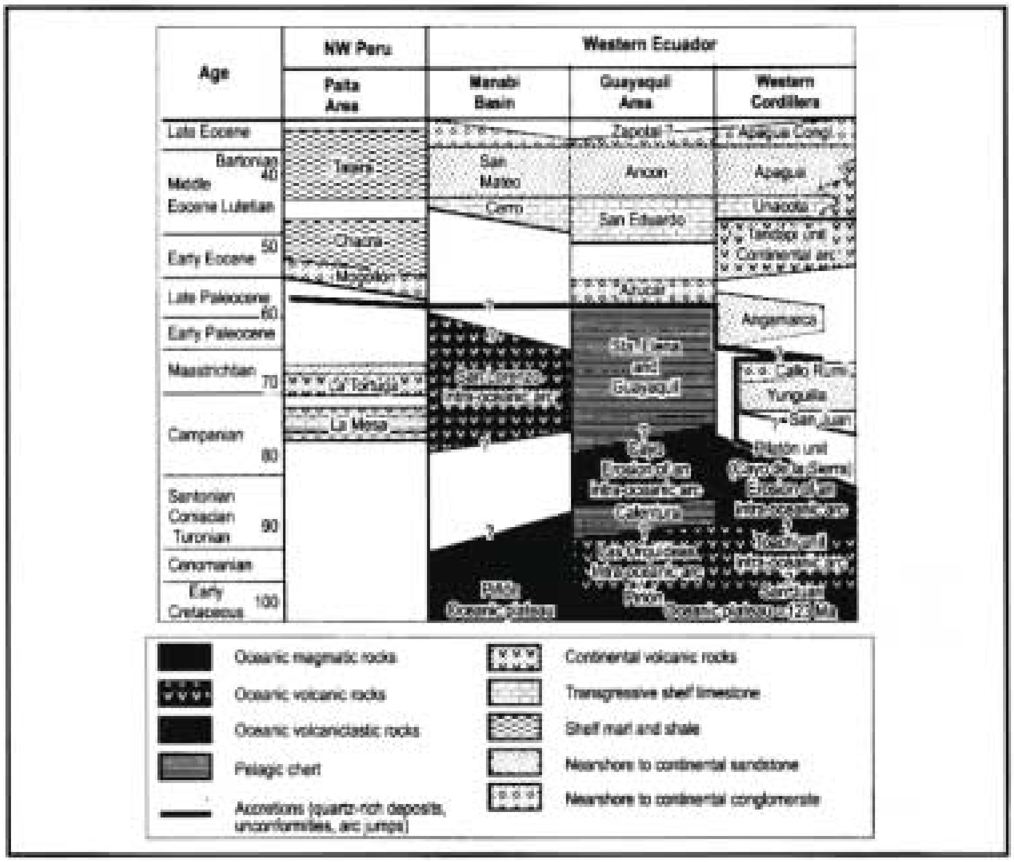


FIGURE 24 - Chronostratigraphic chart of the Middle Cretaceous to early Eocene sedimentary units of the fore-arc zones of Ecuador and northern Peru (after Jaillard et al., 1999).

FIGURE 25 - Chronostratigraphic chart of fore arc stratigraphic successions of northern Peru and western Ecuador. Probable accretion episodes are shown by the black line separating oceanic magmatic and sedimentary rocks (dark colours) from quartz-rich sediments and continental arc volcanism (light colours) (after Cosma et al., 1998).

Since the Coniacian to Early Santonian marine shale is generally overlain by transgressive sandstone of Campanian age, a sedimentary gap of Late Santonian-Early Campanian age is inferred. In southern Peru, and probably in Bolivia, the Late Coniacian-Early Santonian transgression is well marked and forms a thin marine layer (Middle Querque; Vicente, 1981; Jaillard, 1994), which is used as a correlation layer (Middle Vilquechico, Chaunaca formations, Jaillard *et al.*, 1993; Sempere *et al.*, 1997; Carlotto, 1998; Fig. 23).

In the back-arc areas, subsidence significantly increased during the Coniacian-Early Santonian time-span. In Bolivia, the spectacular increase in subsidence is regarded as the result of a foreland-type, flexural subsidence (Sempere, 1994; Sempere *et al.*, 1997; Fig. 23), due to significant tectonic shortening in the western areas (northern Chile). Extension prevailed, however, in these back-arc areas (Soler and Sempere, 1993). The occurrence of disoxigenated deposits (northern Peru, Ecuador) and of evaporites (southern Peru, Bolivia) suggests that the back-arc basin was separated from the open sea by an incipient morphological barrier, which was more pronounced to the S. In the whole area, Coniacian-Santonian marine deposits onlap eastwards onto the Guiana and Brazilian shields (Sempere, 1994). In eastern Ecuador, northeastern and southern Peru and in eastern Bolivia, the Coniacian-Santonian beds are the first Cretaceous marine shales to be deposited, and overlie the Albian (?)-Turonian fluvio-marine massive sandstone units (Jaillard, 1995, 1997; Sempere *et al.*, 1997). This significant eastward migration of the early Senonian depocenter is associated with a reorganization of the isopach maps, which become narrower and elongated parallel to the present-day chain, suggesting that the back-arc basins began to behave as distal foreland basins.

This Late Turonian-Early Coniacian paleogeographic reorganization is associated with local tectonic manifestations. In Bolivia, continental red beds unconformably overlie Middle Cretaceous marine strata (Vilcapujio event, Sempere, 1994) and in Ecuador (Jaillard, 1997), Coniacian silt or shale disconformably overlie the eroded Turonian limestone. In the Oriente Basin of Ecuador, the Late Turonian-Coniacian deposits exhibit significant thickness variations related to syn-sedimentary faulting of Late Turonian-Coniacian age (Christophoul *et al.*, 1999). In this area and in the Eastern Basin of northern Peru, mild compressional deformation has been recognized (Dashwood and Abbots, 1990; Gil *et al.*, 1996; Rivadeneira and Baby, 1999). This, together with the change in sedimentation and paleogeography, the slight retreat of the magmatic arc and the increase in subsidence, indicate that tectonic deformation and mild shortening affected the western areas.

Santonian-Early Campanian tectonic event and Campanian-Middle Maastrichtian evolution (85 - 68 Ma)

The Santonian (Early Campanian?) event is a major turning point in the Andean evolution, recognized a long time ago as the Peruvian phase (Steinmann, 1929). In the fore-arc Celica-Lancones-Basin (northern Peru-southern Ecuador), diachronous, latest Santonian to Middle Campanian transgressive shelf sediments unconformably overlie the deformed turbidite series of pre-Santonian age deposits (Jaillard *et al.*, 1997, 1999; Fig. 24). The compressional closure of the basin seems to be associated with the intrusion of syn-tectonic gabbro (Reyes and Caldas, 1987), locally dated at 82 Ma (Mourier, 1988). In the Talara fore-arc basin of northwestern Peru, the Albian shelf carbonate units are also covered disconformably by Campanian transgressive marine deposits (González, 1976; Macharé *et al.*, 1986; Séranne, 1987; Morales, 1993). In the fore-arc zone of Paita (northern Peru), Middle Campanian transgressive deposits rest unconformably on the Paleozoic basement (Bengtson and Jaillard, 1997; Jaillard *et al.*, 1999). Therefore, the main deformation of these fore-arc zones occurred during the latest Coniacian-Early Campanian time-span.

In the Celica-Lancones Basin, the middle Campanian transgressive beds are overlain by basinal dark shale interbedded with fine-grained turbidite beds of Late Campanian-Early Maastrichtian age (Jaillard *et al.*, 1999). South of Paita, the Middle Campanian transgressive sequence consists of transgressive marlstone and sandstone, rudist-bearing massive limestone, and transgressive marl and limestone grading upwards into sandstone and conglomerate, suggesting the occurrence of a Late Campanian tectonic event (La Mesa, Bengtson and Jaillard, 1997). Farther to the W (La Tortuga), the succession follows with a 3000 to 4000 m-thick series of alluvial to marine breccia, overlain by transgressive nearshore sandstone containing ammonites of Maastrichtian (probably Middle Maastrichtian) age (Bengtson and Jaillard, 1997; Fig. 24). These are unconformably overlain by latest Paleocene-early Eocene coarse-grained conglomerate, suggesting that a new tectonic event deformed this area in the Late Maastrichtian or Paleocene. No information is available about the other fore-arc zones.

In Santonian-Early Campanian times, the Ecuadorian margin underwent the accretion of an oceanic terrane constituted by an oceanic plateau dated at 123 ± 12 Ma (Lapierre *et al.*, 1999; Reynaud *et al.*, 1999) overlain by intra-oceanic island arc series (Fig. 25). This event is marked by a regional hiatus of Campanian age on the continental margin, by a significant thermal event which affected the Eastern Cordillera of Ecuador around 85 - 80 Ma (Cordillera Real, Litherland *et al.*, 1994) and by the abrupt arrival of disconformable quartz-rich turbidites of Late Campanian (?)-Maastrichtian age on the accreted oceanic series (Yunguilla Formation; Faucher *et al.*, 1971; Kehrer and Van der Kaaden, 1979; Cosma *et al.*, 1998). In the oceanic domain, the collision led to the end of the Middle Cretaceous island arc activity (Cayo Formation, Benítez, 1995), and to the onset, farther W, of a new island arc of Late Campanian-Maastrichtian age (San Lorenzo Formation, Lebrat *et al.*, 1987; Ordoñez, 1996). Since the accreted island arc series is locally dated as Coniacian in the Western Cordillera (Faucher *et al.*, 1971), the accretions occurred between the Coniacian and the Late Campanian. This arc jump expresses a significant reorganization of the intra-oceanic subduction zone geometry (Cosma *et al.*, 1998). The accreted oceanic terrane (Pallatanga unit, McCourt *et al.*, 1998), characterized by its association with the Yunguilla Formation, crops out presently along the eastern edge of the Western Cordillera of central and northern Ecuador (San Juan-Pujilí Suture, Juteau *et al.*, 1977; McCourt *et al.*, 1998).

The Early Maastrichtian Yunguilla Formation (Bristow and Hoffstetter, 1977) consists of alternations of basinal shale and medium-grained turbidite beds reworking volcaniclastic and siliciclastic material. These locally overlie units of transgressive limestone of Late Campanian-Maastrichtian age (Kehrer and Kehrer, 1969). This succession, comparable to that of the Celica-Lancones and Paita areas, indicates the creation of a wide fore-arc basin of Middle Campanian-Middle Maastrichtian age (Fig. 24), which extended at least from the Paita area (5°S) to N of Quito (0°).

The Albian volcanic arc series of southern Ecuador are unconformably capped by Late Santonian-Early Campanian transgressive marine deposits (Naranjo Formation, Jaillard *et al.*, 1997), which allow refining the age of the main tectonic event as pre-Campanian. In southern Peru, the Albian volcanics are unconformably capped by the undated Omoye Formation (Vicente, 1981), which has been ascribed to the Santonian (Jaillard, 1994), although it might be younger (Campanian?). In the arc zones, the effects of the Late Albian, Turonian-Coniacian and Late Santonian deformation are, therefore, indistinguishable. In both areas, the transgressive sequence grades into coarser-grained, locally conglomeratic, nearshore to continental deposits, dated in southern Ecuador as Maastrichtian (Cosanga Formation, Baudino, 1995; Jaillard, 1997), which indicate new tectonic movements in the Maastrichtian.

The Santonian-Early Campanian event coincided with the beginning of a significant retreat of the Coastal Batholith of Peru (Soler and Bonhomme, 1990; Fig. 20). This, together with the subsidence of the Late Campanian-Maastrichtian fore-arc basin of northern Peru-southern Ecuador, suggests that tectonic erosion began to act as a significant mass transfer process in the fore-arc and arc zones at that time (Jaillard and Soler, 1996; Jaillard, 1997). The Late Santonian-Early Campanian event is followed by a major plutonic pulse in the Coastal Batholith of central Peru, during which mainly granodiorite bodies were emplaced (85 - 77 Ma episode of Soler, 1991). A probable magmatic gap (77 - 74 Ma) might correspond to the Late Campanian event, and is followed by a new magmatic episode (74 - 69 Ma), which began with dyke swarm emplacement (Soler, 1991). In southern Peru, a plutonic gap (84 - 70 Ma) may coincide with the Santonian and Late Campanian events. The latter are responsible for the major NE-vergent Lluta Thrust, near Arequipa, which resulted in the thrust of Precambrian rocks onto Cretaceous sediments (Vicente *et al.*, 1982). Since it involves Coniacian-Early Santonian beds and is concealed by latest Cretaceous-early Paleogene unconformable conglomerate beds, it is of Late Cretaceous age (Vicente, 1989).

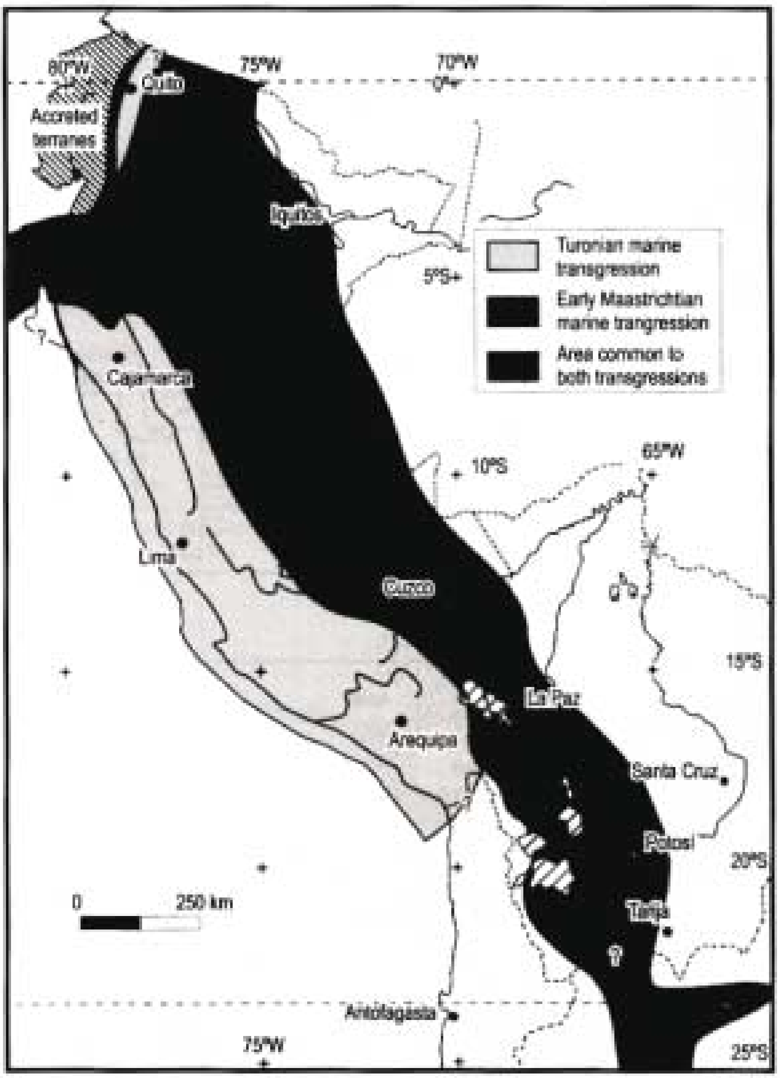
In northern Chile, magmatic activity resumed around 80 Ma ago (Early Campanian, Hammerschmidt *et al.*, 1992), and the location of the magmatic arc significantly shifted eastward, thus indicating that the Middle and Late Cretaceous tectonic events resulted in significant crustal shortening and/or crustal erosion (Scheuber *et al.*, 1994). This significant contractional event dated as 90 - 78 Ma, resulted in the folding, emergence and erosion of the arc zone, the tectonic inversion of the Domeyko Cordillera and creation of the retro-arc Purilactis Basin (Mpodozis and Ramos 1989; Scheuber *et al.*, 1994).

In the back-arc areas of Ecuador and northern Peru, Late Santonian-Early Campanian times are marked by a regional sedimentary gap (Tschopp, 1953; Benavides 1956; Seminario and Guizado 1976; Jaillard 1987, 1997; Mathalone and Montoya, 1995), which coincides with the accretion and related deformations recorded in the westerly zones. In northern Peru and eastern Ecuador, the Santonian marine deposits exhibit a thickening-upward evolution expressing the arrival of sandy detrital material regarded as related to the incoming Late Santonian tectonic movements. In southern Peru and Bolivia, stratigraphic data are insufficient to demonstrate the occurrence and duration of this hiatus in the mostly continental deposits (Middle Vilquechico, Middle Yuncaypata, Chaunaca formations; Sempere *et al.*, 1997; Jaillard *et al.*, 1993; Carlotto, 1998). The Early Santonian age of the last marine deposits in northern Peru and Ecuador, however, supports a Late Santonian age for the main deformational event.

Campanian times are then marked by a short-lived, regional marine transgression, locally dated as Middle Campanian (northern Peru, Mourier *et al.*, 1988), and therefore, probably coeval with the main transgression in the fore-arc zone (Fig. 24). In Ecuador and northeastern and central Peru, this transgression is associated with conspicuous disconformable transgressive sandstone (M-1 Sandstones, Lower Vivian Formation) overlain by a thin layer of marine shale (Augusto *et al.*, 1990; Salas, 1991; Mathalone and Montoya, 1995; Jaillard, 1997). In southern Peru and Bolivia, the Middle Campanian transgression is correlated with a thin layer of charophyte-bearing shale overlain by fine-grained red beds of presumed Late Campanian age (Middle Vilquechico, Middle Yuncaypata, Upper Chaunaca formations, Jaillard *et al.*, 1993; Sempere *et al.*, 1997; Figs. 23 and 27). The hiatus between Campanian and Maastrichtian deposits suggests the occurrence of a tectonic event in the Late Campanian, but an eustatic origin for this sedimentary gap cannot be ruled out. Mid-Campanian alkaline volcanic rocks (80 - 75 Ma) point to an extensional strain in northern Argentina (Viramonte *et al.*, 1999).

A new regional marine transgression occurred in the Early Maastrichtian, which deposited transgressive sandstone units grading upwards into marine shale, which rest disconformably on the Campanian beds. In Ecuador and northern and central Peru, these Early Maastrichtian marine layers are dated by marine microfossils and very scarce ammonites (Lower Tena, Upper Vivian, Areniscas de Azúcar formations, Koch and Blissenbach, 1962; Rodríguez and Chalco, 1975; Vargas, 1988; Mourier *et al.*, 1988; Jaillard, 1997). They are generally overlain by charophyte-bearing fine-grained continental red beds of Maastrichtian age. A disconformity separates these deposits from the overlying fine-grained, continental Paleocene red beds (Upper Tena, Yahuarango, Sol formations), suggesting the occurrence of tectonic movements near the Maastrichtian-Paleocene boundary (Mathalone and Montoya 1995; Jaillard, 1997; Christophoul, in progress).

In southern Peru and Bolivia, the Early Maastrichtian maximum flooding is marked by ephemeral marine



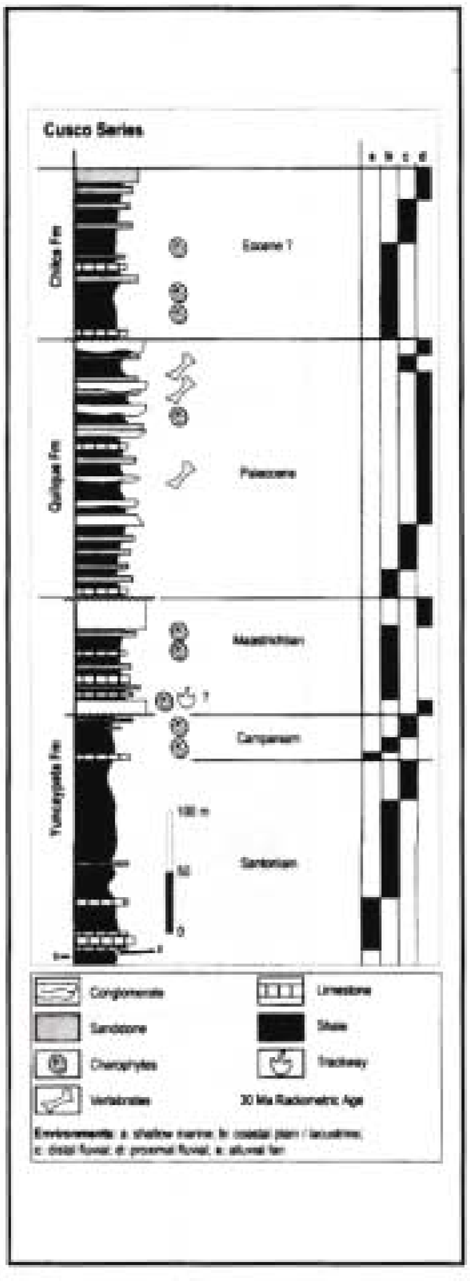


FIGURE 26 - Paleogeographic sketch showing the relative extensions of the Turonian and Maastrichtian marine transgressions.

FIGURE 27 - Late Cretaceous - Eocene stratigraphic succession in the Cusco area (Andes of southern Peru) (after Carlotto, 1998).

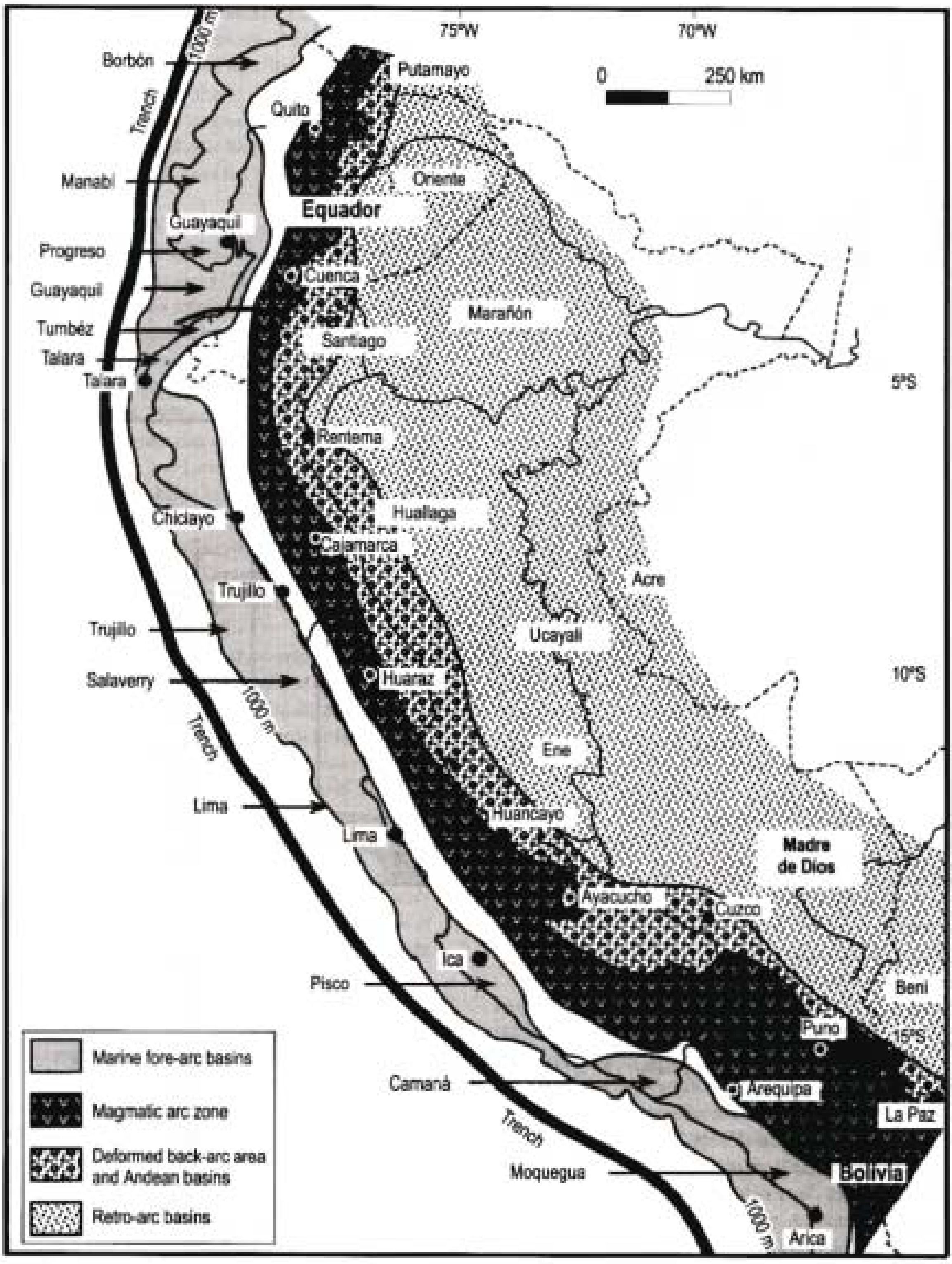


FIGURE 28 - Paleogeographic sketch of Ecuador and Peru for the Paleogene.

conditions (Jaillard *et al.*, 1993; Gayet *et al.*, 1993) and the Maastrichtian deposits significantly onlap onto the eastern border of the Cretaceous basin (Sempere *et al.*, 1997), thus expressing the gradual eastward shift of the depocenter during the Late Cretaceous (Fig. 26). Lacustrine conditions prevailed in part of the Bolivian-Argentine part of the basin (Rouchy *et al.*, 1993; Salfity and Marquillas, 1994). In southern Peru, an erosional disconformity separates the Maastrichtian and Tertiary beds (Jaillard *et al.*, 1993; Carlotto, 1998), but sedimentation is assumed to be continuous in Bolivia (Sempere *et al.*, 1997). The tectonic regime in the back-arc areas of northern Chile and Bolivia is assumed to have been extensional during the Maastrichtian, thus allowing sporadic shallow marine incursion and the outpouring of alkaline basic volcanics (Reyes *et al.*, 1976; Scheuber *et al.*, 1994; Viramonte *et al.*, 1999).

Late Maastrichtian - early late Paleocene (68 - 57 Ma)

The early to early late Paleocene sequence disconformably overlies latest Cretaceous strata. It consists of volcanic rocks (arc zone), and of fine-grained deposits, either marine (fore-arc zone) or continental (back-arc zone). In the arc and fore-arc zones, beside the regional hiatus of Late Maatrichtian-early Paleocene age and the frequent disconformities between Cretaceous and Tertiary beds, a Late Maastrichtian tectonic event is suggested by numerous intrusions (Mukasa, 1986), the emplacement of a centred complex (Mukasa and Tilton, 1985) and high strike-slip rates between 68 and 64 Ma in the Coastal Batholith of Peru (Bussel and Pitcher, 1985), and by numerous K/Ar resets indicating a thermal event at 70 - 60 Ma in the Cordillera Real of Ecuador (Litherland *et al.*, 1994).

Paleocene marine deposits are only known in the fore-arc Talara Basin of northwesternmost Peru (Iddings and Olsson, 1928; González, 1976; Zuñiga and Cruzado, 1979; Séranne, 1987; Morales, 1993; Fig. 29) and maybe in the off-shore fore-arc zone of northern Peru, where the presence of pre-late Eocene sediments has been assumed locally (9°S, Kulm *et al.*, 1982). In the Talara Basin, the Late Cretaceous deposits are disconformably overlain by early Paleocene transgressive sandstone and conglomerate intercalated with marine grey shale, which grade upward into marine dark shale of middle Paleocene age (Weiss, 1955; Paredes, 1958; González, 1976). The upper unit overlaps to the E (Séranne, 1987), and is strongly eroded toward the S by pre-Eocene erosion (Paredes, 1958). To the E, thin continental to nearshore clastic facies grade westward into thick fine-grained deposits of relatively deep marine environment (Séranne, 1987). Farther to the S (Paita), the Cretaceous sedimentation ends up with the 3500 m thick breccia of the Maastrichtian La Tortuga Formation (Olsson, 1944; Fig. 24).

In the presently accreted oceanic terranes of Ecuador, since the San Lorenzo island-arc did not yield ages younger than Maastrichtian (Lebrat *et al.*, 1987; Ordoñez, 1996), its activity may have ceased by Paleocene times. Moreover, McCourt *et al.* (1998) recently identified an early to middle Paleocene quartz-rich turbidite series resting on accreted oceanic island arc series in the Western Cordillera of Ecuador. Since these early Paleocene turbidite beds are found to the W of the outcropping belt of the Yunguilla Formation, they might indicate that an other fragment of oceanic terrane has been accreted during the Late Maastrichtian and/or the earliest Paleocene. However, in some parts, the Maastrichtian-Paleocene sedimentation continued without any noticeable changes (Guayaquil Formation, Benítez, 1995; Keller *et al.*, 1997; Fig. 25).

In the magmatic arc of southwestern Ecuador, subaerial andesitic lava, breccia, agglomerate and acid tuff (Sacapalca Formation) are of latest Maastrichtian (67 Ma) to early Eocene age (Jaillard *et al.*, 1996; Hungerbühler, 1997; Pratt *et al.*, 1998), indicating that volcanic arc activity resumed after a gap that lasted from Late Albian times. In central and northern Ecuador, volcanic arc activity did not start before the early Eocene. In central and northern Peru, ring-complexes (68 - 64 Ma, Cobbing *et al.*, 1981; Soler, 1991) and calc-alkaline intrusions were emplaced (64 - 59 Ma, Cobbing *et al.*, 1981; Beckinsale *et al.*, 1985) and were possibly associated with coeval volcanism (Pararín Formation, Bussel, 1983). These intrusions are volumetrically important, indicating the local resumption of magmatic arc activity, but no significant compositional changes is noted with respect to the Late Cretaceous magmatism (Soler, 1991; Fig. 20). The 68 - 64 Ma period is also marked by important dextral wrench movements (Bussel, 1983; Bussel and Pitcher, 1985), which may be the expression of a Late Maastrichtian tectonic event. A magmatic gap then occurred during the late Paleocene (59 - 54 Ma, Soler, 1991). In southern Peru, plutonic intrusions began during latest Cretaceous times (78 Ma) and exhibit a major pulse during the early to middle Paleocene (62 - 57 Ma, Beckinsale *et al.*, 1985; Mukasa, 1986; Clark *et al.*, 1990). Associated volcanism (Toquepala Formation) consists of 3000 m of dacitic to rhyolitic tuff with minor andesitic intercalations, the composition of which suggests that the Andean crust was not thickened (Boily *et al.*, 1990). It is crosscut by gabbroic and granitic intrusions dated mainly at 66 - 63 Ma (Laughlin *et al.*, 1968; Vatin-Pérignon *et al.*, 1982; Mukasa and Tilton, 1985; Clark *et al.*, 1990). The resumption of arc magmatism recognised in Ecuador and Peru is expressed in northern Chile by abundant Late Maastrichtian-early Paleocene ages in the magmatic arc rocks (Hammerschmidt *et al.*, 1992; Charrier and Reutter, 1994), and by coeval volcanic intercalations in red beds deposited in proximal back-arc basins (Purilactis Group, 64 Ma, Flint *et al.*, 1993).

In the proximal back-arc zones of Peru (present-day Andes), Paleocene deposits unconformably rest on Late Cretaceous rocks (Noble *et al.*, 1990; Jaillard *et al.*, 1993; Mégard *et al.*, 1996), whereas the contact is only locally disconformable in the Eastern Basin (Vargas, 1988; Augusto *et al.*, 1990; Salas, 1991; Mathalone and Montoya, 1995; Gil *et al.*, 1996; Figs. 27 and 31). In addition, compressional deformation due to tectonic inversions near the Cretaceous-Tertiary boundary are common and widespread in the back-arc areas of eastern Ecuador (Dashwood and Abbots, 1990; Rivadeneira and Baby, 1999; Christophoul, in progress), northeastern Peru (Contreras *et al.*, 1996; Gil *et al.*, 1996) and Colombia (Cheilletz *et al.*, 1997).

In the distal back-arc areas of Ecuador and Peru, the Late Campanian-Maastrichtian sequence is overlain by a thick series of Paleocene fine-grained red beds (Upper Tena, Yahuarango, Sol 1, Quilque, Chilca, Santa Lucia formations, Kummel, 1948; Koch and Blissenbach, 1962; Mathalone and Montoya, 1995; Jaillard, 1997; Sempere *et al.*, 1997; Christophoul, in progress) which wedges out toward the W, mainly because of pre-Eocene erosion (Naeser *et al.*, 1991; Jaillard *et al.*, 1993; Carlotto, 1998). In the western zones, Paleocene deposits are usually lacking. However, in the Andes of Central Peru, a series of fluviatile red beds has been assigned to the Paleocene (Casapalca Formation, Jacay, 1994), although it may be younger. These Paleocene fine-grained red beds were deposited in wide, distal alluvial plains or in coastal setting. Clastic material proceeded from the smooth relief of the Paleo-Andes. Microfossils are dominated by charophyte associations (Gutierrez, 1982; Jaillard, 1994), but scarce foraminifera indicate local and sporadic marine influences (Koch and Blissenbach, 1962). In northeastern Peru, the Paleocene beds are disconformably overlain by transgressive conglomerate and marine to brackish beds of Eocene age (Pozo Formation). In Bolivia and northern Argentina, extensional conditions are marked by early Paleocene K-rich lava flows (65 - 60 Ma, Viramonte *et al.*, 1999).

The widespread hiatus, unconformities and detrital sedimentation, as well as the deformation and thermal event suggest that a significant, although poorly known, tectonic event occurred near the Maastrichtian-Paleocene boundary. This event might correspond to the accretion of an oceanic terrane, since part of western Ecuador received early Paleocene quartz-rich sedimentation, and recorded the end of the activity of an island arc (arc jump).

Late Paleocene - late Oligocene

This period corresponds to a transition between the pre-orogenic and the syn-orogenic periods. Compressional deformations became significant and involved the western parts of the back-arc areas, where marine sedimentation no longer occurred, except locally in Ecuador. The subsidence of fore-arc zones, which follows the compressional events, created sedimentary basins. Detrital sedimentation in the eastern area shows evidence for tectonically-induced disconformities. Finally, activity of the volcanic arc resumed, including along the Ecuadorian margin where arc magmatism was unknown since Late Jurassic times.

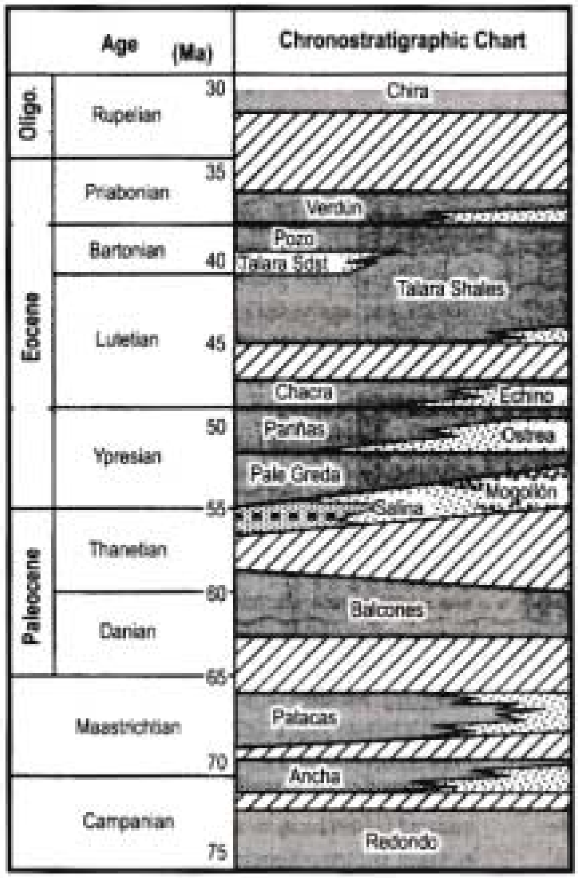
Because of the ongoing crustal shortening, eastward migration of the magmatic front, and uplift of the Andean domain, the paleogeographic pattern progressively changed during this period (Fig. 28). The fore-arc zones roughly correspond to the present-day coastal and offshore parts of the margin. Due to tectonic erosion, shortening, and/or flattening of the slab, the arc zone migrated and enlarged eastwards through time. It corresponded to the western part of the present-day Western Cordillera. The back-arc areas can be divided into a western, deformed and usually emergent area, also referred to as the "Paleo-Andes", and an eastern area, which still received sedimentation, and evolved through time toward a foreland retro-arc basin.

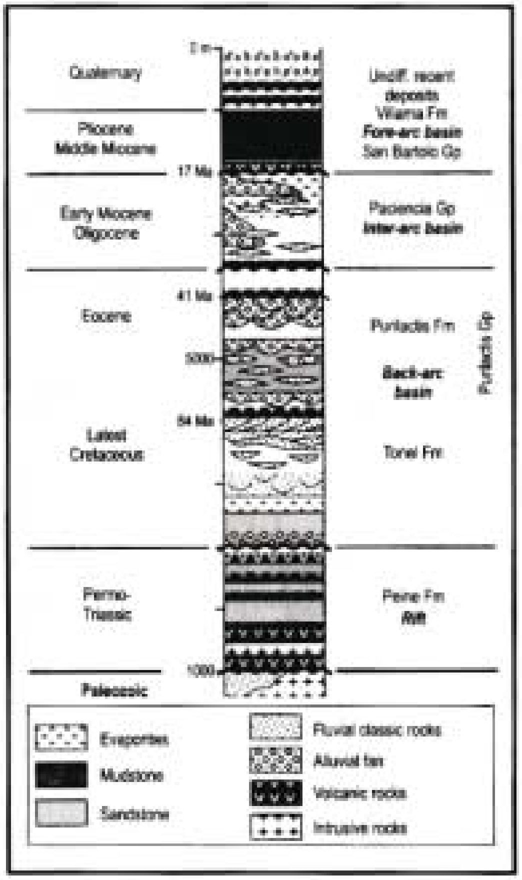
Late Paleocene event (58 - 55 Ma) and Eocene Sequence (55 - 40 Ma)

The late Paleocene event, first suspected by Cobbing *et al.* (1981) and Bussel and Pitcher (1985), is one of the major events in the Andean history (Marocco *et al.*, 1987; Noble *et al.*, 1990; Sempere *et al.*, 1997; Jaillard, 1997). It is coeval with an important plate kinematic reorganization in the Pacific realm, dated at 58 - 56 Ma, which resulted in a change in the convergence direction of the Farallón Plate. The latter changed direction from N or NNE to NE (Pilger, 1984; Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987; Atwater, 1989). The late Paleocene event is followed by the deposition of disconformable, well-identifiable sedimentary or volcanic sequences of early to early late Eocene age.

In the fore-arc zone of Ecuador, the collision of an oceanic terrane resulted in locally intense deformation of early late Paleocene chert (Santa Elena Formation) belonging to the accreted terrane (Benítez, 1995; Jaillard *et al.*, 1995). Since a thick series of quartz-rich coarse-grained turbidite beds of latest Paleocene age conceals the accretion, the latter occurred in the late Paleocene (Jaillard, 1997; Fig. 25). A further tectonic event of probable earliest Eocene age deformed the unconformable quartz-rich turbidite beds (Jaillard, 1997). In the Talara fore arc basin of northern Peru, sandstone and conglomerate (Basal Salinas) disconformably overlie Paleocene marine shale (Séranne, 1987) and grade southwards into diachronous, latest Paleocene to early Eocene, coarse-grained alluvial conglomerate (Mogollón Formation, Morales, 1993; Fig. 29). In the Paita area, the latter contain boulders of intra-oceanic origin, thus demonstrating that accretion occurred before the early Eocene in northern Peru. Since oceanic terranes are presently in Ecuador, they were subsequently displaced northwards along dextral faults, with a minimum estimate rate of 5 mm/y (Pecora *et al.*, 1999).

Due to the late Paleocene event, the fore-arc zones of Ecuador are marked by a widespread sedimentary hiatus encompassing most of the early Eocene (Benítez, 1995; Jaillard *et al.*, 1995). Sedimentation resumed diachronically since the end of the early Eocene. The Eocene sequence typically begins with breccia, slumped shale or transgressive peri-reefal limestone concealing fault-controlled relief. Diachronism and tectonic figures indicate an extensional tectonic regime, related to the subsidence that led to the deposition of the overlying thinning and shallowing-upwards sequence of marls interbedded with turbiditic sandstone (Jaillard *et al.*, 1995). The Eocene sequence ends up with disconformable, locally conglomeratic sandstone of nearshore to continental environment, dated as late middle to early late Eocene (Jaillard *et al.*, 1995). These indicate the beginning of the late Eocene tectonic movements (Jaillard, 1997). A comparable sequence is known in most of the coastal area (Benítez, 1995; Jaillard *et al.*, 1995) and the Western Cordillera (Santos *et al.*, 1986; Egüez, 1986; Bourgois *et al.*, 1990). This suggests that most of western Ecuador underwent a similar sedimentary evolution during the middle Eocene, and that most of oceanic terranes were probably already accreted to the continental margin by middle Eocene times (Jaillard, 1997; Cosma *et al.*, 1998). In the Western Cordillera, however, an island arc unit (Macuchi





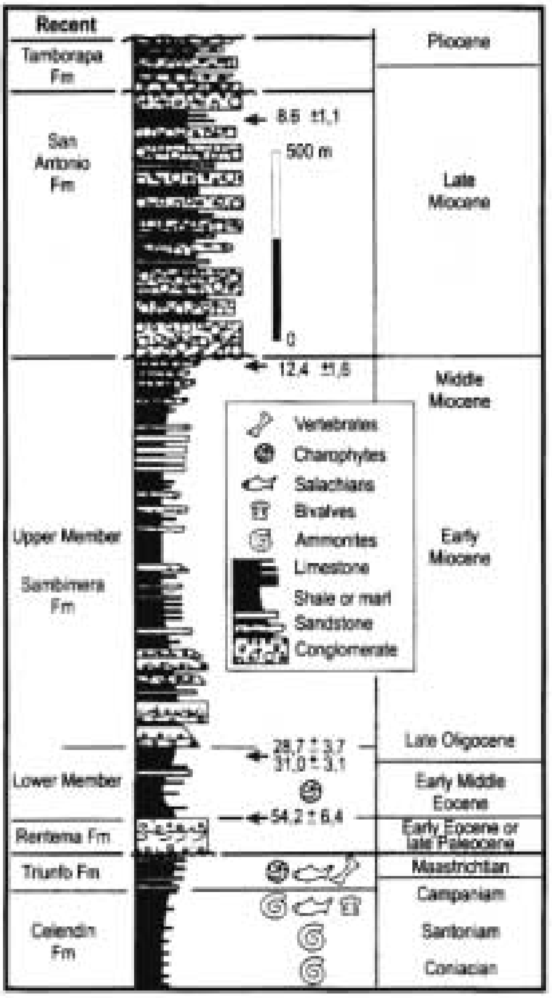


FIGURE 29 - Chronostratigraphic chart of the Late Cretaceous-Paleogene Talara Basin (forearc zone of northern Peru) (after Morales, 1993).

FIGURE 30 - Mesozoic-Quaternary stratigraphic succession in the Atacama area (northern Chile) (after Flint et al., 1993). Due to the eastward migration of the arc zone, this area evolved from a back-arc to a fore-arc setting between latest Cretaceous and Pliocene times.

FIGURE 31 - Late Cretaceous-Pliocene stratigraphic succession in the Bagua area (northern Peru) (after Naeser et al., 1991).

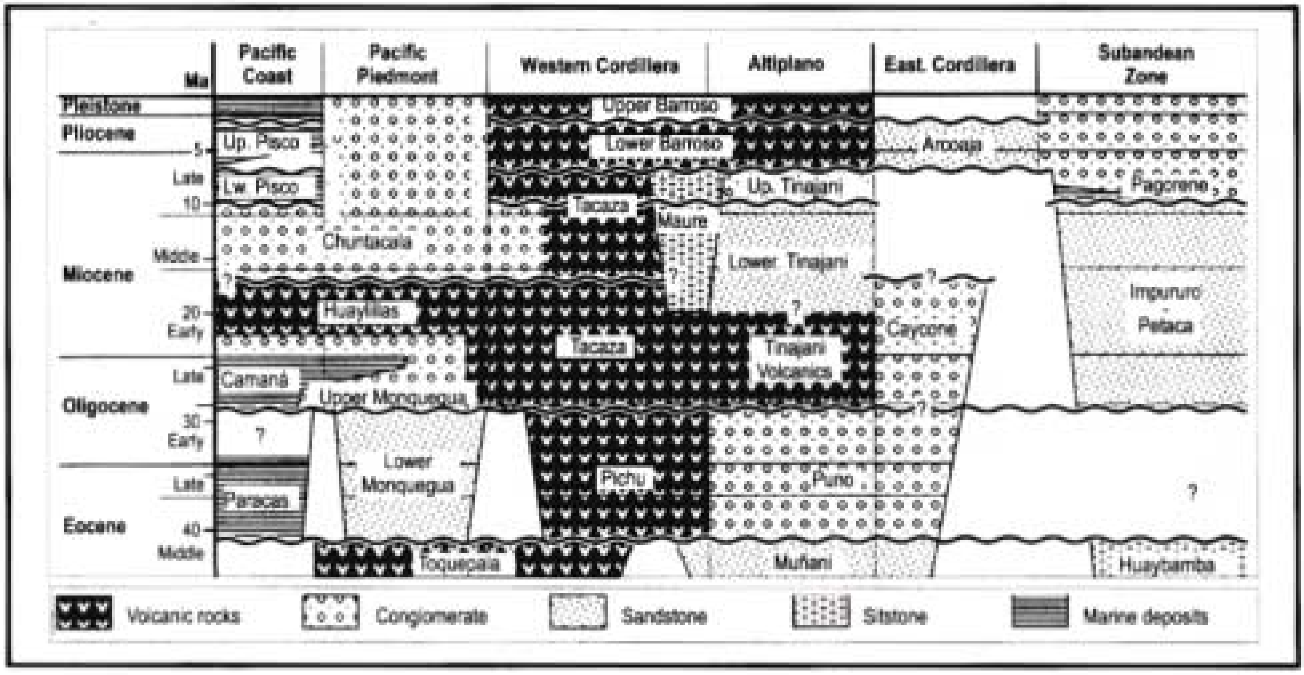


FIGURE 32 - Chronologic chart of the sedimentary and volcanic successions and deformational events in southern Peru (after Sébrier et al., 1988)

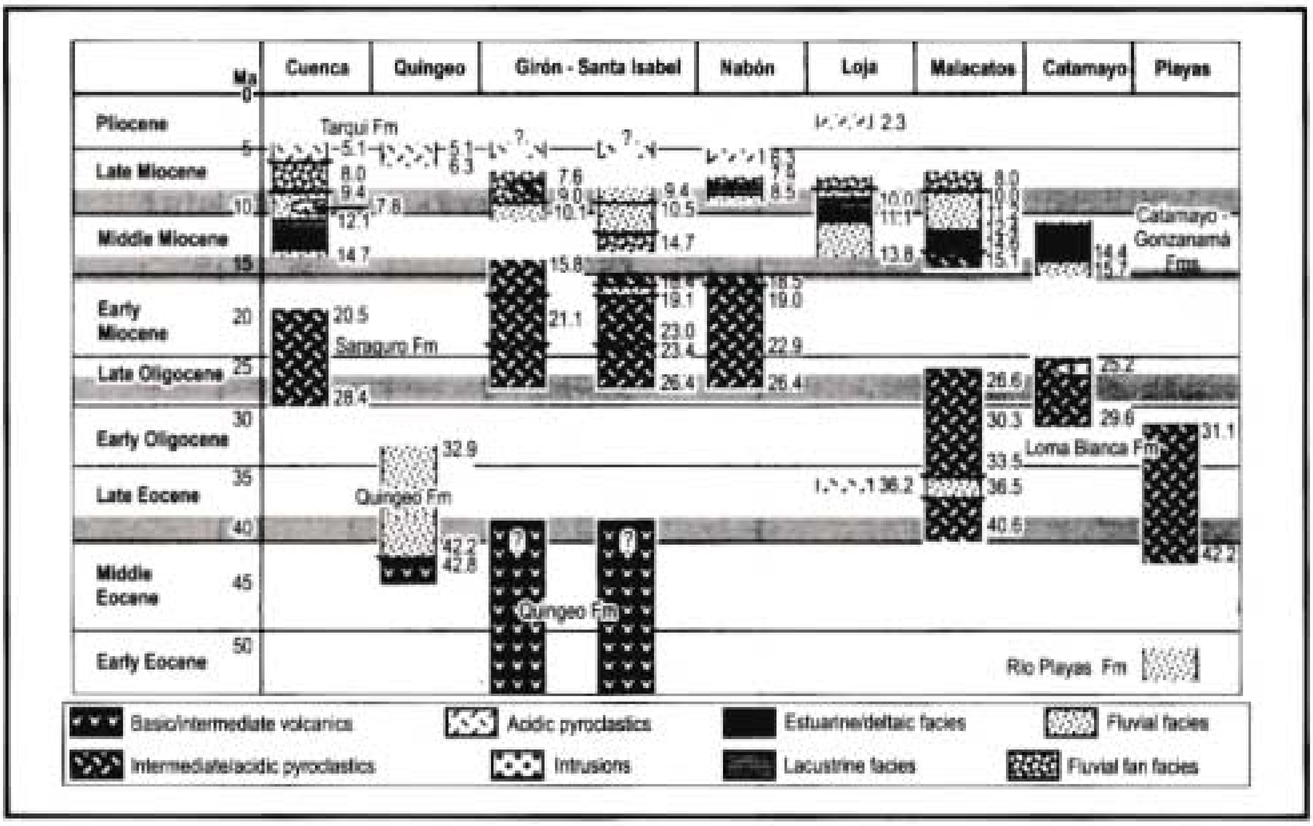


FIGURE 33 - Chronostratigraphic chart of the volcanic and sedimentary successions in the Andes of southern and Central Ecuador (after Hungerbühler, 1997, Steinmann et al., 1999).

Sandstone) seems to be associated with Eocene pelagic chert, and was therefore not yet in contact with the Andean margin (Hughes and Pilatasig, 1999).

In the fore-arc zone of Peru, although early Eocene deposits have been locally mentioned (Kulm *et al.*, 1982, Suess *et al.*, 1988), deposits of that age are well known only in the Talara Basin (Fig. 29). There, the very thick Eocene series comprises three main sedimentary sequences limited by disconformities of earliest Eocene, latest early to early middle Eocene, and late middle Eocene age, respectively (González, 1976; Séranne, 1987; Morales, 1993). The diachronous transgression begins with disconformable marine sandstone and conglomerate of latest Paleocene to early Eocene age, from W to E (Morales, 1993), the base of which contains boulders of oceanic origin, which post-date the accretion of oceanic terranes (Pécora *et al.*, 1999). The Eocene sequences generally consist of marine shale and fine to coarse-grained sandstone, which grade laterally E or NE into coarse-grained continental sandstone and conglomerate (González, 1976; Séranne, 1987). Pre-middle Eocene erosions removed part of the early Eocene sequence, and the second sequence locally rests on the basal conglomerate (Paredes, 1958). The upper part of the late middle Eocene sequence exhibits compressional syn-sedimentary deformations related to a transpressional regime (Séranne, 1987; Becerra *et al.*, 1990). The third sequence (late middle to early late Eocene) is made of unconformable massive coarse-grained sandstone of nearshore environment (González, 1976; Morales, 1993), announcing the late Eocene tectonic event. Overlying open marine shale beds are ascribed to the late Eocene (González, 1976) or the early Oligocene (Morales, 1993).

Near the early-middle Eocene boundary, subsidence of the fore-arc zone triggered the creation of a new generation of fore-arc basins, where thick middle to early late Eocene deposits unconformably rest upon Paleozoic to early Eocene units. In the offshore fore-arc basins of central and northern Peru, drillholes crosscut as much as 2000 m of middle Eocene shale, siltstone and sandstone of shallow marine, high-energy environment (Ballesteros *et al.*, 1988; Suess *et al.*, 1988), which usually unconformably overlies Cretaceous to Paleozoic rocks (Macharé *et al.*, 1986). It ends up locally with breccia, suggesting a tectonic instability of late middle Eocene to late Eocene age. Subsidence and basin tectonics is controlled by NNE trending normal faults (Macharé *et al.*, 1986; Azálgara *et al.*, 1991). The middle Eocene series is commonly directly overlain by Miocene marine sediments, thus providing evidence for a widespread sedimentary hiatus encompassing late Eocene, Oligocene, and often early Miocene times (Suess *et al.*, 1988; Von Huene *et al.*, 1988).

In the arc zone of southern Ecuador, subaerial arc volcanism and associated continental volcaniclastic sedimentation occurred during the late Paleocene and early Eocene (Sacapalca Formation, Jaillard, 1997; Hungerbühler, 1997). In the rest of Ecuador, resumption of arc volcanism is dated as early or middle Eocene (53 - 45 Ma, Egüez 1986; Wallrabe-Adams, 1990; Van Thournout *et al.*, 1990; Steinmann, 1997; Dunkley and Gaibor, 1998). It is commonly associated with subaerial volcaniclastic red beds of middle Eocene age deposited in proximal back-arc basins (Silante Formation, Wallrabe-Adams, 1990; Hughes and Pilatasig, 1999; Quingeo Formation, Steinmann, 1997; Playas Formation, Hungerbühler, 1997; Fig. 33). The renewal of arc activity along the Ecuadorian margin is due to the more easterly convergence direction, subsequent to the late Paleocene plate tectonic re-organization (Pilger, 1984; Pardo-Casas and Molnar, 1987).

In northern Peru, thick calc-alkaline subaerial volcanic series are dated between 55 and 40 Ma (Laughlin *et al.*, 1968; Cobbing *et al.*, 1981; Noble *et al.*, 1990; Soler, 1991), and post-date compressional deformation (Cobbing *et al.*, 1981; Bussel, 1983; Bussel and Pitcher, 1985). Although the chemical signature did not change with respect to the Paleocene magmatic arc (Soler, 1991), the Eocene magmatic activity is marked by a slight eastward shift of the western magmatic front, a decrease of plutonic intrusions in the Coastal Batholith and the beginning of the enlargement of the magmatic arc, which reached the present-day Western Cordillera (Noble *et al.*, 1990; Soler, 1991; Fig. 20). In contrast, a magmatic gap occurred in southern Peru during the early Eocene (Mukasa, 1986; Boily *et al.*, 1990; Clark *et al.*, 1990; Soler, 1991). However, subvolcanic stocks and associated porphyry copper deposits were emplaced in the Toquepala prospect between 57 and 52 Ma (Sébrier *et al.*, 1988; Clark *et al.*, 1990). Farther to the E (Inner Arc), the enlargement of the magmatic arc is marked by the emplacement of several calc-alkaline plutons, among which the large is the Andahuaylas-Yauri Batholith (48 - 34 Ma, Carlier *et al.*, 1996).

In northern Chile, the magmatic arc drastically shifted eastward between 55 and 48 Ma (Hammerschmidt *et al.*, 1992; Scheuber *et al.*, 1994), thus suggesting that significant crustal shortening and/or tectonic erosion occurred in the early Eocene (Fig. 15). This event has been ascribed to the late Eocene tectonic phase (Scheuber *et al.*, 1994), but is better correlated with the late Paleocene event. The latter is followed by a significant resumption of the magmatic activity along the whole margin. In northern Chile, the resumption of arc magmatism is expressed by numerous volcanic outcrops dated as middle Eocene (48 - 38 Ma, Hammerschmidt *et al.*, 1992). Products of this arc were deposited in a proximal, extensional back-arc basin, which received a thick, coarsening-upward pile of mainly volcaniclastic rocks, which may rest conformable on Late Cretaceous sediments or unconformably on older rocks (Hartley *et al.*, 1992; Flint *et al.*, 1993; Purilactis Formation of Charrier and Reutter, 1994; Fig. 30). Tectonic regime in the back-arc zone is no longer extensional (Scheuber *et al.*, 1994).

In the western back-arc areas of Ecuador (Cordillera Real), K/Ar age resets near 65 - 50 Ma indicate the occurrence of a noticeable thermal event (Aspden and Litherland, 1992; Litherland *et al.*, 1994), related to the late Paleocene tectonic event. In the Andes of northern Peru, early Eocene volcanic rocks dated at 55 to 50 Ma unconformably rest on deformed Cretaceous sediments (Cobbing *et al.*, 1981; Bussel, 1983; Noble *et al.*, 1990; Soler, 1991). In the Andes of central Peru, continental red beds bearing Paleocene-Eocene charophytes unconformably rest on Cretaceous sediments (Mégard, 1978; Mégard *et al.*, 1996).

In the present-day Andes of central Peru, unconformable continental red beds are locally dated by latest Maastrichtian charophytes, whereas in other parts, apparently similar, but conformable, red beds yielded late

Eocene-early Oligocene charophyte oogons and 40 to 37 Ma K/Ar dates (Mégard *et al.*, 1996). Therefore, some red beds may be of late Paleocene-middle Eocene age, but their characteristics and extension are unknown so far. Undated fluviatile red beds (Casapalca Formation) overlying the latest Cretaceous strata have been ascribed to the Paleocene (Jacay, 1994), although they may be younger. In southern Peru, Paleocene fluviatile red beds (Quilque Formation) are disconformably overlain by lacustrine deposits (Chilca Formation, Jaillard *et al.*, 1993; Carlotto, 1998), possibly of Eocene age.

The eastern back-arc areas are marked by a regional unconformity below massive coarse-grained sandstone and conglomerate of early Eocene age (Tiyuyacu Formation of Ecuador, Dashwood and Abbots, 1990; Benítez *et al.*, 1993; Jaillard, 1997; Rivadeneira and Baby, 1999; Rentema and Basal Pozo formations of Peru, Naeser *et al.*, 1991; Robertson Research, 1990; Cayara Formation of Bolivia, Sempere, 1994; Sempere *et al.*, 1997; Figs. 13, 23, and 31). They often post-date a sedimentary gap, which encompasses a large part of the Paleocene. Moreover, no early Eocene deposits have been accurately dated so far in the back-arc areas of Ecuador and Peru. Early to middle Eocene times are then marked by a regional marine transgression.

In Ecuador, the unconformable Lower Tiyuyacu Formation is overlain by marine to brackish beds of Eocene age (Benítez *et al.*, 1993). In eastern Peru, the early Eocene basal transgressive lag is overlain by a marine to brackish fine-grained layer of Eocene age and by coarsening-upward lacustrine to fluviatile red beds of middle to late Eocene age (Pozo Formation, Kummel, 1948; Müller and Aliaga, 1981; Robertson Research, 1990; upper part of Sol 3 Formation, Koch and Blissenbach, 1962; Gutierrez, 1982). The Ucayali Basin seems to be marked by a sedimentary hiatus of early Eocene age (Koch and Blissenbach, 1962). On the western border of the basin (Rentema), conglomerate beds dated at 54 Ma are conformably overlain by early to middle Eocene lacustrine deposits, equivalent to the Pozo Formation (Naeser *et al.*, 1991; Jaillard, 1994). Comparable lacustrine deposits of Eocene age are known in the Altiplano of southern Peru (Chilca Formation, Carlotto, 1998) and locally in Bolivia (Cayara Formation, Sempere *et al.*, 1997). Probably due to subsequent erosions, the overlying late Eocene succession is frequently lacking on the borders of the basin (Ecuador - Subandean Zone; Peru - Rentema, Ucayali Basin).

Middle-late Eocene event (40 - 35 Ma) and Oligocene evolution (35 - 28 Ma)

The late Eocene event has long been recognized in the Andes of Peru (Incaic phase, Steinmann, 1929), and has been further documented on the basis of radiometric data. It is followed by the deposition of unconformable beds of latest Eocene-middle Oligocene age. Sedimentation is chiefly tectonically driven in the eastern intermontane and foreland continental basins. This period ended with the late Oligocene Aymara tectonic event (28 - 26 Ma, Sébrier *et al.*, 1988; Sempere *et al.*, 1990; Fig. 32).

In the fore-arc basins, the late Eocene tectonic event was announced by the deposition of late middle Eocene disconformable coarse-grained deposits (42 - 40 Ma). It culminated in the late Eocene (37 - 35 Ma) with the deformation and emergence of many external fore-arc basins (Macharé *et al.*, 1986; Séranne, 1987; Ballesteros *et al.*, 1988; Jaillard *et al.*, 1995; figs. 25 and 29). This event, together with the late Oligocene crisis, is responsible for a widespread Oligocene hiatus in the fore-arc zone. However, sedimentation occured in a few basins (Talara, locally), and some internal (eastern) fore-arc basins were affected by significant subsidence, which allowed the deposition of late Eocene to early Oligocene marine (Pisco) or continental sequences (Moquegua) (Macharé *et al.*, 1988; De Vries, 1998). This subsidence pulse announced the accelerated subsidence, related to tectonic erosion processes, which affected the Andean fore-arc zones from the Eocene (Suess *et al.*, 1988; Bourgois *et al.*, 1990; Von Huene and Scholl, 1991).

In Ecuador, undated coarse-grained conglomerate beds of fan-delta environment, which unconformably overlie the Eocene sequence may be ascribed either to the latest Eocene-early Oligocene (Jaillard *et al.*, 1995; Fig. 25), or to the late Oligocene (Benítez, 1995). Marine shale, siltstone and fine-grained sandstone are dated as middle Oligocene (Playa Rica Formation, Benítez, 1995). They rest disconformably on the Eocene sequence and are separated from the Miocene deposits by a sedimentary hiatus (Benítez, 1995). In northern Peru (Talara Basin), the late middle Eocene sandstone beds are overlain by pelagic shale of debated, possibly early Oligocene, age (Chira Formation, Morales, 1993). In southern Central Peru (Pisco Basin), 600 m of transgressive shale, siltstone and subordinate sandstone of intertidal to nearshore environments are regarded as of late Eocene, maybe early Oligocene (?), age (Paracas Formation, Newell, 1956; Marocco and De Muizon, 1988; Macharé *et al.*, 1988). A middle Oligocene marine sequence has been recently described (De Vries, 1998), which probably correlates with the Oligocene beds of Ecuador (and northern Peru?). In Southern Peru (Moquegua), transgressive fan conglomerate, fluvial sequences and evaporite-bearing lacustrine silt and shale are ascribed to the Eocene, and infill an extensional, fault-controlled basin, probably created after the late Eocene event (Marocco *et al.*, 1985). As for many fore-arc basins of Peru, these beds unconformably overlie Precambrian to Mesozoic rocks deformed by the Late Cretaceous to late Eocene tectonic phases.

In the arc zone of central Ecuador (Cuenca), volcanic rocks of early middle Eocene age (43 Ma, Steinmann, 1997) are overlain by a 1000 m thick series of fluvial conglomerate and sandstone beds of late middle to late Eocene age (Quingeo Formation, 42 - 35 Ma, Steinmann, 1997; Fig. 33). In central and northern Peru, intrusions in the Coastal Batholith ceased by latest Eocene times (35 Ma, Beckinsale *et al.*, 1985; Mukasa and Tilton, 1985; Soler, 1991). With respect to the Cretaceous-Paleocene intrusions, the late Eocene-recent arc magmatism exhibits significant geochemical changes, regarded as resulting from the late Eocene tectonic event (Soler, 1991; Fig. 20). In the present-day Andes of northern and central Peru, the late Eocene event is materialised by a widespread unconformity, the age of which is bracketed between 44 and 40 Ma (Noble *et al.*, 1974, 1979, 1990; Mégard *et al.*, 1996). In southern Peru, the middle to late Eocene batholiths of the inner arc are intruded by acid, calc-alkaline subvolcanic stocks of earliest Oligocene age (34 - 32 Ma), thus indicating a strong uplift event during the late Eocene (Cárlier *et al.*, 1996; Carlotto, 1998).

In the arc zone of northern Chile, a significant angular unconformity is dated at 39 - 38 Ma (Hammerschmidt *et al.*, 1992). Late Eocene upright anticlines, which account for 25% shortening in the arc zone, were associated with arc-parallel dextral strike-slip movements and with E-vergent folds and reverse faults in the back-arc zone (Scheuber *et al.*, 1994). In the proximal back-arc area, the middle-late Eocene tectonic event is recorded by the post 42 Ma unconformity which separates the Purilactis and Paciencia groups (Flint *et al.*, 1993; Fig. 30).

In the arc zone of central-southern Ecuador, the late Eocene event is followed by an important pulse of arc volcanism (andesite, dacite and subordinate rhyolite) dated as latest Eocene-Oligocene (39 - 23 Ma, Saraguro Group; Egüez *et al.*, 1992; Dunkley and Gaibor, 1998). Within this pile, Dunkley and Gaibor (1998) identified erosional periods of latest Eocene-earliest Oligocene (36 - 34 Ma) and middle Oligocene age (30 - 27 Ma). In northern Ecuador, volcanic activity seems to have decreased in the Oligocene, but chronological data are scarce (Egüez, 1986; Wallrabe-Adams, 1990).

In the arc zone and the paleo-Andes of central Peru, a plutonic pulse of late middle and late Eocene age (42 - 36 Ma) is followed by a minor pulse of middle Oligocene age (31 - 30 Ma), the latter being restricted to the Paleo-Andes (Soler, 1991). Volcanic activity displays a correlative evolution, since the volcanic Calipuy Formation yielded ages of 41 - 35 Ma, and 31 - 29 Ma (McKee and Noble, 1982; Noble *et al.*, 1979; Soler, 1991). The middle to early late Oligocene magmatic quiescence is correlated with a low convergence period (31 - 26 Ma, Sébrier and Soler, 1991), and is marked by a subtle change in the geochemical composition of the arc magmatism (Soler, 1991). In the Altiplano and Eastern Cordillera of southern Peru, a significant episode of high-K alkaline magmatism occurred between 30 and 27 Ma (Bonhomme *et al.*, 1985; Bonhomme and Carlier, 1990), which express a local extensional regime (Carlier *et al.*, 1996) and is coeval with the emplacement of monzogabbro at the southern edge of the Altiplano (30 Ma, Clark *et al.*, 1990). These late Oligocene-earliest Miocene alkaline, shoshonitic and high-K calc-alkaline effusions and intrusions are interpreted as the result of partial melting of an enriched mantle wedge (Sébrier and Soler, 1991).

In the arc zone of northern Chile, Oligocene times are marked by the deposition of mainly sedimentary, fluvial beds, which indicate a period of magmatic quiescence (40 - 28 Ma, Azapa Formation and Paciencia Group, Coira *et al.*, 1982; Flint *et al.*, 1993; Garcia, 1997; Fig. 30). In the Paleo-Andes, the middle-late Eocene event is well-expressed. In the Western Cordillera of Ecuador, late Eocene times are marked by the deposition of subaerial conglomerate on the Eocene marine sequence, interpreted by some authors as the result of the accretion of the Western Cordillera terrane (Bourgois *et al.*, 1990; Litherland *et al.*, 1994; McCourt *et al.*, 1998).

Deformation is maximum in the Western Cordillera where E-verging fold and thrust belts developed (Mégard, 1978; Ángeles, 1987; Mourier, 1988). In southern Peru, late Eocene times (42 - 38 Ma) are also marked by thrusting to the NE along the southern border of the Altiplano (Laubacher, 1978; Farrar *et al.*, 1988; Carlotto, 1998), and also by SW-verging thrust faults NE of the Altiplano (Huancané Fault Zone, Laubacher, 1978). Farther to the NE, the middle-late Eocene event is responsible for widespread unconformities in the arc zone and the Altiplano, and for disconformities in the eastern areas (Sébrier *et al.*, 1988; Farrar *et al.*, 1988; Noble *et al.*, 1999).

In the present-day Western Cordillera of northern and central Peru, Late Cretaceous strata are folded and faulted, and in the eastern part, compressional deformations result in a 50 km-large, NE-verging fold and thrust belt (Marañon FTB, Mégard 1984, 1987), which occurs on the western border of the Mesozoic positive zone (Marañon Geanticline), and interpreted as the result of the tectonic inversion of normal paleo-faults (Mourier, 1988). This belt expresses a significant shortening of the continental crust and its overlying cover. They are associated with coarse-grained deposits exhibiting internal unconformities (Pacobamba Formation, Ángeles, 1999). In southern Peru, the Incaic Deformation resulted in a comparable NE-verging fold and thrust belt to the S of the Cusco-Puno Swell (Mañazo FTB, Jaillard and Santander, 1992; Carlotto, 1998), and in the SW-verging Huancané Fault Zone (Audebaud *et al.*, 1976; Laubacher, 1978). The subsequent erosion period is concealed by the deposition of widespread, unconformable coarse-grained conglomerate (Chanove *et al.*, 1969), and documented locally by Oligocene terrestrial faunas preserved in karst excavations (Hartenberger *et al.*, 1984). Farther to the NE, the middle-late Eocene event is responsible for widespread unconformities on the Altiplano, and by disconformities in the eastern areas (Laubacher, 1978; Sébrier *et al.*, 1988; Farrar *et al.*, 1988).

In the Cusco area (southern Peru), 5000 to 6000 m of alluvial red beds (San Jerónimo Group), formerly ascribed to the Late Cretaceous (Gregory, 1916; Jaillard *et al.*, 1993; Noblet *et al.*, 1995), are presently dated as late Eocene(?)-middle Oligocene age (Carlotto, 1998; Fig. 34). They comprise two thick coarsening-upward sequences affected by large-scale progressive unconformities showing evidence for syn-sedimentary compressional or transpressional deformation (Córdova, 1986; Noblet *et al.*, 1987; Carlotto, 1998). Farther to the E, as much as 2000 m of very coarse-grained fanglomerate and sandstone of late Eocene to middle Oligocene age (Anta Formation) unconformably overly Cretaceous to middle Eocene rocks, thus providing evidence for strong magmatic and tectonic activity (Carlotto, 1998). Coeval deposits are known farther to the SE from isolated basins exhibiting changing sedimentary and paleogeographic evolutions, and yielding scarce early Oligocene ages (30 - 27 Ma, Carlotto, 1998).

In the Altiplano Basin of Bolivia, the Paleozoic basement is unconformably overlain by a 3000 m thick series of red shale and sandstone beds, with evaporite units in the lower part, dated at 30 - 29 Ma (Tiwanacu Formation, Rochat *et al.*, 1998; Fig. 35). This succession exhibits eastward paleocurrents and is interpreted as the foreland sequence of the Western Cordillera deformed during the late Eocene event (Sempere *et al.*, 1990; Sempere, 1995, Rochat *et al.*, 1998). However, Lamb *et al.* (1997) determined westward paleocurrents and proposed that the Eastern Cordillera was also uplifted along an E-verging thrust fault during the middle-late Eocene deformation, and thus, separated the Altiplano Basin from the incipient eastern foreland basin (Lamb and Hoke, 1997).

In the back-arc basins of Ecuador late Eocene-Oligocene times are represented by disconformable quartz-rich conglomerate (Upper Tiyuyacu Formation), overlain by fine-grained red beds (Orteguaza Formation) exhibiting a conspicuous transgressive layer of partly marine glauconitic sandstone (Benítez *et al.*, 1993; Rivadeneira and Baby, 1999). In eastern Peru, the late Eocene event accounts for a widespread sedimentary hiatus encompassing the late Eocene-middle Oligocene time-span in the western and southern zones (Koch and Blissenbach, 1962; Naeser *et al.*, 1991; Figs. 31 and 36) and for a slight unconformity farther to the E and NE. There, Robertson Research (1990) identified a thin lacustrine unit of probable Oligocene age. Although available stratigraphic data are scarce and sometimes conflicting, they suggest a noticeable decrease of the tectonic subsidence during the late middle to late Eocene interval (40 - 35 Ma, Thomas *et al.*, 1995; Berrones and Cotrina, 1996; Contreras *et al.*, 1996).

The frequent lack of late Eocene-middle Oligocene deposits in the Oriente Basin contrasts with the thick accumulations of coeval deposits in the Paleo-Andes, which seem to have been marked, however, by an E to NE drainage system. This suggests that, either these deposits have been eroded due to a significant late Oligocene uplift of the Eastern Basin, or the entire detrital sediments have been trapped within the Andean basins, which acted therefore as the proximal foreland basins of the Western Cordillera FTB (Sempere, 1995; Carlotto, 1998), the Eastern Basin constituting a by-pass zone for low discharge rivers.

Orogenic Evolution of the North-Central Andes (Late Oligocene - Present)

Late Oligocene - middle Miocene evolution (28 - 10 Ma)

The late Oligocene "Aymara" tectonic event (28-26 Ma)

A major tectonic and geodynamic event occurred in the late Oligocene (28 - 26 Ma). It has been described by Sébrier *et al.* (1988) and Sempere *et al.* (1990; Fig. 32). The late Oligocene event is related to a major plate dynamics reorganization that occurred at 26 Ma. This consisted of the break up of the Farallón Plate into the Cocos and Nazca plates, accompanied by a change in the direction of convergence (Pilger, 1984; Pardo-Casas and Molnar, 1987). Convergence became approximately E-W, which triggered a progressive rotation of the strain, from NNE-SSW during the late Oligocene, to E-W at the end of the Miocene (Noblet *et al.*, 1988).

This tectonic event was marked by regional unconformities in the Andes (Sébrier *et al.*, 1988; Sempere *et al.*, 1990), by the deposition of disconformable coarse-grained conglomerate in the Eastern Basin, by the inception of eastward thrusting in the Sub-Andean Zone (Sempere *et al.*, 1990), and by a sharp increase of the subsidence rates in the Eastern Basin (Thomas *et al.*, 1995; Berrones and Cotrina, 1996). It also triggered or accelerated the subsidence related to subduction-related tectonic erosion in the fore-arc zones, since in most areas, pelagic Miocene deposits disconformably overlie Eocene shelf deposits (Macharé *et al.*, 1986; Suess *et al.*, 1988). This event is also marked by pre-23 Ma disconformities in the Andes of northern Peru (Mourier, 1988), by some resets of K/Ar ages in the Eastern Cordillera of Ecuador (35 - 25 Ma, Litherland *et al.*, 1994), by unconformities at the base of the late Oligocene volcanics of Ecuador (base of the Saraguro Formation, 29 - 26 Ma, Dunkley and Gaibor, 1998; Steinmann *et al.*, 1999), and by unconformities and syn-tectonic sediments in Bolivia (Sempere *et al.*, 1990; Rochat *et al.*, 1998).

Latest Oligocene-early Miocene evolution (26 - 17 Ma)

The major plate dynamics reorganization of late Oligocene age provoked a renewal of tectonic activity, which accelerated the shortening and uplift of the Andes, and induced thick continental sedimentation in the intermontane and retro-arc foreland basins. The late Oligocene tectonic event is post-dated by the creation of a nearly continuous belt of fore-arc (Macharé *et al.*, 1986) and by a sharp increase of the subsidence rates in the Eastern Basin (Thomas *et al.*, 1995). Offshore northern Peru, tectonic subsidence is marked by the unconformable rest of Middle Miocene pelagic deposits upon Eocene shelf deposits (Macharé *et al.*, 1986; Suess *et al.*, 1988; Von Huene *et al.*, 1988; Bourgois *et al.*, 1990; Fig. 37).

In coastal Ecuador, the early Miocene sequence begins locally with transgressive conglomerate overlain by marine shale and siltstone rich in planktic foraminifera and radiolaria (Dos Bocas and Villingota formations, Evans and Whittaker, 1982; Benítez, 1995). The upper part, of early middle Miocene age, locally grades laterally into coarser-grained subaerial deposits (Benítez, 1995). Rapid subsidence of the northern Talara and Tumbes basins is expressed by the deposition of a 250 to 1000 m thick transgressive series of locally conglomeratic sandstone beds, with marly and carbonate-rich intercalations of paralic environment (Máncora Formation), which unconformably rest on Paleozoic rocks (León, 1983). Further subsidence allowed the deposition of as much as 1000 m of shale, marlstone and sporadic turbidite units rich in planktonic foraminifera, which indicate a significant deepening of the basin during the early to middle Miocene (Heath Formation, León, 1983). Offshore northern Peru, early Miocene marine deposits are mentioned only locally. They overlie directly middle Eocene strata (Ballesteros *et al.*, 1988).

In southern central Peru (Pisco Basin), the latest Oligocene-early Miocene deposits consist of a 60 to 300 m thick series of transgressive shale, siltstone and fine-grained sandstone, which unconformably rests on Paleozoic to early

Oligocene rocks (Caballas Formation, Marocco and De Muizon, 1988; Macharé *et al.*, 1988). Farther to the S (outer fore-arc), the late Oligocene series is represented by sandstone, conglomerate and shale of nearshore environment (Camaná Formation, Rüegg, 1956; Macharé *et al.*, 1988). In the inner fore-arc zone of southern Peru (Moquegua Basin; Fig. 32), as much as 700 m of coarse-grained fluviatile deposits dated by volcanic intercalations (25 - 23 Ma, Upper Moquegua Formation, Noble *et al.*, 1985; Marocco *et al.*, 1985) recorded local and short-lived marine transgressions. On its eastern border, superimposed alluvial fans indicate a coeval tectonic uplift of the paleo-Andes. The coarse-grained fore-arc deposits indicate that the paleo-Andes of southern Peru were more actively uplifted than those of Central Peru during late Oligocene-early Miocene times. A similar situation is recorded in the Azapa Basin of northern Chile. The Chilean margin shows in its northern part a well-expressed extensional, asymmetric basin (Muñoz and Fuenzalida, 1997), similar to the Neogene basins occurring farther N (Von Huene and Scholl, 1991), and a horst and graben topography (Fig. 2C).

In the magmatic arc of southern Ecuador, a significant pulse of mainly acid to intermediate arc magmatism of late early Oligocene-early Miocene age has been recognized (33 - 16 Ma, Saraguro Formation, Aspden *et al.*, 1992; Lavenu *et al.*, 1992; Hungerbühler, 1997; Dunkley and Gaibor, 1998). Detailed analysis discloses ignimbritic events at 28 - 26 Ma, 24 - 22 Ma and 20 - 18 Ma, suggesting an extensional tectonic setting in central and southern Ecuador (Steinmann, 1997; Fig. 33).

In northern Peru, only continental greywacke and conglomerate yielded a 23 Ma K/Ar age (Noble *et al.*, 1990). In the magmatic arc zone and Paleo-Andes of Central Peru, after the 30 - 26 Ma plutonic gap, effusion of calc-alkaline basaltic andesite, andesite and dacite resumed near the Oligocene-Miocene boundary (26 - 19 Ma), and occupied a 150 km wide area (Sébrier and Soler, 1991; Soler, 1991; Fig. 20) They originated in the mantle wedge modified by fluids or melts issued from the subducting slab (Sébrier and Soler, 1991).

In the arc zone of southern Peru, after a long volcanic gap (55 - 27 Ma), huge volumes of basaltic to dacitic flows and tuff were outpoured (Tacaza and Sillapaca formations, Barroso Group, 27 - 15 Ma, Lefèvre, 1979; Tosdal *et al.*, 1984; Klinck *et al.*, 1986; Sébrier *et al.*, 1988; Clark *et al.*, 1990; Fig. 32). Their chemical signature indicates a significant crustal contamination, indicative of the beginning of crustal shortening and thickening (Boily *et al.*, 1990). Granodiorite plutons (25 Ma, Bonhomme *et al.*, 1985) and abundant calc-alkaline volcanism occurred along the western Cordillera-Altiplano boundary (Carlier *et al.*, 1996). This period is also marked by a significant peraluminous magmatism in the Eastern Cordillera of southern Peru (28 - 23 Ma, Kontak *et al.*, 1986; Laubacher *et al.*, 1988; Clark *et al.*, 1990; Carlier *et al.*, 1996), which would result from melting of crustal continental material (Sébrier and Soler, 1991). It is associated with and/or followed by emplacement of shoshonitic bodies and K-rich to high-K minettes exhibiting lamproitic affinities (25 - 20 Ma, Bonhomme *et al.*, 1985; Kontak *et al.*, 1986; Carlier *et al.*, 1996). The latter would reflect the thickening of the lithosphere due to incipient

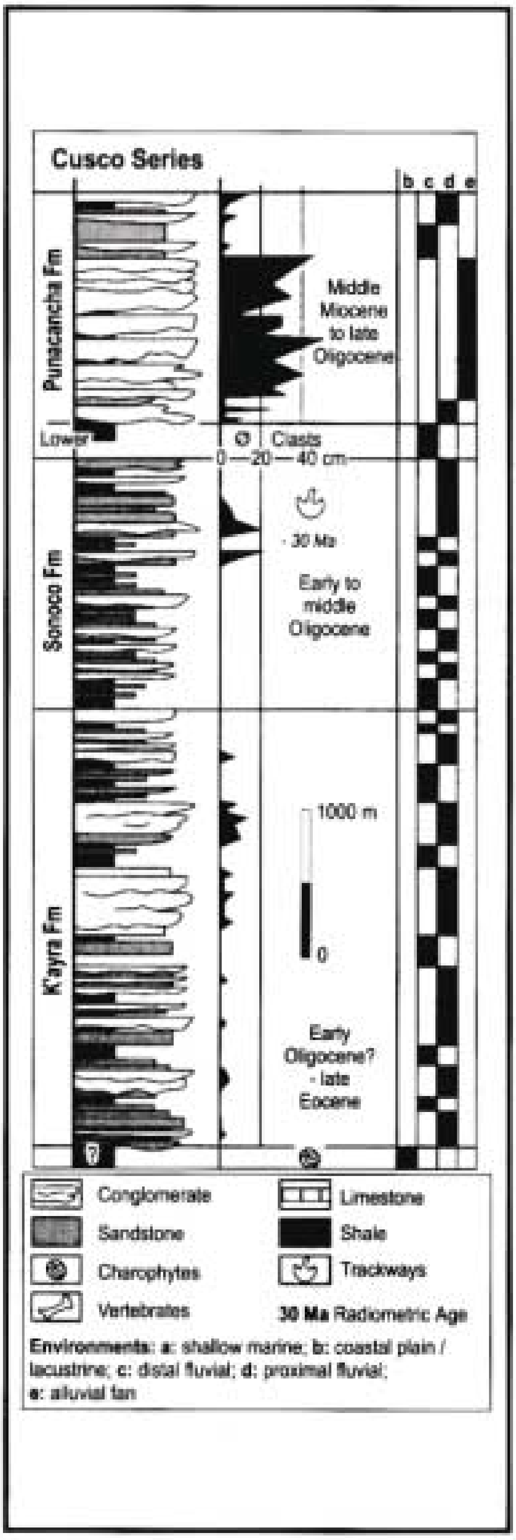


FIGURE 34 - Stratigraphic succession of the Eocene-Miocene "Red Beds" of the Cusco area (Andes of southern Peru) (after Carlotto, 1998).

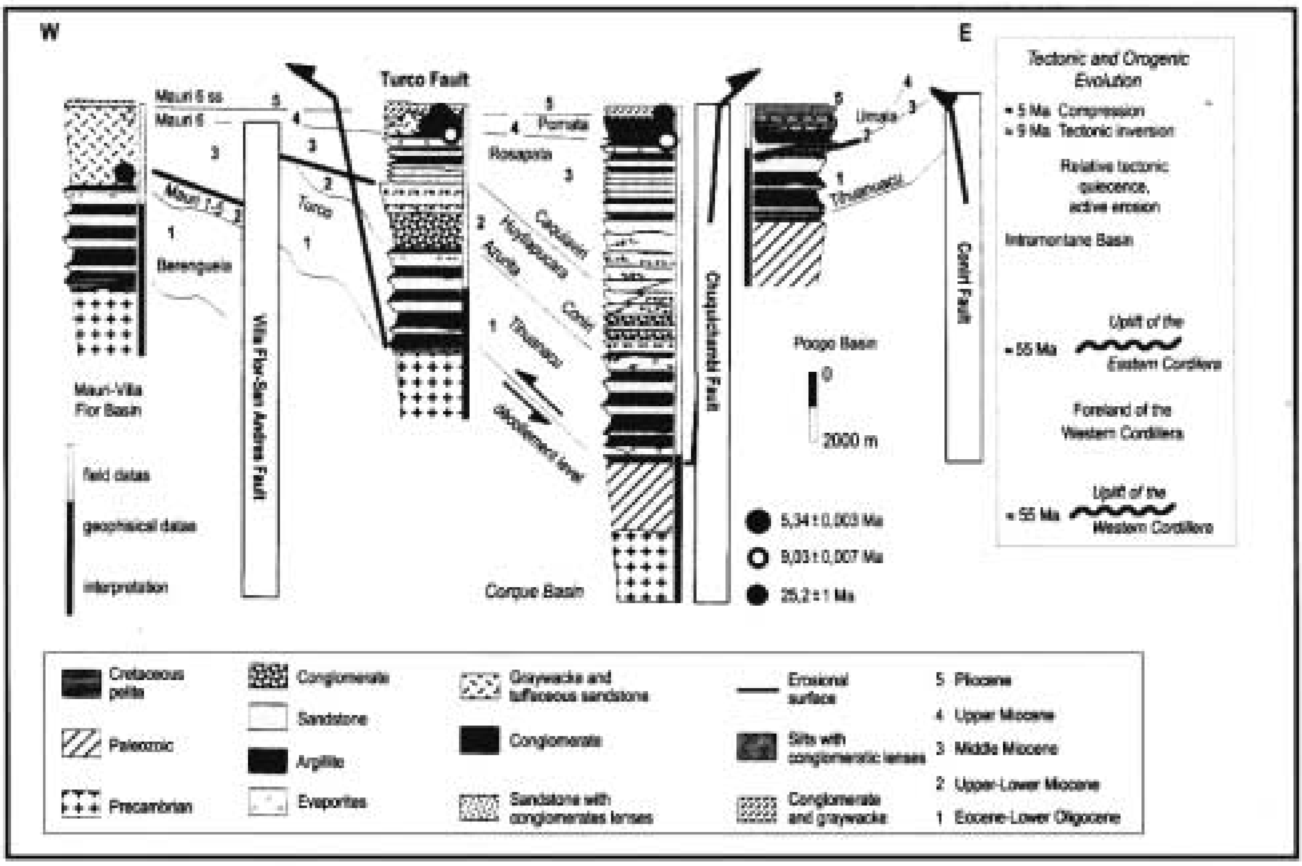


FIGURE 35 - Representative stratigraphic successions accross the Altiplano Basin of Bolivia (after Lamb et al.*, 1997, Rochat* et al.*, 1998).*

thrusting to the NE of the Altiplano onto the Brazilian Shield, triggering also the partial melting of the continental crust (Carlier *et al.*, 1996).

In northernmost Chile, volcanic activity resumed by 25 Ma Oligocene terrestrial sediments (Azapa Formation, Muñoz and Charrier, 1996; Garcia, 1997; Garcia *et al.*, 1999; Fig. 38) are unconformably overlain by late Oligocene-middle Miocene acid tuff associated with fluvial sediments deposited in an extensional environment (Lupica and Oxaya formations, 25 - 18 Ma). Farther to the S (Atacama area), no deposits are known between the playa and fan sediments of the upper Paciencia Group (28 Ma) and the unconformable volcanic rocks of the San Bartolo Group, the base of which is dated at 17 Ma (Flint *et al.*, 1993).

In the present-day Andes of Peru and Ecuador, the late Oligocene-early Miocene was thought to be marked by the creation of a remarkable belt of intermontane basins (Marocco *et al.*, 1995; Noble *et al.*, 1999). However, recent F-T dates led to the consideration that most intermontane basins of Ecuador are late early to early middle Miocene in age (18 - 9 Ma, Hungerbühler, 1997; Steinmann, 1997; Steinmann *et al.*, 1999; Fig. 40).

In the Cuzco area, the Oligocene red beds are disconformably overlain by a fine-grained unit, in turn unconformably overlain by a 4000 m thick series of sandstone and conglomerate beds of late Oligocene-early Miocene age (Punacancha Formation, Carlotto, 1998; Fig. 34). These mainly represent reworked volcanic rocks and exhibit compressional syn-sedimentary deformation, thus evidencing noticeable coeval volcanic and tectonic activity (Marocco and Noblet, 1990; Carlotto, 1998).

On the Altiplano and in the Eastern Cordillera, the late Oligocene-early Miocene sequence is marked by the arrival of coarse-grained conglomerate units (Azurita and Coniri formations, Kennan *et al.*, 1995; Rochat *et al.*, 1998; Fig. 35).

The back-arc area is marked by a strong increase in the

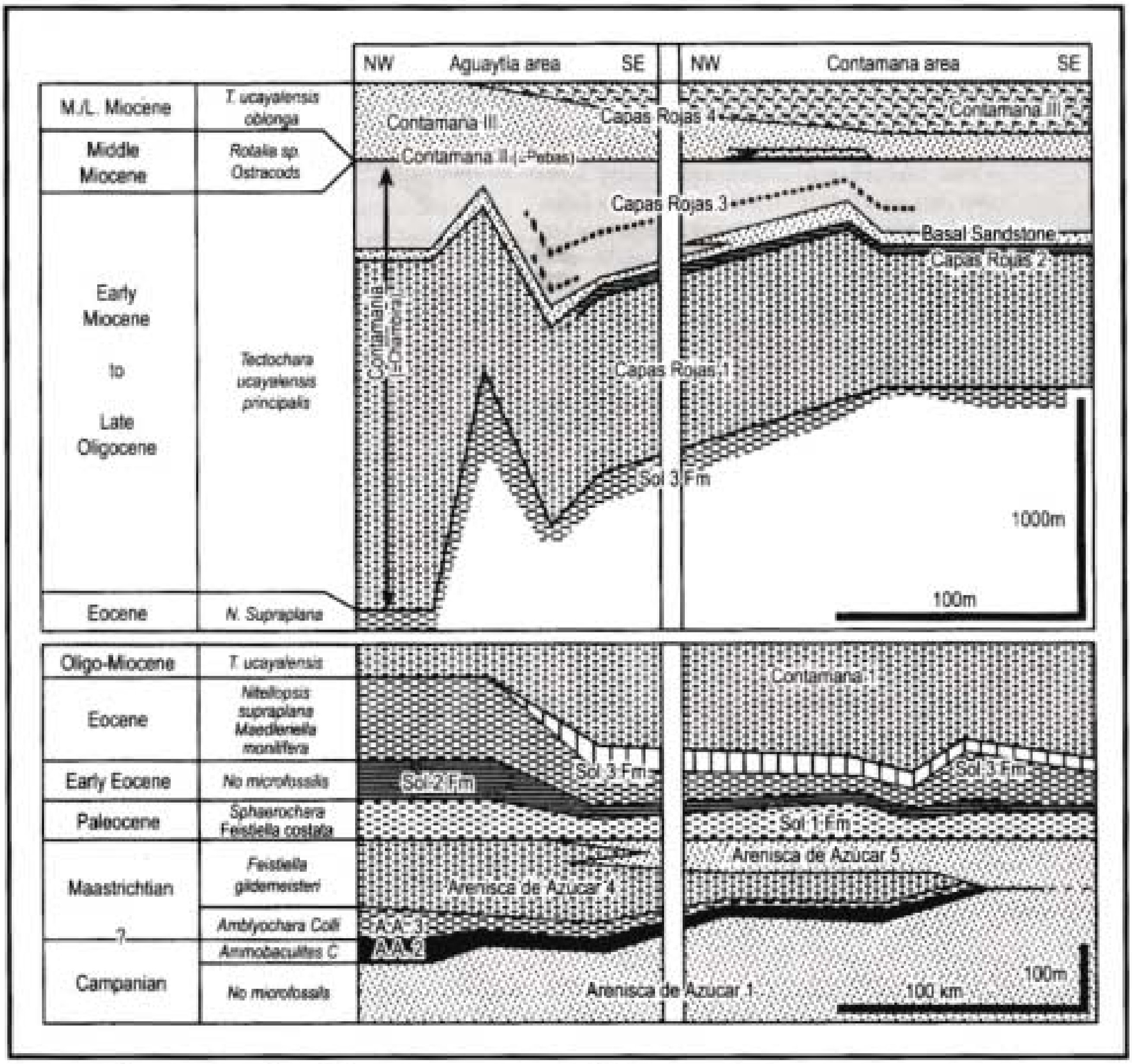


FIGURE 36 - Paleogeographic profiles of the latest Cretaceous-Miocene "Red Beds" of the Ucayali Basin (eastern Central Peru) (after Koch and Blissenbach, 1962; after Müller and Aliaga, 1981).

tectonic subsidence, which marks the start of its evolution as a true retro-arc foreland basin (Sempere *et al.*, 1990; Dashwood and Abbots, 1990; Thomas *et al.*, 1995; Baby *et al.*, 1995; Berrones and Cotrina, 1996; Contreras *et al.*, 1996; Fig. 39). The latest Oligocene-middle Miocene sequences overlie a marked unconformity representing a long period of non-deposition. In Ecuador, Eocene red beds are unconformably overlain by 500-1000 m of siltstone with thin interbeds of sandstone or evaporite, of lacustrine to continental environment, ascribed to the early (to middle?) Miocene (Chalcana Formation). Near Rentema (northern Peru), a 1000 m thick series of alluvial siltstone, sandstone and conglomerate beds (Upper Sambimera Formation) is probably of late Oligocene to middle Miocene age (Naeser *et al.*, 1991; Fig. 31). Farther to the E and SE (Marañon, Ucayali basins), the distal equivalent of this series (Chambira Formation, Contamana I) consists of red siltstone, shale and thin-bedded sandstone, with local evaporite and coal beds of brackish to lacustrine environment. They overlie directly the Eocene beds (Koch and Blissenbach, 1962; Marocco *et al.*, 1993; Fig. 36). These deposits express a resumed erosion of the paleo-Andes subsequent to the late Oligocene phase (Marocco *et al.*, 1993). Farther to the SE (Madre de Dios Basin), the Tertiary succession has never been studied.

In the Cordillera Real of Bolivia, an important magmatic pulse is responsible for the intrusion of numerous S-type granitoid plutons dated between 28 and 23 Ma (Ávila-Salinas, 1990). Farther to the E, the Eastern Basin of Bolivia is marked by a late Oligocene-early Miocene succession of continental sandstone associated with subordinate conglomerate and shale, interpreted as early foreland deposits related to the uplift of the Eastern Cordillera (Petaca Formation, Marshall *et al.*, 1993; Sempere, 1995).

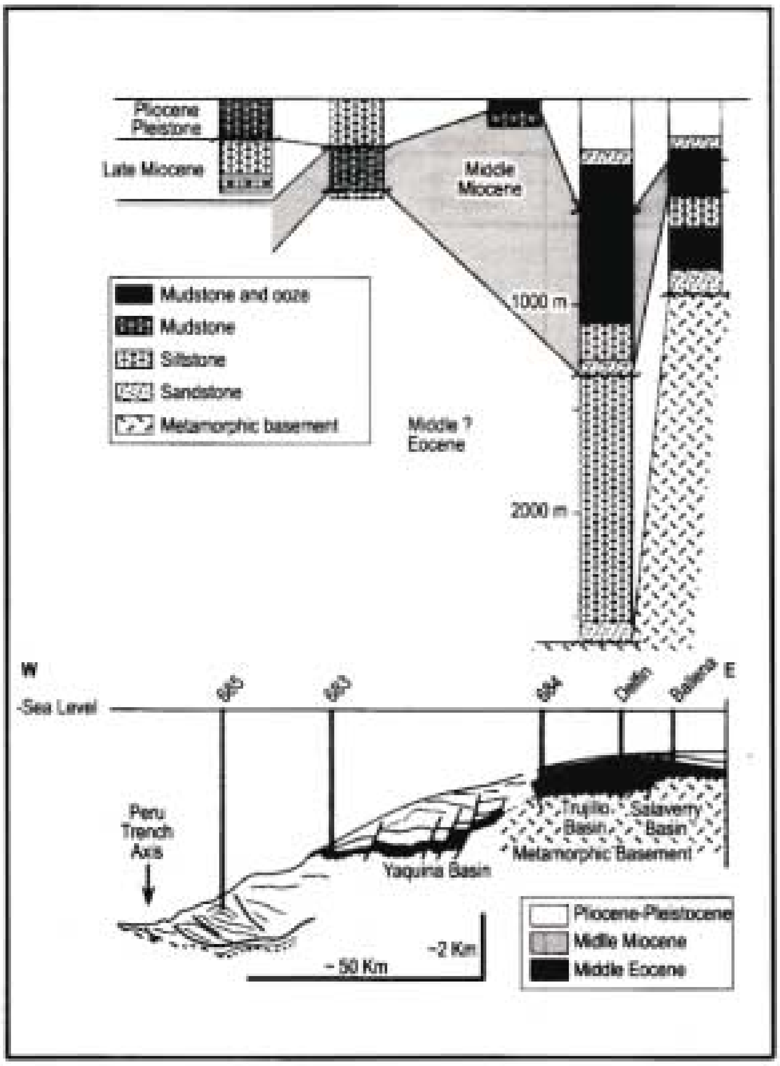
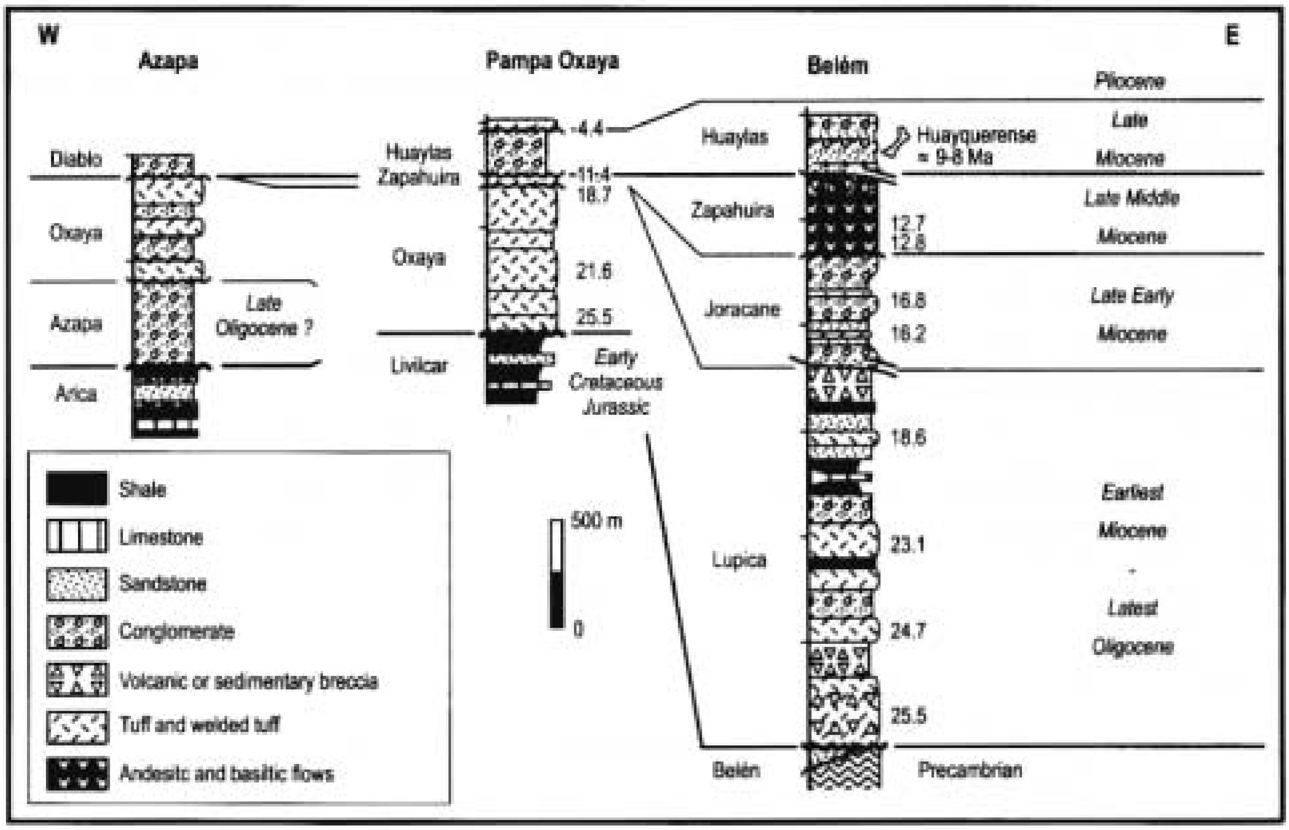


FIGURE 37 - Interpretative structural section and well successions across the fore-arc zone of northern Peru (Trujillo and Naquina basins) (after Suess et al., 1988).

FIGURE 38 - Representative sections in the eastern fore-arc zone of northernmost Chile (after Garcia, 1997; Garcia et al., 1999).



"Quechua 1" event (17 - 15 Ma) and middle to late Miocene evolution (15 - 9 Ma)

In the fore-arc basins of Ecuador, no good record of the event, which occurred near the early-middle Miocene boundary (17 - 15 Ma), is known. This event may be recorded by the arrival of moderate amounts of detrital sediments and by local transgression (Progreso Basin). In the Andes of Ecuador, F-T dates on sediments indicate an uplift stage around 18 Ma (Steimann *et al.*, 1999). In the Eastern Cordillera of Ecuador, resets of K/Ar ages at 20 - 15 Ma may be due to compressional deformations (Litherland *et al.*, 1994). This interval is followed by the opening of the Guayaquil Gulf (Deniaud *et al.*, 1999).

In central Peru, the middle-late Eocene structures were re-activated in the Western Cordillera and in the Marañon Fold and Thrust Belt (Mégard, 1984; Mégard *et al.*, 1984), which fold late Cretaceous red beds. A possible magmatic gap occurred around 19 - 18 Ma in the arc of central Peru (Soler, 1991). In southern Peru, it is responsible for monoclinal folds and reverse faults, large scale incisions due to the resumption of erosions triggered by a pulse of uplift (Sébrier *et al.*, 1988). The latter is thought to be responsible for a 400 m uplift (Sébrier *et al.*, 1988).

In the arc and fore-arc zones of northern Chile (Fig. 42), W-verging fault thrusts are assumed to have began around 18 - 15 Ma (Garcia *et al.*, 1996; Muñoz and Charrier, 1996; Charrier *et al.*, 1999). This event is associated in northernmost Chile, with unconformable late middle Miocene lava flows (Zapahuira Formation, 13 - 11 Ma) on *the early Miocene tuff and conglomerate* (Garcia, 1997; Fig. 38). Farther to the S (Antofagasta), the volcanic rocks of the San Bartolo Group (17 Ma) overlie Oligocene fluvial sediments with an angular unconformity (Flint *et al.*, 1993).

In the fore-arc basins of Ecuador, early Miocene marlstone beds are disconformably overlain by middle Miocene sandstone and marlstone (Subibaja-Angostura Formation). They are in turn overlain by early late Miocene conformable mudstone (Onzole Formation, Benítez, 1995). In the Progreso Basin, the middle Miocene disconformity (Subibaja and San Antonio transgressive limestone members) is followed by deposition of nearshore sandstone of late Miocene age (Progreso Formation), related to the beginning of the first opening stages of the Guayaquil Gulf. After an opening stage (late Oligocene-early Miocene) the Tumbes Basin of northernmost Peru received deep-marine turbidite beds (León, 1983). In the fore arc of central and northern Peru, the early Miocene deposits are conformably overlain by thick middle Miocene marine mudstone (Macharé *et al.*, 1986; Ballesteros *et al.*, 1988; Fig. 37).

Except in Ecuador, the arc zones are marked by the resumption of significant amounts of volcanic products. In central Peru, a magmatic pulse occurred in the Eastern Western Cordillera and the Altiplano between 18 and 13 Ma, which comprises abundant volcanism (Soler, 1991; Fig. 20). In southern Peru, the shoshonitic and High-K magmatism came to an end, while calc-alkaline magmatism went on (Carlier *et al.*, 1996). In the arc and fore-arc zone of northernmost Chile, late middle Miocene lava flows (Zapahuira Formation, 13 - 11 Ma) unconformably overlie early Miocene tuff and conglomerate (Garcia, 1997; Fig. 38). Farther to the S, a volcanic unit (San Bartolo Group, 17 Ma) overlies Oligocene fluvial sediments with an angular unconformity (Flint *et al.*, 1993; Fig. 30).

In the paleo-Andes of Ecuador and Peru, the opening of the "Miocene" intermontane basins had been dated as mainly early Miocene (K/Ar, 26 - 22 Ma, Lavenu *et al.*, 1992) and were regarded as pull-apart basins, opened by the play of an oblique NE trending strain exerted on pre-existing NNE and ENE trending faults (Noblet *et al.*, 1988; Baudino *et al.*, 1994; Marocco *et al.*, 1995; Barragán *et al.*, 1996). However, recent F-T concordant dates support a late early to early middle Miocene age (16 - 14 Ma) for the creation of most intermontane basins of Ecuador (Figs. 33 and 40), which have been alternatively regarded as the result of an E-W extensional regime (Steinmann, 1997; Hungerbühler, 1997; Steinmann *et al.*, 1999). An extensional regime is also thought to have governed the creation of many Miocene intermontane basins of Peru (Noble *et al.*, 1999).

In Ecuador, the rapidly opened intermontane basins (16 - 15 Ma) were filled by middle Miocene fine-grained lacustrine deposits (14 - 10 Ma) representing a period of relative tectonic quiescence (Noblet *et al.*, 1988; Marocco *et al.*, 1995). The fine-grained deposits are overlain by a coarsening-upward sequence of sandstone and conglomerate, coeval with the compressional closure of these basins (Mégard *et al.*, 1984; Noblet *et al.*, 1988; Baudino *et al.*, 1994; Marocco *et al.*, 1995; Hungerbühler *et al.*, 1995; Hungerbühler, 1997; Fig. 40). The Cuenca and Loja basins of southern Ecuador contain middle Miocene marine intercalations, indicating that they were located at or very close to sea level and are regarded as embayments of fore-arc basins (15 - 11 Ma, Hungerbühler 1997; Steinmann *et al.*, 1999; Fig. 40). In Peru, the non marine volcaniclastic and sedimentary in-fill of the middle Miocene basins is usually unconformably overlain by mainly volcanic deposits dated at 10 to 7 Ma (Mégard *et al.*, 1984; Marocco *et al.*, 1995; Noble *et al.*, 1999).

The Altiplano Basin of Bolivia is a peculiar case of intermontane basin consisting of N-S-trending half-grabens (Fig. 35). In this basin, shale, sandstone and subordinate conglomerate of middle Miocene age rest conformably on the early Miocene deposits (Lamb *et al.*, 1997; Rochat *et al.*, 1998). The middle Miocene sequence was deposited with very high sedimentation rates, especially in the Corque Basin of southern Bolivia (Roperch *et al.*, 1999a). Clastic sediments mainly derived from the Eastern Cordillera, the erosion of which allowed the development of regional-scale flat morphological surfaces. These are the Chayanta (13 - 14 Ma) and San Juan de Oro (10 Ma) surfaces, which can be observed and traced from northern Argentina up to the La Paz region (Servant *et al.*, 1989; Hérail *et al.*, 1993; Gubbels *et al.*, 1993). This period coincides with a low uplift rate of the Eastern Cordillera.

In the back-arc basins, a conspicuous shallow marine transgression is recorded during the late middle Miocene (15 Ma, Pebas Formation of Peru, Hoorn, 1993), which was connected to the open marine realm through the Maracaibo area (Hoorn *et al.*, 1995) and possibly the Guayaquil seaway in southern Ecuador (Steinmann *et al.*, 1999). A similar shallow marine invasion of early late Miocene age is known in the Eastern Basin of Bolivia (Yecua Formation, Marshall *et al.*, 1993), which was connected southward to the Atlantic Ocean. In Ecuador and northern Peru, this period is marked by slight decrease of the tectonic subsidence (Thomas *et al.*, 1995; Contreras *et al.*, 1996; Fig.39). However, in eastern Bolivia, a strong increase of tectonic subsidence has been related to the deformation of the eastern Cordillera (Marshall *et al.*, 1993; Sempere, 1995).

"Quechua 2" event (9 - 8 Ma) and late Miocene evolution (9 - 6 Ma)

This period begins with the Quechua 2 tectonic phase (9 - 8 Ma, Mégard, 1984; Sébrier *et al.*, 1988). Rather than a major deformational event, the Quechua 2 event is a turning point in the evolution of the northern Central Andes, which corresponds to the change from a depositional period characterized by thick and relatively widespread fining-upward sequences, to a compressional and uplift period marked by erosions and depositional areas restricted to the fore arc and retro arc domains. This is interpreted as the result of the beginning of the nearly *en bloc* eastward thrusting of the Paleo-Andes onto the Brazilian and Guiana shields, which resulted in crustal thickening and rapid uplift of the arc zones and paleo-Andes, and the transfer of active deformation into the Subandean thrust and fold belts.

The fore-arc zone are mainly marked by uplift (Sébrier *et al.*, 1988), unconformities and reverse faulting. In coastal Ecuador, the middle Miocene disconformity is overlain by transgressive marine sandstone (Angostura Formation) and then by a thinning-upwards succession of marine nearshore sandstone of late Miocene age (Progreso and Lower Onzole formations), related to the beginning of the opening of the Guayaquil Gulf (Deniaud *et al.*, 1999). In the fore-arc zone of Peru, the unconformity between middle and late Miocene marine deposits in the fore-arc basins of central Peru (Lima Basin, 11°S) has not been recognized farther N (Yaquina Basin, 9°S, Ballesteros *et al.*, 1988). In the Lima Basin, low energy shelf deposits of late Miocene age evolved toward shelf to slope deposits (Ballesteros *et al.*, 1988), indicating a noticeable deepening of the depositional environment, due to the subsidence related to tectonic erosion of the fore-arc zones (Von Huene *et al.*, 1988). In the Yaquina Basin, middle Miocene low energy turbidite beds grade upwards into higher energy turbidites of late Miocene-Pleistocene age (Ballesteros *et al.*, 1988).

In the fore-arc zone of northern Chile, where W-verging thrusting went on (Fig. 42), a same kind of unconformity is recognized between late middle Miocene lava and late Miocene alluvial deposits (Huaylas Formation, 9 - 8 Ma, Garcia, 1997; Fig. 38). In the magmatic arc of Peru, no location change is observed in the magmatic activity, with respect to the former periods. A major magmatic pulse occurred between 12 and 7 Ma (peak activity around 10 Ma). It corresponds to numerous intrusions, and very abundant effusive products (Soler, 1991). A possible gap, or at least magmatic quiescence, is mentioned at 9 - 8 Ma, which may coincide with a compressional event (Soler, 1991; Fig. 20).

The Paleo-Andes and surrounding areas are marked by a general and rapid uplift (Sébrier *et al.*, 1988; Steinmann *et al.*, 1999), the local rates of which remain to be specified. In Ecuador, the estimates of mean rock uplift rate since the late Miocene (9 Ma) is of 0.7 mm/y and the mean surface uplift is of 0.3 mm/y (Hungerbühler, 1997; Steinmann *et al.*, 1999), which is consistent with estimates by Delfaud *et al.* (1999) in the same area, by Sébrier *et al.* (1988) in southern Peru, and by Parraguez *et al.* (1997) in northern Chile. In the latter area and in Bolivia, uplift involved both the Western (Sébrier *et al.*, 1988) and Eastern (Benjamin *et al.*, 1987) Cordilleras.

In Ecuador, this period is marked by the compressional closure of the Miocene intramontane basins (Noblet *et al.*, 1988; Marocco *et al.*, 1995; Hungerbühler, 1997; Steinmann *et al.*, 1999), which are filled by coarsening and thickening-upward clastic deposits. This is interpreted as the result of a change from an extensional (15 - 10 Ma) to a compressional regime (9 - 8 Ma), which resulted in the uplift of southern Ecuador, the establishment of terrestrial conditions in the intermontane basins of southern Ecuador and rising relief in the Eastern Cordillera (Hungerbühler, 1997). Latest Miocene times are then marked by the development of smaller-scale intermontane basins filled with mainly volcanic and volcanogenic rocks (Lavenu *et al.*, 1996; Hungerbühler, 1997; Fig. 41).

In northern Peru, although timing constraints are poorer, and the change from extensional to compressional regime is assumed to be of late Miocene age. In central Peru, N-S shortening induced mainly dextral movements along NW trending faults (Mégard, 1984). In the Ayacucho Basin, compressional deformation occurred between 9.5 and 8.5 Ma (Mégard *et al.*, 1984). In southern Peru, latest Miocene transpressional stress is thought to be responsible for the closure of intermontane basins (Paruro Basin) opened about 12 Ma ago (Carlotto, 1998). Farther to the S, late Miocene times are marked by the contraction of the Altiplano, related to the tectonic inversion of the pre-existing normal faults defining the hemi-grabens located W of the Altiplano Basin (Kennan *et al.*, 1995; Lamb *et al.*, 1997; Rochat *et al.*, 1998, 1999; Fig. 35). In the Altiplano Basin, tuff beds dated at 9 Ma are disconformably overlain by a thin sequence of conglomerate reworked Paleozoic basement rocks (Lamb *et al.*, 1997; Rochat *et al.*, 1998).

In northern Chile, compressional tectonic activity went on in the Pre-Cordillera, with the development of the western thrusts of the W-vergent thrust system (Fig. 42), whereas an extensional regime prevailed farther to the W (Longitudinal Valley and Coastal Cordillera; Muñoz and Charrier, 1996; Garcia *et al.*, 1999).

Regarding the Eastern Basin, the eastward migration of the deformation front during the late Miocene is the prevailing feature (Sheffels, 1990; Baby *et al.*, 1997). From this time onwards, most of the deformation and shortening of the Andean margin is accommodated by the eastern areas, which received W-proceeding coarse-grained clastic, terrestrial sediments. In the Oriente Basin of Ecuador, late Miocene deposits consist of thick sequences of poorly dated coarse-grained conglomeratic sequences separated from each other by disconformities (Christophoul, 1999). In the western part of the Eastern Basin of northern Peru (Bagua

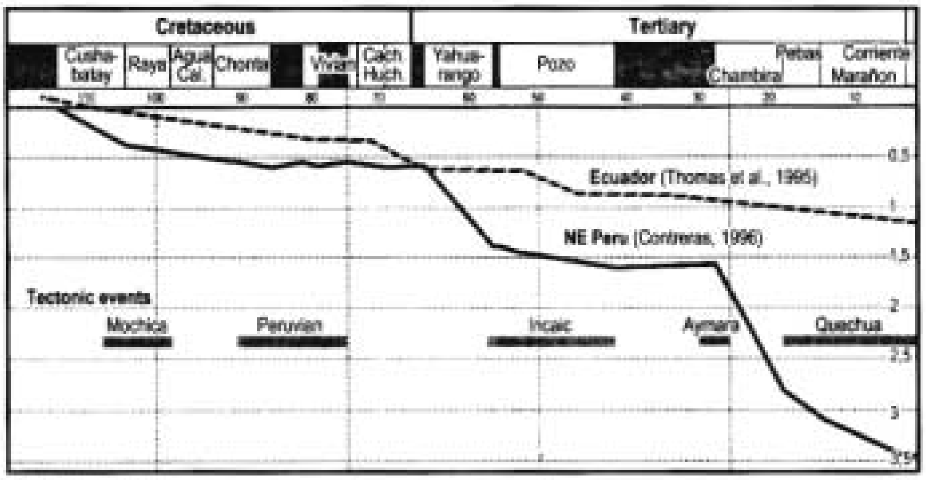
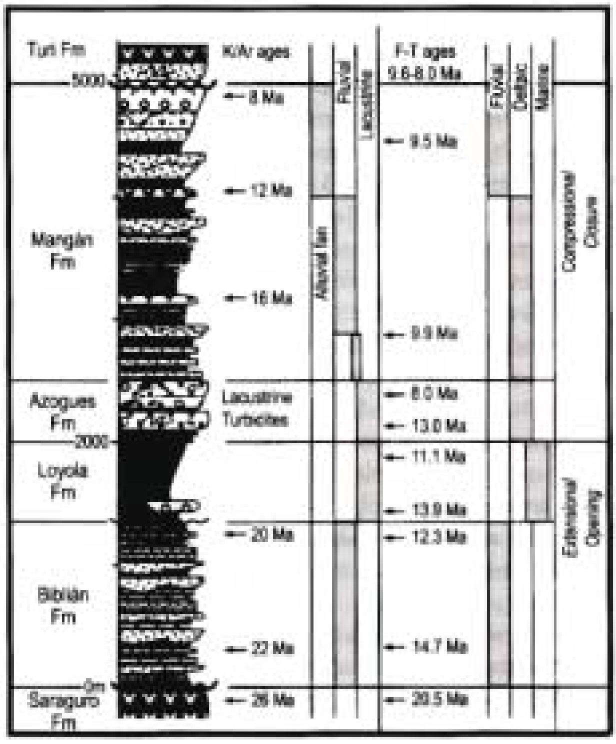
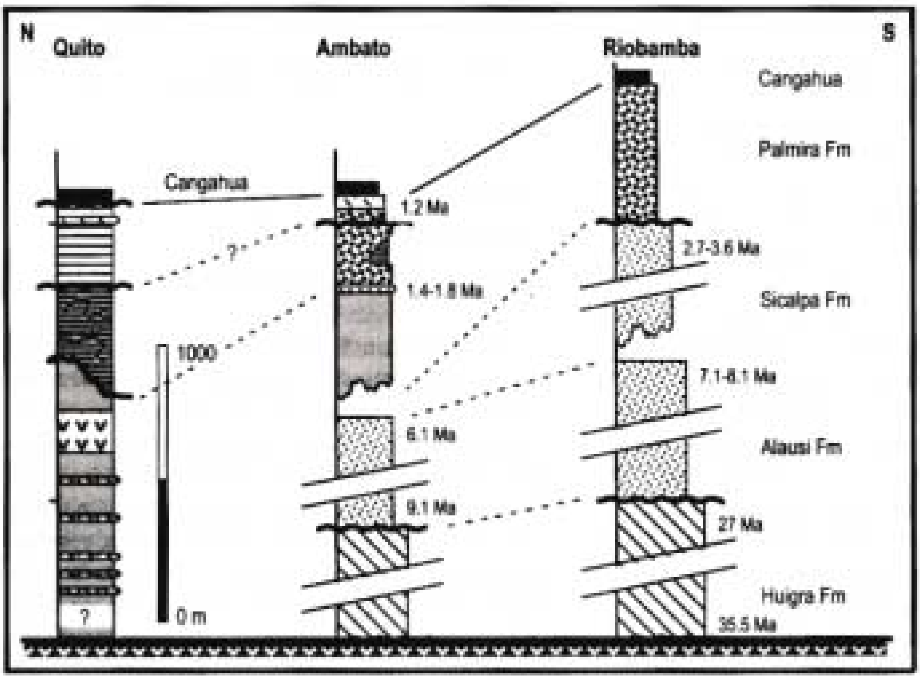


FIGURE 39 - Average tectonic subsidence curves for the Oriente Basin of Ecuador (Thomas et al., 1995) and the Marañon Basin of northern Peru (Contreras et al., 1996).

FIGURE 40 - Lithology, age, environments and evolution of the sedimentary infill of the Cuenca Basin (Ecuador); after Marocco et al., 1995; Steinmann et al., (1999).

FIGURE 41 - Representative successions of the volcanogenic infill of the inter-Andean basins of central Ecuador (after Lavenu et al., 1996).





area), a marked unconformity (10 Ma) separates fluvial sandstone and conglomerate of middle Miocene age, from late Miocene coarse-grained fanglomerate (San Antonio Formation, Mourier *et al.*, 1988; Naeser *et al.*, 1991; Fig. 31). In the Ucayali Basin, fluvial siltstone and sandstone of late Miocene age abruptly overlie marine to brackish beds (Koch and Blissenbach, 1962; Fig. 36).

Quechua 3 event (7 - 5 Ma) and latest Miocene-Present evolution (6 - 0 Ma)

At the beginning of this period, the significant 7 - 5 Ma contractional event is marked by a chiefly E-W shortening (Mégard, 1984; Sébrier *et al.*, 1988). It is marked by folding and reverse and strike-slip faulting in southwestern Peru (Sébrier *et al.*, 1988), and by the onset of the sub-Andean thrust and fold belts, which accommodate most of the shortening in the Andean Chain during the Pliocene (Roeder, 1988; Sheffels, 1990; Baby *et al.*, 1992).

In the fore-arc zone the deformation and uplift were influenced by the incipient subduction of the Nazca and Carnegie aseismic ridges (7 to 3 Ma ago, Suess *et al.*, 1988; Von Huene and Scholl, 1991, Benítez, 1995), and the ongoing tectonic erosion of the fore-arc zones. In the fore-arc zones of Ecuador, a regional disconformity dated at the Miocene-Pliocene boundary (5.5 Ma) precedes the deposition of a coarsening and shallowing-upward sequence, related to the increased uplift and erosion of the Andean Chain from 9 Ma (Benítez, 1995; Deniaud *et al.*, 1999). In the Guayaquil Gulf, however, a strong subsidence due to transtensional movements allowed the deposition of huge volumes of clastic sediments, especially during the early Pleistocene (Deniaud *et al.*, 1999). A similar disconformity and hiatus seem to be recorded at 5 Ma in the offshore basins of Peru (Von Huene *et al.*, 1988).

In central Peru (6° - 14°S), the subsidence of the fore-arc zones related to the tectonic erosion is recorded in the Lima Basin by Pliocene low energy turbidite beds, which indicate deepening of the environment (Ballesteros *et al.*, 1988), by the local transition from uplift to subsidence at 6 Ma (Von Huene *et al.*, 1988), and by the lack of uplifted terraces in the coastal zone (Macharé and Ortlieb, 1993). On the contrary, significant uplift movements affected the coast of northern (4° - 6°S, < 0.2 mm/y) and southern Peru (14° - 18°S, < 0.7 mm/y) since the late Pliocene (Macharé and Ortlieb, 1992; Von Huene *et al.*, 1988). In this latter case, uplift is due to the subduction of the Nazca aseismic ridge, which induced a regional extensional stress regime (Macharé and Ortlieb, 1992)

All segments of the Andes are marked by the continuation of the major and rapid uplift. In Ecuador, uplift rate estimated by F-T evidence a slow down between 6 and 4 Ma, and an increase from 3 Ma (Steinmann *et al.*, 1999). Other methods estimate 1000 - 1200 m of net uplift since 5 Ma (0.2 mm/y, Delfaud *et al.*, 1999). Pliocene-Quaternary times are also marked by the uplift of the sub-Andean Zone of Ecuador (Baby *et al.*, 1999). In the Andes of southern Peru, uplift is estimated at 1300 m since the late Miocene, of which 200 - 300 m would be of Quaternary age.

The arc zones of Ecuador (Steinmann *et al.*, 1999), and Northern and Central Peru are marked by an effusive pulse centred around 5 - 4 Ma. In Peru, it corresponds mainly to ignimbritic tuff associated with rhyolitic dykes in the Western Cordillera (Soler, 1991). In southern Peru, the emplacement of alkaline, peraluminous and shoshonitic suites along major fault systems suggests that an extensional regime prevailed around 6 - 5 Ma (Carlier *et al.*, 1996). Effusions of shoshonite, minette, lamproite and peraluminous rhyolite and dacite went on during the past 3 Ma, and took place along the fault systems limiting the Altiplano, and interpreted as sinistral wrench-faults (Carlier *et al.*, 1996; Carlotto, 1998).

In the eastern basins, latest Miocene-Pliocene times are marked by a strong flexural subsidence allowing thick accumulations of foreland clastic deposits (Thomas *et al.*, 1995; Contreras *et al.*, 1996; Baby *et al.*, 1995; Fig. 39), whereas Recent times are marked by a strong decrease of the sedimentation rate and local uplifts. However, Recent sedimentation continues in restricted and/or more easterly areas (Ucamara depression of easternmost Peru; Pastaza-Marañon alluvial fan).

In the Oriente Basin of Ecuador, localized coarse-grained fanglomerate are incised by present-day rivers (Christophoul, 1999). In northeastern Peru, apatite fission tracks data indicate that the Santiago Basin underwent a rapid uplift (0.4 mm/y) during the last 10 Ma, probably related to the onset of the Santiago Fold and Thrust Belt during the latest Miocene (Pardo, 1982; Mégard, 1984). In the Marañon Basin, a sedimentary hiatus separates middle-late Miocene fine-grained deposits from disconformable coarse-grained fanglomerate of latest Miocene-Recent age (Mathalone and Montoya, 1995) and Pliocene times are marked by the uplift of the area (Contreras *et al.*, 1996). In the Ucayali Basin, no post-Miocene deposits are known (Koch and Blissenbach, 1962). In the Madre de Dios Basin, late Miocene deposits fill incised valleys, and the recent alluvial terrace morphology shows a Pleistocene uplift of the area. In the sub-Andean foreland basin of northern Bolivia, a significant increase in the subsidence allowed the accumulation of about 5000 m of late Miocene-Pliocene clastic sediments. This is interpreted as the result of the rapid eastward migration of the Andean deformation 10 to 6 Ma ago (Gubbels *et al.*, 1993; Baby *et al.*, 1995). However, no foredeep sedimentation occurs at present (Roeder, 1988).

Tectonic and kinematic evolution of the North-Central Andes

From Bolivia to Ecuador, structural style of the Andes changes dramatically. Geometry of the present-day deformation of the Bolivian and North-Chilean Andes results from Neogene thin-skinned tectonics, whereas the Ecuadorian Andes have been structured by thick-skinned and wrench tectonics since Cretaceous times. Figures 2A, 2B and 2C illustrate global changes in structural geometry and chain width. These two parts of the Andes form two extremes. Neogene tectonic events seem to occur contemporaneously, but express two shapes of orogenic belt. Three Neogene orogenic stages, late Oligocene-early

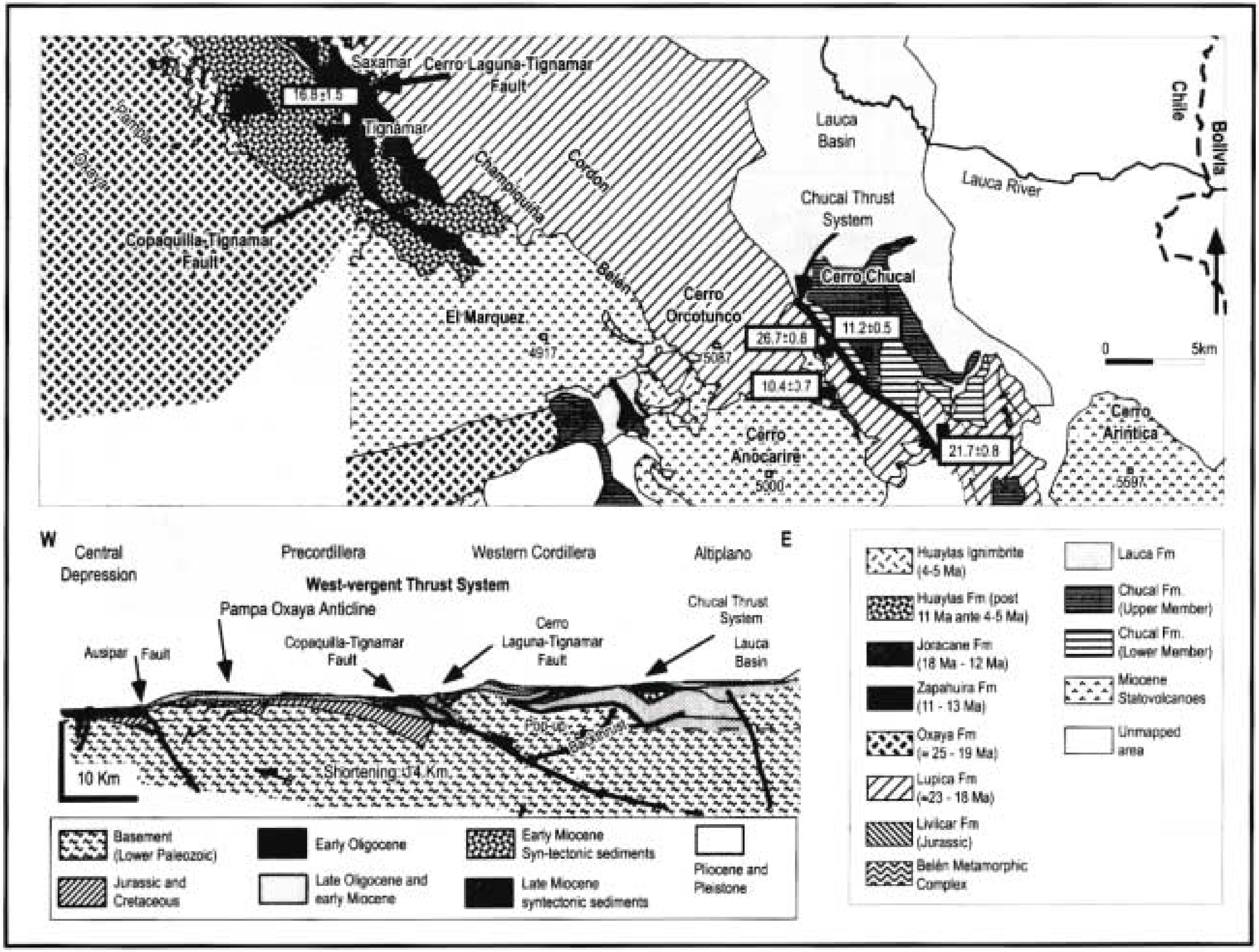


FIGURE 42 - Geology and structural sketch of the Altiplano and Precordillera of northern Chile.

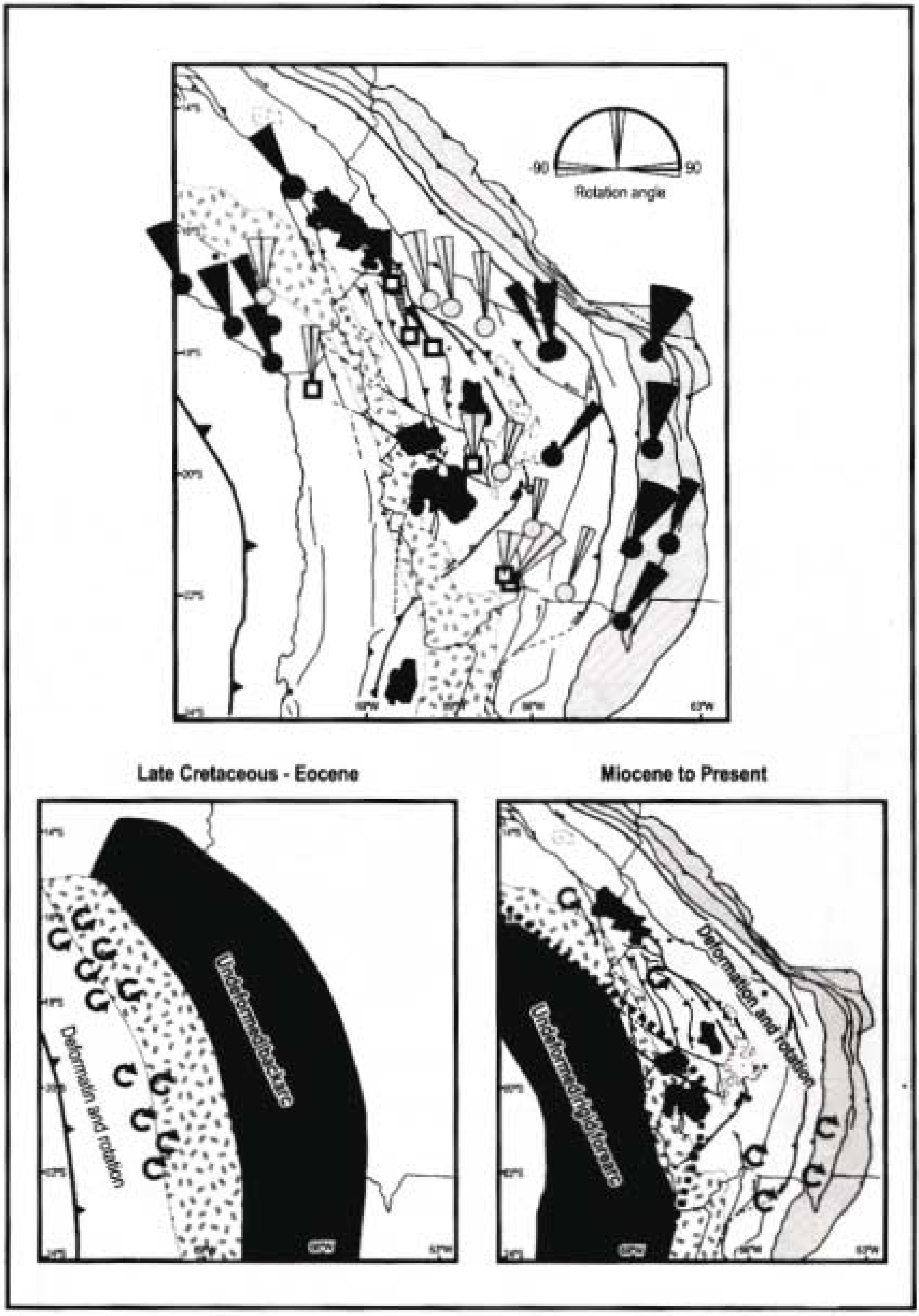


FIGURE 43 - Tectonic rotations in the Central Andes. Important rotations characterize the fore-arc's evolution before the Miocene. The Arica elbow probably developed during the Upper Cretaceous and Eocene. The rotations found in the Altiplano-Puna are linked to the pushing of the rigid fore-arc and to the paleogeographic control of the propagation of the deformation in the subandean belt (after Roperch et al., 1999b).

Miocene, late Miocene, and Pliocene-Quaternary, have been distinguished according to the deformation style and rate. They have been recorded in fore-arc, intermontane and back-arc basins. Sedimentary successions involved in the deformations consist of Cambrian to Oligocene pre-orogenic strata, and Oligocene-Miocene to Recent continental syn-orogenic in-fill.

Central Andes of Bolivia

Recent work has shown the importance of crustal shortening for the development of the structural pattern of this part of the Andes (Allmendinger *et al.*, 1983; Isacks, 1988; Roeder, 1988; Sheffels, 1990; Sempere *et al.*, 1990; Baby *et al.*, 1992, 1997; Gubbels *et al.*, 1993; Schmitz, 1994; Kley and Reinhardt, 1994; Dunn *et al.*, 1995; Rochat *et al.*, 1999; Giese *et al.*, 1999).

Crustal structures and Neogene shortening

The Central Andes are divided from E to W into several morpho-tectonic units (Fig. 43). The Chaco and Beni plains correspond to a slightly deformed Neogene foreland basin underlain by the Brazilian Shield. It is overthrust by the Subandean Zone, a complex thin-skinned fold and thrust belt characterized in its central part (Santa Cruz elbow) by large-scale transfer zones (Baby *et al.*, 1996). The northern branch of the Subandean Zone is characterized by large scale thrust sheets (10 - 20 km of offset) and broad synclines (Roeder, 1988) filled by up to 6000 m of syn-tectonic Neogene sedimentary rocks (Baby *et al.*, 1995a). Surface mapping, seismic reflection data, and drilling information show that the main detachment planes are located in the Ordovician, Silurian, Devonian and Permian shaly levels (Baby *et al.*, 1995b). The slope of the base of the foredeep is 4° toward the SW. The amount of shortening is 74 km, i.e. 50%.

In the southern branch of the Subandean Zone, a regional E-verging thrust (Mandiyuti Thrust) divides the southern Bolivian Zone into two fold and thrust belts, which differ according to their thrust system geometry. Mainly fault-propagation folds and fault-bend folds characterize the western belt, whereas fault-propagation folds and passive-roof duplexes characterize the eastern belt. Main detachments are located in Silurian dark shale, Lower Devonian shale, and at the base and top of the Middle to Upper Devonian dark shale. The Silurian-Devonian succession is covered by more than 2000 m of upper Paleozoic and Mesozoic sandstone with no potential detachments; in some places it is also covered by several thousand metres of syn-orogenic Neogene sedimentary rocks (Moretti *et al.*, 1996). The base of the foredeep slopes at 2°W. Total shortening decreases southward from 140 km (50%) at 20°S, to 86 km (35%) at 22°S.

The Interandean Zone and Cordillera Oriental are deformed by E-vergent thrusts which involve basement rocks (Kley, 1996), and associated thin-skinned thrusts and back thrusts. Mainly Silurian, Devonian, and Carboniferous strata are exposed in the Interandean Zone. In the Cordillera Oriental, the Neogene thrust system is superimposed on a deeply eroded pre-Cretaceous fold belt that deformed Ordovician anchimetamorphic sedimentary rocks. Shortening is concentrated in the W-verging thrust system of the western part of the Cordillera Oriental and of the Interandean Zone. The Cordillera Oriental is characterized by small Neogene piggyback basins (Fornari *et al.*, 1987; Hérail *et al.*, 1996). Good surface data allowed us to construct some balanced cross sections, according to which total shortening may be estimated between 80 and 100 km. The Altiplano is a complex Neogene intermontane basin deformed by both extensional and compressional tectonics. Surface mapping, seismic reflection data, and drilling information made possible the construction of balanced cross-sections. The total shortening calculated is 20 km and 13 km in the southern and northern parts of the Altiplano, respectively.

The western areas include several morphological units. The Cordillera Occidental includes Plio-Quaternary volcanoes. Its western part is formed by a W-verging thrust system (Muñoz and Charrier, 1996; Fig. 42), characterized by the reactivation of high angle faults and the lack of Paleozoic cover. On the eastern part of the Precordillera, back thrusts limit a blind pop-up structure below the Tertiary deposits (Riquelme and Hérail, 1997). Shortening is close to 18 km. In the Central Valley, within which Late Cretaceous-Paleocene magmatic arc and associated deposits are deformed by a mild Plio-Pleistocene extensional tectonics (Parraguez *et al.*, 1997). The Coastal Cordillera shows low relief constituted by Jurassic - Early Cretaceous magmatic arc rocks. The Chilean margin exhibits a horst and graben topography, and in its central part, a well expressed extensional, asymmetric basin (Muñoz and Fuenzalida, 1997), similar to the Neogene basins located farther N (Von Huene and Scholl, 1991).

Crustal balancing across the Central Andes between 15°S and 18°S (Fig. 2) on the basis of a normal pre-orogenic crustal thickness (according to the location of the Palaeozoic basin and the lack of significant Meso-Cenozoic extension) allows us to calculate 210 km of shortening during the Neogene (Baby *et al.*, 1997). At the latitude of the Arica elbow, shortening is associated with the clockwise rotation of crustal blocks controlled by inherited faults (Fig. 43), due to the compression exerted by the fore-arc zone that behaves as a rigid buttress. These rotations are coeval with the compressional deformation, but the elbow shape of the Bolivian orocline has been acquired prior to this deformation, and is probably of Late Cretaceous or Eocene age (Roperch *et al.*, 1999b).

The Moho shape and the Nazca Plate geometry at this latitude are well constrained by geophysical studies (James, 1971; Cahill and Isacks, 1992; Dorbath *et al.*, 1993; Beck *et al.*, 1996; Zandt *et al.*, 1996). Deep crustal structures are imaged by lower crust reflectors located at different structural levels (Wigger *et al.*, 1994; Allmendinger and Zapata, 1996). The crustal duplexes below the Eastern and Western Cordillera are insufficient to produce the crustal thickening evidenced by geophysical data below the Altiplano and the fore-arc zone. Duplex structures in the lower crust (Lamb and Hoke, 1997) cannot explain the over balanced volume (7216 km^2 in cross-section; Fig. 2), since the lower crust structures have been taken into account in the crustal balancing. Asthenosphere wedge as well as significant volumes of magmatic addition cannot account for the observed thickness (Rochat *et al.*, 1999). The significant tectonic erosion of the Chilean margin (Rutland, 1971; Cloos and Shreve, 1996; Von Huene and Scholl, 1991) and associated extensional deformations suggest that deep crust material removed from the continental edge has been underplated below the fore-arc zone and Altiplano (Schmitz, 1994; Baby *et al.*, 1997).

Timing of Neogene deformations

In the Central Andes, the back-arc thrusting started in the late Oligocene (Sempere *et al.*, 1990; Baby *et al.*, 1997). The first W-vergent thrust motions in the fore-arc zone occurred in the late Oligocene-lower Miocene along the median thrust plane (Garcia *et al.*, 1996). Meanwhile, the Altiplano corresponded to an endorheic basin (Rochat *et al.*, 1998, 1999) situated at the back of the more internal crustal thrust of the Eastern Cordillera. During the upper Miocene, the median thrust plane of the W-vergent thrust system was reactivated (Garcia *et al.*, 1996) and crustal back thrusts produced the partial expulsion of the Altiplano, which represented, therefore, a broad piggy-back basin carried over the crustal duplex of the Eastern Cordillera (Baby *et al.*, 1997). Activity of the Subandean fold and thrust system started at the same period (Gubbels *et al.*, 1993); its eastward propagation accelerated in the Pliocene and continues presently.

Kinematic and dynamic analysis

Tectono-sedimentary studies of the Altiplano (Rochat *et al.*, 1998) indicate a local type isostatic behaviour (deep basin controlled by vertical motion along pre-existing high angle faults). Predicted topography from 10 km of deposits, assuming a normal crust isostatically compensated, is 1.5 km (Rochat *et al.*, 1999). However, no significant absolute subsidence and uplift occurred in the Altiplano during the Neogene. The continuity of the sedimentation in the centre of the Altiplano shows that the topography was archived by filling up of the thick syn-orogenic deposits and progradation over the uplifting borders.

The Neogene filling of the fore-arc extensional basin (Von Huene and Scholl, 1991) is coeval with the sedimentary overloading of the Altiplano, which corresponds to 30% of the volume eroded from the back-arc and the Cordillera Occidental (Rochat, 1999). Timing of both processes indicates that deep tectonic erosion and underplating were able to maintain isostatic equilibrium and consequently the vertical aggradation of the Altiplano level. Along the fore-arc zone, structural traps (like the Altiplano crustal piggyback basin) do not exist. The intensity of Neogene erosion shows that these areas were overcompensated by the deep underplating. The upper Plio-Pleistocene decrease of sedimentation areas in the Altiplano (Rochat *et al.*, 1998) was associated with exorheic drainage. Consecutive minor uplift, as is shown by lacustrine over-deepening and extensional deformations (Lavenu, 1995), show that the equilibrium between superficial sedimentation-erosion and deep erosion underplating was broken. In an ongoing convergence tectonic context, deep up-drive will involve destruction of the Altiplano by erosion and associated collapse.

Andes of Ecuador

The Ecuadorian Andes (1°N - 4°S), are one of the narrowest and most active part of the Andean Belt. It is deformed by NNE-SSW right-lateral transpressive shear zones (Tibaldi and Ferrari, 1992) and underwent an intense Holocene tectonic and volcanic activity. The Dolores-Guayaquil Megashear constitutes an important dextral transcurrent boundary which marks roughly the suture zone between the South American continental margin and the Coastal Block with oceanic basement, accreted during Late Cretaceous-Paleogene times (Juteau *et al.*, 1977; Lebrat *et al.*, 1987; Cosma *et al.*, 1998; Reynaud *et al.*, 1999). Deep geophysical data are not numerous enough to constrain the Moho geometry. Below the chain, the average depth of the Moho is about 50 km (Prévot *et al.*, 1996).

Crustal structures and Neogene deformations

The Ecuadorian Andes are divided from E to W into six morphotectonic units (Fig. 2). The Amazonas foreland basin is deformed by two major NNE-SSW trending, transpressional right-lateral fault zones, which correspond to inverted Mesozoic rift systems (Baby *et al.*, 1999). Positive flower structures were developed along these trends and formed the main oil fields of Ecuador. No Quaternary sedimentary sequences are cropping out in this basin, which seems to undergo uplift presently.

The Subandean Zone is formed by two en echelon NNE-SSW trending positive flower structures (Napo and Cutucú uplifts, Baby *et al.*, 1999), which are still seismically and volcanically active. They result also from transpressional dextral movements, and are separated by a Quaternary pull-apart basin (Pastaza Depression).

The Cordillera Real is a metamorphic belt, intruded by Jurassic batholiths and strongly deformed by wrench tectonics. A W-dipping, high angle reverse fault zone separates it from the Subandean Zone. In the Interandean Valley, thick alluvial, lacustrine and volcaniclastic continental sediments were deposited in several Neogene intermontane basins, controlled by regional strike-slip faults limiting the Interandean Valley (Marocco *et al.*, 1995; Barragán *et al.*, 1996; Hungerbühler, 1997).

The Western Cordillera and Coastal area are part of the allochthonous oceanic terranes accreted to the Andean margin during Late Cretaceous-early Tertiary times (Lebrat *et al.*, 1987; Cosma *et al.*, 1998; Hughes and Pilatasig, 1999; Reynaud *et al.*, 1999). The Western Cordillera is made of oceanic plateau and island arc magmatic rocks and their Cretaceous-Eocene flysch cover, overlain and crosscut by continental arc magmatic rocks. The Coastal area is characterized by four main Neogene fore arc basins (Borbón, Manabí, Progreso, Guayaquil; Fig. 1) related to dextral strike-

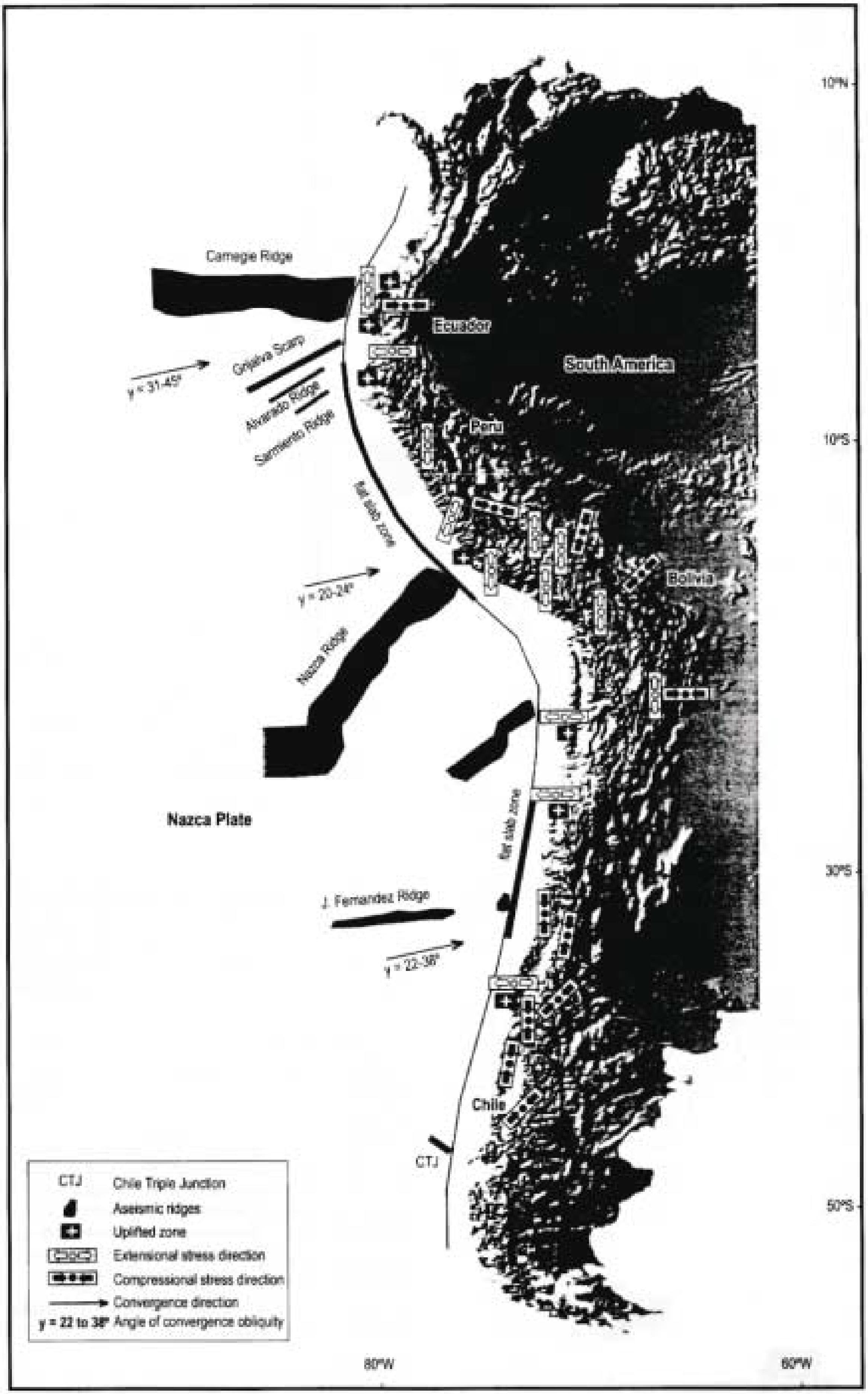


FIGURE 44 - Principal directions of stress deduced from structural analysis of Quaternary and active faults.

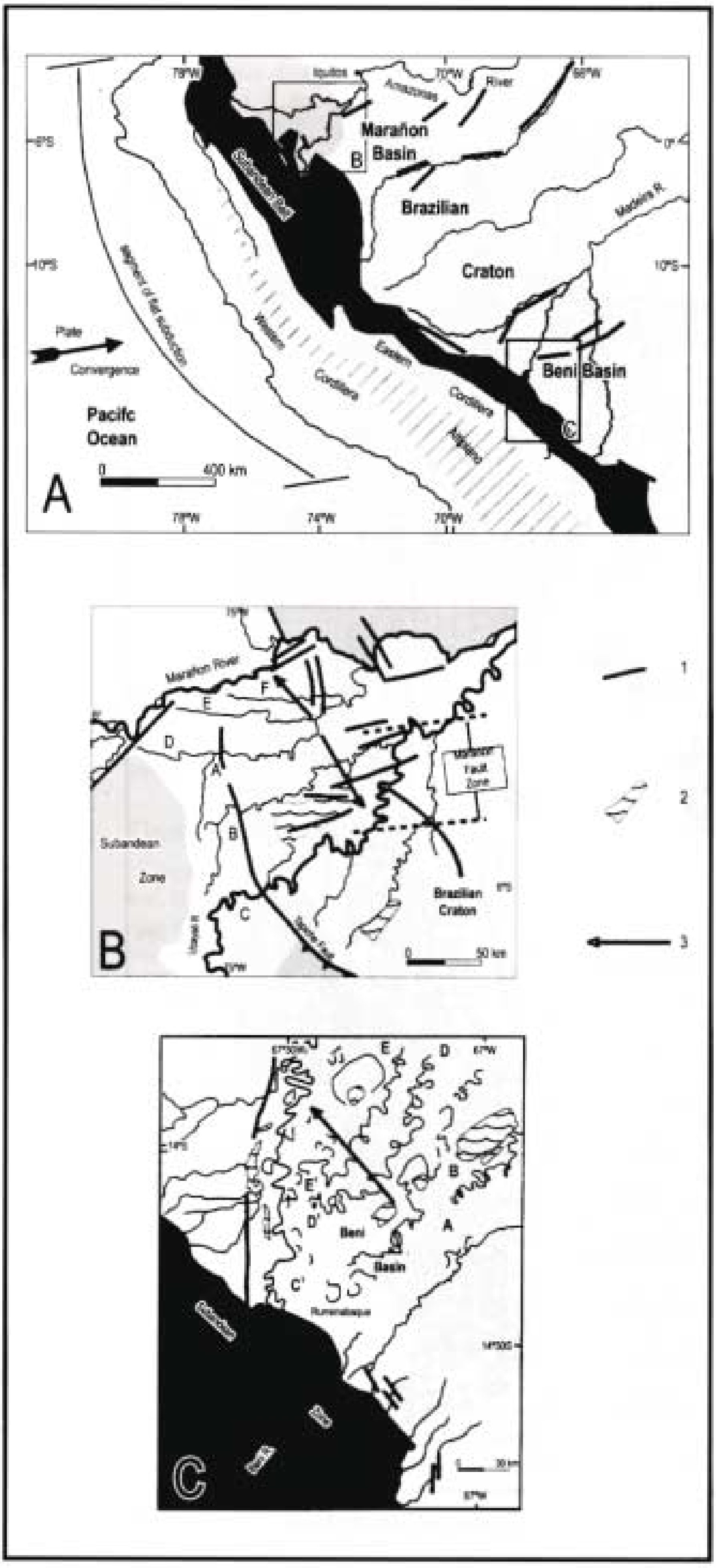


FIGURE 45 - Geodynamic environment of the Marañon and Beni basins.
A: Location. B: Rivers and basement structures in the Marañon Basin. Shift of the inflection points from A to C (Ucayali River) and from D to F (Marañon river). 1 - Tapiche Fault, front of Subandean thrust tectonics;
2 - main basement faults;
3 - structural elongated lakes ;
4 - main geographical direction of river shifts.
C: Beni Basin and shifts of the Beni River. A to E show the successive lateral shifts of the river, which correspond to the deflection points C' to E'. Lakes in the Beni Basin are not of tectonic origin.

slip displacements (Deniaud *et al.*, 1999). The Gulf of Guayaquil is the deepest Neogene fore-arc basin. It corresponds to a pull-apart basin developed between the Dolores-Guayaquil Megashear Zone to the E, and the oblique convergent Nazca-South America Plate boundary to the W (Deniaud *et al.*, 1999).

Timing of Neogene deformations

Neogene deformations have been recorded in the fore-arc, intermontane and foreland basins. The creation of the intermontane basins started at about 28 - 26 Ma (Marocco *et al.*, 1995), like the Bolivian Altiplano Basin. In the coastal area, the evolution of the Manabí and Progreso Neogene basins began in the early Miocene. In the Amazonas Basin, stratigraphic data are insufficient to specify the age of the onset of the Neogene foreland basin. It is marked by the eastward wedge of the subaerial Arajuno Formation (Petroproducción seismic information), of probable early Miocene age.

The upper Miocene is characterized by the closure and the piggyback evolution of the intermontane basins (Marocco *et al.*, 1995), and the closure of the Manabí and Progreso fore-arc basins (Deniaud *et al.*, 1999). At the same time, the Amazonas foreland was mildly deformed and invaded by marine incursions (Hoorn *et al.*, 1995).

The Pliocene showed an acceleration of the deformation and marked the onset of the strongest orogenic stage of the Andes, which is still active. The opening of the pull-apart basin of the Gulf of Guayaquil started during the Pliocene, and sedimentation rate reached a maximum (8600 m/Ma in the depocenter) in the lower Pleistocene (Deniaud *et al.*, 1999b). The uplift of the Subandean Zone (Napo and Cutucú uplifts) occurred during this period and continues presently.

Overview of the neotectonics of the North-Central Andes (Ecuador, Peru, Bolivia, and Northern Chile)

The main features of the Andean Cordillera were acquired during the Miocene, and few changes occured since then. However, significant modifications of the topography are produced by neotectonic deformations, resulting in the present-day topography. During this period, altiplano basins are formed or maintained in Ecuador and Bolivia, the Nazca and Carnegie aseismic ridges are introduced in the subduction zone, leading the coast to rise, and the two depressions of the Marañon and Beni basins are individualized, giving birth to the present Amazonas River.

This evolution of the landscape is better approached and understood considering three different aspects of neotectonic studies and methods. The first determines the state of stress (Fig. 44), as deduced from fault analysis. As far as Quaternary terranes are considered, a comparison with the present state of stress can be carried out. The second deals with the vertical movements along the coast, as determined from the study of marine terraces. The third type of neotectonic studies is dedicated to the foreland basins, where river locations and shifts are controlled by neotectonic deformation of the basin surface (Fig. 45).

The convergence vector is oblique with respect to the plate boundary zone. The mode of the oblique accommodation is problematic, specially the relationship between the overriding plate deformations and the subduction. The geometry of the coast and the subduction system suggests that it strongly controls the building of the range. The coastal areas of South America are generally submitted to extensional tectonics, mostly because it lies over the subducted plate without lateral constraints. However, the direction of extension is variable. In areas of oblique convergence and relatively wide coastal zone (comprised between trench and Western Cordillera) it is orthogonal to the direction of plate convergence (Ecuador, southern Peru). In Chile, where there is a narrow belt between the trench and the Main Cordillera, the extension is orthogonal to the margin, and interpreted as related to gravitational post-seismic effects.

In the coastal region of Ecuador, the stress pattern is dominated by a N-S extension (Dumont *et al.*, 1997), due either to the general northward escape of the Andean Block, or, more locally, to the northward increasing obliquity of the convergence, from the Gulf of Guayaquil to Colombia. The triangle-shaped Andean Block accommodates the deformation at the triple junction between the South American, Caribbean, and Nazca plates. At the southern tip of the Ecuadorian Coastal Block, which forms the southern corner of the Andean Block, the Gulf of Guayaquil opened as a result of the right lateral movement of the Andean Block with respect to the South American Plate. This right lateral movement is accommodated along the Pallatanga Fault, which extends northeastwards towards the Interandean Depression (located between the Eastern and Western Cordilleras) and farther N to other fault segments (*e.g.*, Chingual-Sofia Fault). Southwestwards, the Pallatanga Fault extends into the Gulf of Guayaquil, by the means of a system of transcurrent and normal faults. The calculated average Quaternary extensional rate of the Gulf of Guayaquil is of 2.5 ± 1.1 mm/y.

The subduction of the Carnegie Ridge during early and middle Pleistocene is an important parameter of the coastal uplift. Several Quaternary abrasion surfaces at elevations ranging from 7 m to as much as 330 m (*e.g.*, the surfaces of the Tablazos Formation) are observed between the Gulf of Guayaquil and Esmeraldas, suggesting a maximum uplift rate of about 0.2 mm/y during the Quaternary. Along the Pacific coast of Peru, the Quaternary faults evidence a N-S trending extension. In the Pacific lowlands of southern Peru, this state of stress is about neutral, due to a topographic effect related to the proximity of the deep Peru-Chile Trench. On the northern Peruvian Coast, the present-day elevation of the abrasion surfaces suggests an uplift rate of 0.2 mm/y during the Quaternary. In southern Peru, in front of the Nazca Ridge, uplifted marine terraces located at 300 to 700 m high, suggest an average uplift rate of 0.18 mm/y and a maximum uplift rate of 0.7 mm/y for the same period (Macharé and Ortlieb, 1993).

In Chile, in localised coastal areas, which are the closest to the trench (80 to 100 km); the observed state of stress during the upper Pleistocene is an E-W extension. This E-W *stretching is related to uplifted terraces, located over a* crustal bulge due to the subduction. Thus the extension is interpreted as resulting from an accommodation of the rising topography, related to body forces. The E-W trending sxx is s3 (sHmin, Horizontal minimum principal stress, or tensional deviatoric stress), syy N-S trending is s2 (intermediate deviatoric stress) and szz, vertical, is s1 (Hmax, maximum principal stress or compressional deviatoric stress). The state of stress is szz > syy > sxx. In northern Chile (23° to 27°S) the Quaternary marine abrasion surfaces, located at an elevation of 200 m, suggest a maximum uplift rate of about 0.2 mm/y for the Quaternary (Ortlieb *et al.*, 1996).

In the Main Range of Central Andes, present-day stress field and crustal deformations are not homogeneous along strike. In the whole Andean range of Ecuador, the present-day stress field appears to be homogeneous and the Quaternary dominant tectonic regime is an E-W trending compression. Near the trench, the state of stress is s1 = N81°E; in the High Cordillera between 0° and 1°S, the state of stress is N77°E < s1 < N120°E; and in the Subandean Zone, s1 is N99°E (Ego *et al.*, 1996). In the northern part of Ecuador, along the right lateral Chingal-La Sofia Fault, the lateral slip rate displacement is of 7 ± 3 mm/y for the last 37 ka BP (Ego, 1995). In the Interandean Depression, in the restraining bend of the Latacunga Zone, the shortening rate is of 1.4 ± 0.3 mm/y since 1.4 Ma (Lavenu *et al.*, 1995). Along the right lateral strike-slip Pallatanga Fault, the horizontal rate motion is of 4 ± 1 mm/y for the same period (Winter and Lavenu, 1989; Winter, 1990; Winter *et al.*, 1993).

The present-day state of stress in the Peruvian Andes has been deduced from the structural analysis of Quaternary and active faults and seismic data (Sébrier *et al.*, 1985, 1988, Mercier *et al.*, 1992). The crustal deformation of the High Andes is characterized by normal faulting, excepted within the Eastern Cordillera of central Peru, whereas the western and eastern boundaries of the High Andes (fore-arc and foreland) are characterized by thrust mechanisms that indicate compressional deformations.

In the High Andes, two tectonic regimes occur. In the Western Cordillera, recent and active deformations result from a N-S trending extensional tectonics. In the Eastern Cordillera, seismicity and active strike-slip faults result from both a N-S trending extension and an E-W trending compression. In the Subandean Zone, reverse faulting is in agreement with an E-W trending compression. Close to the trench, at the contact between the Nazca and South American plates, focal mechanisms of earthquakes evidence an E-W trending compression roughly parallel to the convergence between the two plates. In southern Peru, the state of stress in the High Andes as well as in the Pacific Lowlands results from a N-S trending extension.

Thus, the state of stress in the Andes of central Peru and those of the Andes of southern Peru may be interpreted as an effect of compensated high topography. However, compressional tectonics affects the High Andes of central Peru but not those of southern Peru.

In central Peru, the stress model is such that the vertical stress szz increases with the topography and the compressional stress sHmax is considered constant and *trends* E-W, *i.e.*, *roughly parallel to the convergence* direction. In the Western Cordillera of the High Andes, szz becomes 1, Hmax is s2 and Hmin is s3, trending N-S. In the Eastern Cordillera, szz becomes s2, sHmax is s1 and trends E-W, and sHmin is s3 trending N-S. In the Eastern Cordillera, the compressional strike-slip faulting may be explained by an effect of topography, between the high Western Cordillera and the Subandean Lowlands. The Eastern Cordillera being undercompensated, its elevation should be lower in an isostatic equilibrium. The change between the compressional regime in the Subandean Zone and the strike-slip regime in the Eastern Cordillera should take place between 1000 and 2000 m in elevation.

In southern Peru, the state of stress is different in the High Andes; there, szz is s1, sHmax is s2 and sHmin is s3 and trends N-S, *i.e.*, roughly perpendicular to the convergence vector (Sébrier *et al.*, 1985, 1988; Mercier *et al.*, 1992). In the Bolivian High Andes, during the Pliocene (6 - 3 Ma), the tectonic regime was extensional; s3 is sHmax and trends E-W (Lavenu and Mercier 1991). During uppermost Pliocene-lower Pleistocene (3 - 2 Ma) a compressional tectonics affected this region, which is characterized by s1 trending E-W, parallel to the convergence. This tectonic event is characterized by a weak deformation, and by the reactivation of old faults as reverse and strike-slip faults. This stress regime is followed by a nearly coeval N-S trending compressional tectonics. Since lower Pleistocene to Present, the whole range is affected by an extensional tectonics with s3 trending N-S (kilometric normal faults with hectometric throw). In the Altiplano and the High Andes, Quaternary tectonic regime is extensional with sHmin = s3 and trends N-S, sHmax = s2 and trends E-W, and s1 is vertical. As in Peru, this stress field results from body forces due to a compensed high topography. The E-W trending horizontal stress sHmin = s2 is roughly parallel to the convergence direction; szz (s1) increases with the topography due to the range load.

The intermediate zones (*e.g.*, Tarija, 1900 m in elevation) are characterized by two superposed stress regimes. One is a relatively weak strike-slip compressional stress, with s2 vertical, s1 = sHmax, E-W trending, and s3 = sHmin, N-S trending. The other one, more intensive, is an extensional, axial stress, with s1 vertical; s2 trends E-W and is equivalent to s3, which trends N-S. If we admit that the vertical stress szz is the result of the weight of an isostatically compensated topography, the strike-slip state of stress is consistent with the intermediate location of the basin, between the Subandean Zone and the High Andes (Lavenu and Mercier, 1991).

Along the Chilean coast, the Quaternary regime is extensional and of an E-W strikes. This deformation characterizes the westernmost portions of the continental fore-arc, close to the trench axis (80 km). This deformation does not appear to be directly linked to boundary forces due to the convergence, but could be the consequence of co-seismic crustal bending with subduction-related earthquakes. It could be topographic accommodation to the uplift of this part of the coast (body force due to topography), sxx striking E-W becomes s3, syy striking N-S is s2, and szz is s1. The state of stress is such that szz > syy > sxx. This phenomenon could be related to the zones of maximum coupling between the oceanic and continental plates in the Central Andes, which could act as a buttress zone.

The partition of the deformation across the plate boundary zone shows that the tectonic regime of the Quaternary is more complex than previously recognized. In the southern Andes (Chile), as well as in the northern Andes (Ecuador), the Cordilleran segments, linked to large strike-slip faults and high angle convergence obliquity, slides toward the North. A part of the energy, transmitted from the subducting plate to the overriding plate, is absorbed by the free escape of fore arc slivers, parallel to the margin. The lack of important crustal thickening and widening of the range characterizes these parts of the Andes. On the contrary, in the Cordilleran segments linked to low angle convergence obliquity, the progressive stop of the lateral movements is due to buttress zones and the energy is absorbed by the crustal thickening and widening of the range (Bolivia).

The Subandean Zones of the Central Andes are dominated by a compressional stress regime. In the Subandean Zone of Ecuador, the Quaternary stress field is compressional and trends E-W. In the Subandean Zone of central Peru, reverse faulting is in agreement with an E-W trending compression, whereas deformations result from a *N-S trending compression (s1 is sHmax) in the Subandean* Zone of southern Peru. In the Subandean lowlands of Bolivia, deformations are compressional, with sHmax as sxx, is s1, E-W trending.

The Marañon and Beni basins are respectively situated at the northern and southern ends of the Peru-Bolivia Andean segment (Fig. 45A). This segment corresponds to the flat slab subduction of the Nazca Plate beneath the Andes. Specific structures of the foothills of the Subandean Zone control two basins, each one having only one outlet, the Amazonas and Madeira rivers, respectively. The flat surface of these basins shows a complex network of flowing and fossil river traces. These active and abandoned fluvial traces are used, together with neotectonic, seismotectonic and subsurface structural data (Dumont and Fournier, 1994), to determine the neotectonic evolution of the Peruvian and Bolivian foreland basins (Dumont, 1996). The phenomena exemplified below refer to short term neotectonics, occurring during the Holocene (0.1 - 0 Ma).

In both basins, recent directional shifts of the main rivers are controlled by the offset of faults. The Ucayali River flows northwards along a N-S intra-subandean basin, then enters the Marañon Basin where it has been deflected to the NE (Fig. 45B). The successive deflection points shifted upstream and along the foothills. The line joining the deflection points (Fig. 45B, points A, B and C) lies just behind the Andean Frontal Thrust, represented here by the Tapiche Fault (Fig. 45B). Contemporaneously, the Marañon River was deflected to the NE, lined up with the straight, NE trending lower reaches of the Huallaga River, which is controlled by a fault observed on satellite images. In the eastern part of the Marañon Depression, the rivers trend NE-SW, parallel to the strike of the main basement faults of the Marañon Structural Zone (Laurent, 1985). Elongated lakes are situated over structurally downwarped blocks (Dumont, 1993).

Successive shifts of the Beni River (Fig. 45C) show the northward migration of the deflection point made by the Beni River entering the basin. A N-S trending fault crossing the foothill margin controls this downstream increment. The present regional state of stress in the Subandean region is roughly E-W, except in the southern part of the Marañon Basin where it is NE-SW (Assumpção, 1992). Quaternary normal faults displaying a NNW-SSE extension in the distal part of the Marañon Basin, as well as rising of bulges on the eastern margins of both basins are consistent with the present-day state of stress. The interpretation emphasizes that in the distal areas of the Marañon Basin the river traces are guided by topographic lows along tensional faults, or basement blocks, uplifted or downwarped by tensional faulting. Near the piedmont, river shifts are controlled by the increment of fault movements toward the basin. However, the effect of faults in the near piedmont is more difficult to explain than in the distal areas. Topographic effect between the High Andes and the foreland basin (Assumpção and Araújo, 1993) may explain that the geometry of active faults on the foothills piedmont depend also on the local trend of the Cordillera. The interpretation of the successive shifts involves the knowledge of paleoclimate oscillations. It appears that a river moves toward a new formed depocenter at the onset of a wetter period, and can stay in place, *even if tectonic deformation progress, during* relatively dryer periods (Schumm *et al.*, 1998).

In summary, the study of the Recent state of stress in the Andes shows several types of behaviour of the continental plate along the active margin. This behaviour is linked to the dip of the subducted plate, the obliquity of convergence between the oceanic and continental plates, the body forces and boundary forces, the presence or absence of buttress zones in the upper plate, and the possibility for the coastal blocks of free escaping (Fig. 46).

In Ecuador, where the convergence obliquity is very high (g = 31° to 45°), the Coastal Block is pushed northwards, and is affected by a N-S trending extension. Since the elevation of the Andean range is relatively low, the topographic effect is weak. The range is separated from the Coastal Block by a large strike-slip fault, and an E-W trending compressional stress developed, due to boundary forces.

In Peru and Bolivia, convergence obliquity is relatively low (g = 20° - 24°). The High Andes, which present the highest average altitude, are affected by a N-S extension, mainly due to the body forces. A compressional regime is observed only along the boundary between the range and the Brazilian Shield (boundary forces).

In central and southern Chile, convergence obliquity is intermediate (g = 22° to 30°). The fore-arc and intra-arc zones of the Cordillera, the topography of which is lower than in Peru and Bolivia, are affected by a N-S to NE-SW compression.

Regarding the uplifted marine terraces, the type and rate of vertical movements are thought to be relatively independent on the rate and direction of convergence, the convergence obliquity, and the age and dip of the subducted plate (Macharé and Ortlieb, 1993). Conversely, these vertical movements are tightly dependent on the morphology of the subducting plate (aseismic ridges), and on the structure and

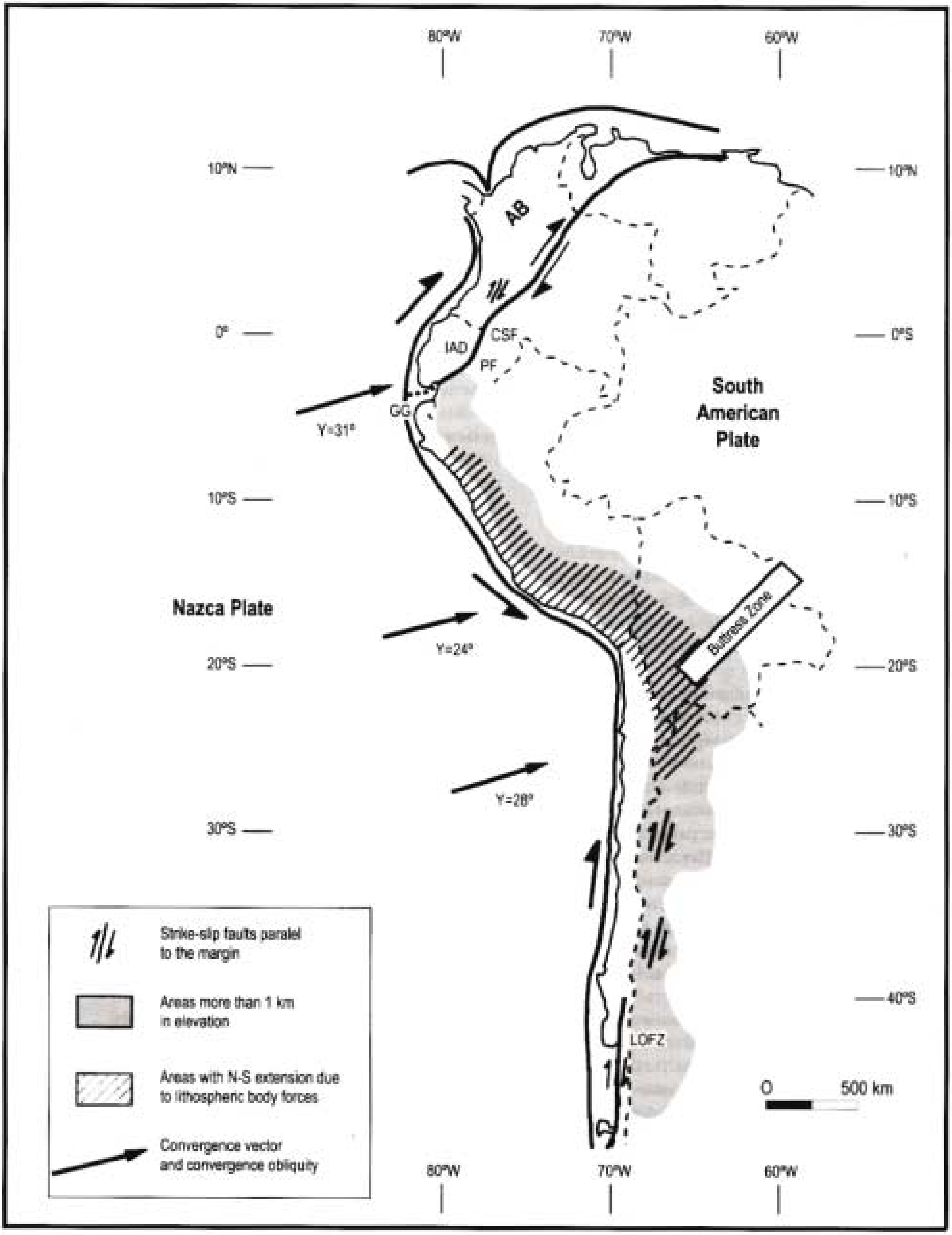


FIGURE 46 - Geodynamic features and relationships between convergence obliquity and Andean Cordillera. AB - Andean Block; CSF - Chingual-La Sofia Fault; GG - Gulf of Guayaquil; IAD - Interandean Depression; LOFZ - Liquiñe-Ofqui Fault Zone; PF - Pallatanga Fault (after Ego, 1995).

density distribution in the overriding plate. In Chile, as well as in Peru, the vertical motion of coastal areas is dependent on the distance from the coast to the trench, and this deformation characterizes the westernmost portions of the continental fore arc, 80 km close to the trench axis. The relationships between Quaternary vertical movements and seismic activity are still poorly understood.

Conclusions : Geodynamic Processes of the Northern-Central Andean Orogeny

Plate Kinematics Framework

The evolution of the central and northern Andean system can be divided into four main periods with different sedimentary, tectonic and magmatic characteristics, indicating distinctive geodynamic situations and convergence directions.

Late Permian - Late Jurassic: Tethyan Period

The Early Mesozoic evolution of the Andean margin is influenced by the Tethyan rifting and evolution (Jaillard *et al.*, 1990, 1995), and is characterized by the onset of a southeastwards subduction along the Colombian, Ecuadorian and Peruvian margins (Fig. 47). After the Late Paleozoic coalescence of Pangea, the Triassic evolution of the northern and central Andes was dominated by an extensional regime responsible for the creation of grabens and the extension of alkaline volcanics. This tectonic context is clearly related to the westward propagation of the Tethyan break-up between Laurasia and Gondwana.

The Early Jurassic evolution of the northern Central Andes was dominated by the destruction of the Late Triassic-Liassic carbonate platform, caused by a general extensional tectonic activity that progressed diachronously southwards. This is thought to have been induced by the rifting of the E-W trending Tethyan system. Meanwhile, no significant absolute motion of the South American Plate occurred relative to the surrounding continental plates (Africa, North America).

Between late Early and early Late Jurassic times, the Tethyan breakline resulted in the opening of NE-trending oceanic-floored rhombochasms (Alpine and central Atlantic oceans, Bernoulli and Lemoine, 1980) linked by E to ENE trending sinistral transform zones (*e.g.*, the Caribbean Transform Zone). The opening of the central Atlantic Ocean began before the late Middle Jurassic (157 Ma, Klitgord and Schouten, 1986) and possibly as early as latest Liassic (180 Ma, Scotese *et al.*, 1988). In the Andes, this period (190 - 140 Ma) was marked by the emplacement of I-type plutons and calc-alkaline volcanics along the NNE trending Ecuadorian-northern Peruvian margin, which should have been coeval with an active subduction beneath this part of the Andean margin. According to the pre-break-up reconstruction, this situation can be interpreted in two ways:

1 - If the Colombian segment was facing continental blocks, the subduction must have involved the new oceanic crust of the Tethyan arm created between the Colombian segment and these blocks (Jaillard *et al.*, 1990; Litherland *et al.*, 1994).

2 - If the Colombian segment directly faced the oceanic paleo-Pacific Plate, subduction was probably active before the Early Jurassic, and the creation of a magmatic arc may have resulted from more rapid subduction, due to an accelerated accretion rate in the paleo-Pacific system.

Whatever the case, the roughly southeastward subduction beneath the Ecuadorian segment must have induced oblique subduction along the Peruvian margin, associated with a strong sinistral strike-slip component, manifested by the creation of the large NW trending south-Peruvian turbiditic pull-apart basin (Vicente *et al.*, 1982) and by transtensional features in the back-arc zones (Sempere *et al.*, 1998; Fig. 47). The Kimmeridgian-Berriasian time-span is transition period. Along the Colombian-Ecuadorian segment, this period was marked by accretions of displaced terranes, compressional deformation, and the end of magmatic activity, while along the Peruvian segment varied tectonic events were associated with the resumption of subduction-related volcanic activity (Aspden *et al.*, 1987; Jaillard *et al.*, 1990, 1995). All this clearly resulted from an important, global-scale geodynamic change. In the west-Tethyan realm (central Atlantic, Alpine oceanic ridges), spreading rates significantly decreased (Olivet *et al.*, 1984; Klitgord and Schouten, 1986; Savostin *et al.*, 1986). If a Tethyan-Colombian oceanic arm did exist, the motion vector of the Phoenix oceanic plate was the sum of the expansion vectors of the Tethyan and Pacific ridges (Duncan and Hargraves, 1984). As a result, slowdown of Tethyan expansion would have induced a northeastward convergence between the Phoenix and South American plates (Duncan and Hargraves, 1984; Jaillard *et al.*, 1990). Moreover, the outpouring of a large oceanic plateau along the Pacific Ridge in Tithonian times may have modified the accretion direction of the East-pacific paleo-plate (Nakinishi *et al.*, 1989).

Early Cretaceous-Paleocene: South Atlantic Period

During this period, the development of the South Atlantic Ocean controlled the westward drift of the South American Plate and the variations in the convergence rate along the subduction zone. These are thought to determine the sedimentary, tectonic and magmatic evolution of the Andean margin. During the Early Cretaceous, the sudden arrival of a great amount of east-derived sands can be interpreted as the result of the westward doming of eastern South America due to the incipient rifting of the South Atlantic Ocean. Although no reliable geodynamic reconstruction is available, the lack of significant tectonic or magmatic activity along the Pacific margin of the South American Plate N of 18°S would indicate a slow, steep-dipping subduction of the paleo-Pacific slab.

The definitive opening of the South Atlantic Ocean at equatorial latitudes during Albian times (Emery and Uchupi, 1984; Scotese *et al.*, 1988) induced the beginning of the absolute westward motion of the South American Plate. Therefore, as noted by various authors (Frutos, 1981; Mégard, 1987; Soler and Bonhomme, 1990), the beginning of compressional deformation along the Peruvian and Colombian segments during the Late Albian (100 - 95 Ma) coincides with the onset of the trenchward motion of the upper plate (Uyeda and Kanamori, 1979; Cross and Pilger, 1982; Jarrard, 1986).

The Albian-Turonian period coincides with a period of high convergence rate and with the mid-Cretaceous magnetic quiet zone (Larson, 1991). In the Central Andean margin, it is characterized by important magmatic activity, a high average subsidence rate (Jaillard and Soler, 1996) and probably significant dextral strike-slip movement (Bussel and Pitcher, 1985). The latter are probably related to the north-northeasterly motion assumed for the paleo-Pacific slab during Late Cretaceous times (Pilger, 1984; Gordon and Jurdy, 1986; Pardo-Casas and Molnar, 1987; Fig. 48). This convergence direction accounts for the lack of Cretaceous arc magmatism along the NNE trending Ecuadorian margin. The NW trending subduction zone along the Peruvian margin extended probably northwestwards into the oceanic domain as an intra-oceanic subduction zone allowing the development of Cretaceous island arcs, such as those known in western Ecuador.

The Coniacian-late Paleocene interval was marked by a significant slowdown in the convergence rate (85 - 75 Ma), followed by a period of low mean convergence rate between the Phoenix and South American plates (80 - 58 Ma, Soler and Bonhomme, 1990). This period was characterized, however, by the beginning of the Late Cretaceous Andean compressional events. In the northern part of the studied area, significant deformation was restricted to the fore-arc and arc zones. However, the Late Cretaceous and Paleogene tectonic events are coeval with a noticeable decrease of the subsidence rate in the back-arc areas, which favoured detritital deposits, sedimentary hiatus and unconformities. This suggests that during this period of oblique subduction, most of the convergence was accommodated by lateral displacements of fore-arc slivers along the edge of the margin, rather than by shortening and thickening of the upper plate. In the southern Peru and northern Chile, however, deformations seem to have been more important, and a significant increase of the subsidence rate in Bolivia is interpreted as the result of a foreland-type subsidence related to the deformation and tectonic loading of the margin (Sempere, 1994).

Late Paleocene-late Oligocene: Transition Period

The late Paleocene to late Oligocene interval (55 - 25 Ma) is a key period in the whole Andean evolution. Displaced terranes were accreted or obducted along the Colombian segment, important compressional deformation occurred in the Andean realm and sedimentary gaps and unconformities occurred in the eastern domains.

These events coincided with global plate kinematic reorganization (Scotese *et al.*, 1988). In late Paleocene-Eocene times, the convergence of the paleo-Pacific plate changed from N or NNE to NE or ENE (Pilger, 1984; Pardo-Casas and Molnar, 1987), provoking the change from a dominantly dextral transform zone to a nearly normal convergent regime in the Colombian-Ecuadorian segment (Figs. 47 and 48). Such dramatic changes explain how terranes that were previously situated W of the Andean margin, were drifted eastwards and accreted to the northern Andean margin at that time (Jaillard *et al.*, 1995). On the other hand, the more easterly convergence direction allowed subduction to take place beneath the NNE trending Ecuadorian margin and triggered the resumption of arc magmatism in this area by early Eocene times.

A second major reorganization occurred by late Oligocene times, as the Farallón Plate splitted into the Cocos and Nazca plates (Wortel and Cloething, 1981). Convergence direction evolved from ENE to nearly W-E and convergence rate subsequently increased (Pilger, 1984; Pardo-Casas and Molnar, 1987; Tebbens and Cande, 1997; Somoza, 1998; Fig. 48). As a consequence, this period is marked by a significant increase of the orthogonal component of the convergence velocity between the paleo-Pacific oceanic plate and the Andean margin. The subsequent increased coupling along the subduction zone was responsible for a significant eastward migration of the deformed zone, which involved the former arc zone and proximal back-arc areas, *i.e.*, the present-day Eastern Cordillera of Ecuador, Western *Cordillera of Peru and Bolivia, and eventually the Eastern* Cordillera of Bolivia. This suggests a significant decrease of the play of lateral displacement along the Andean margin in the accommodation of convergence, and a correlative increase of the shortening and thickening of the overriding continental plate. Note that this period does not correspond to the subduction of a younger plate (Fig. 49).

Late Oligocene to Present: Pacific Period

From the late Oligocene onwards, the northern central Andean margin was completely controlled by the W to WNW motion of South America and the E to ENE motion of the paleo-Pacific Plate, that determined a roughly E-W couple and a nearly normal subduction system (Fig. 48). During this period, the subducting slab is rejuvenating, the convergence rate is relatively high (Fig. 49) and aseismic ridges arrived in the subduction zone (Fig. 3). This geodynamic pattern, which remains relatively stable and differs significantly from the preceding ones, corresponds to the classical Chilean-type convergent margin.

From 30 Ma onwards, the age of the oceanic plate rejuvenated slightly, becoming probably more buoyant, and favouring, therefore, a low-dipping angle of subduction. Although the convergence rate did not change significantly, late Oligocene-early Miocene times are marked by an acceleration, while the Pliocene is marked by a slight deceleration (Fig. 49). Correlation of these rate variations is difficult to link with specific tectonic events. During this period, the deformed zone significantly migrated eastwards and enlarged, eventually involving crustal-scale thrusting in the Peruvian and Bolivian parts of the chain (Figs. 51 and 52). Due to this large-scale thrust movement, the chain was considerably uplifted, most of the present-day altitude being acquired during the last 8 to 9 Ma. On the other hand,

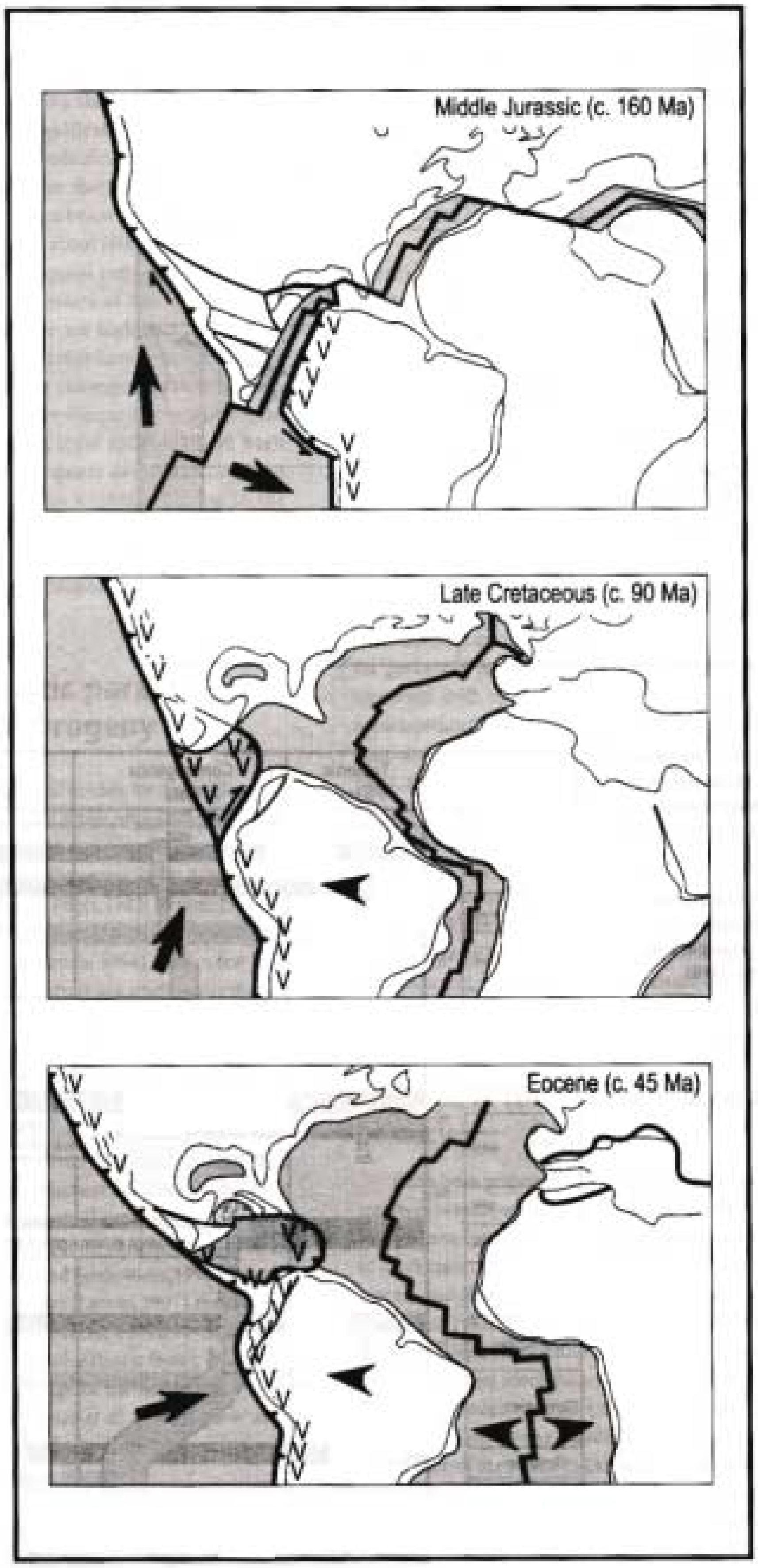


FIGURE 47 - Sketch of the plate tectonic evolution of the Andean margin since Early Mesozoic times (after Jaillard et al., 1990, 1995).

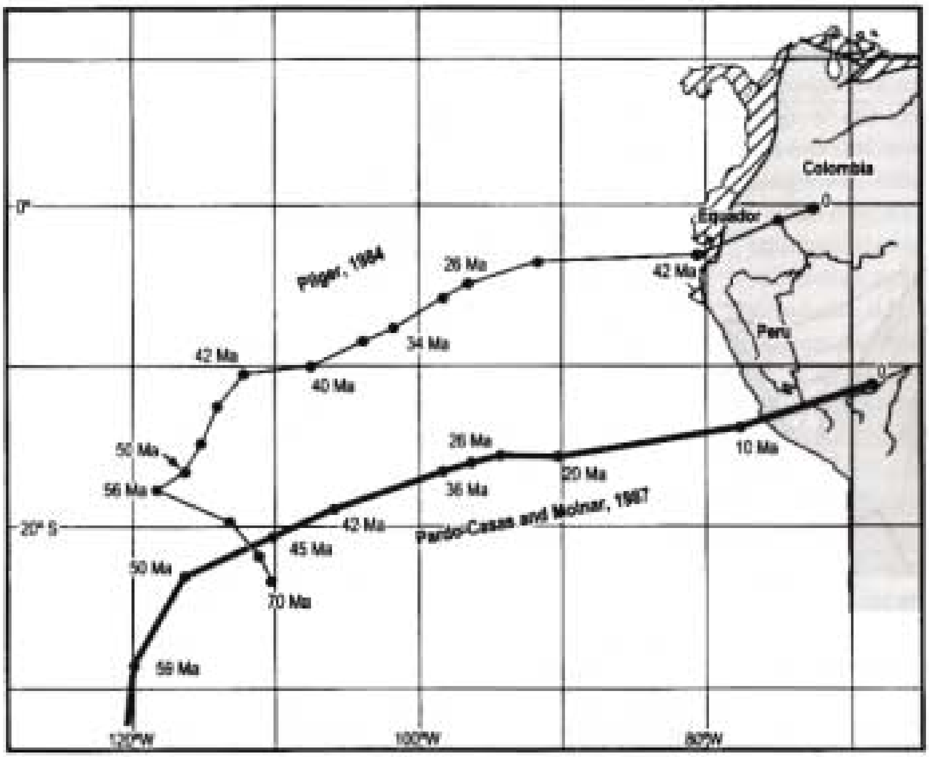
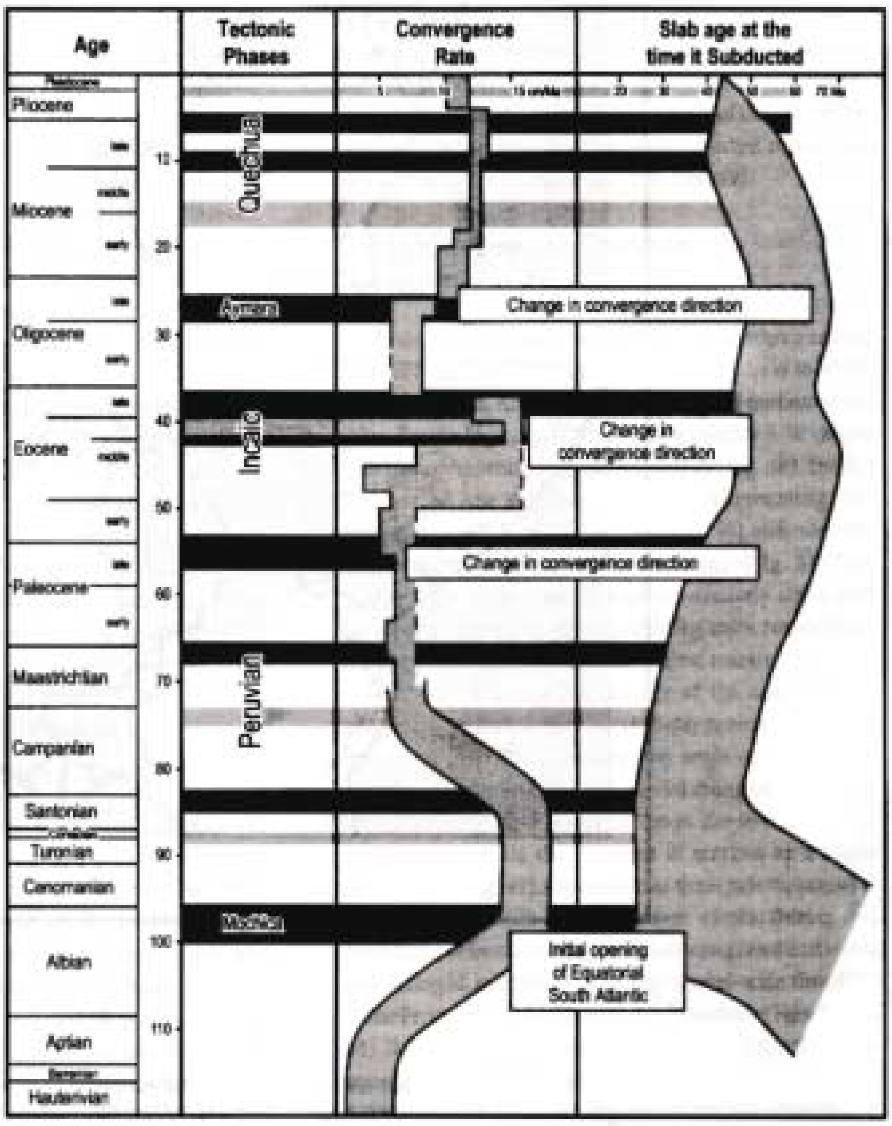


FIGURE 48 - Evolution of the convergence direction of the paleo-Pacific oceanic plate relative to the Andean margin since the latest Cretaceous, (after Pilger 1984; Pardo-Casas and Molnar, 1987).

FIGURE 49 - Evolution of the convergence rate and the age of the slab while subducted, since Middle Cretaceous times (after Soler and Bonhomme, 1990).



subsidence in the eastern basins of Bolivia and southern Peru increased, due to flexural loading. In contrast, subsidence did not increase in Ecuador and even decreased in northeastern Peru (Fig. 39), thus indicating that the tectonic processes differ significantly in both regions. Meanwhile, the fore-arc zones subsided considerably, in such a manner that Pliocene outer shelf deposits are found presently at more than 4000 m below sea-level, indicating an average subsidence rate of about 1000 m/Ma.

In spite of a nearly orthogonal convergence direction, the rate of dextral displacements of fore-arc slivers or terranes of western Ecuador are high (0.5 cm/y). This suggests that such movements may have been much higher in the Late Cretaceous, when convergence rate was much higher (Fig.49) and much more oblique than in the Neogene (Fig. 48). In the same way, local rotations have been significant during the Neogene, demonstrating that contractional shortening was a leading process in the thickening and bending of the Altiplano Orocline, and suggesting that the cumulated amount of rotation may have been significant since the Cretaceous.

Role of kinematic parameters in the Andean Orogeny

Most classical geodynamic models for the origin of the tectonic phases in continental active margins are based on the observation and comparison of various present-day active margins (Uyeda and Kanamori, 1979; Scholl *et al.*, 1980; Uyeda, 1982; Cross and Pilger, 1982; Jarrard, 1986), or through physical modelling (Bott *et al.*, 1989; Whittaker *et al.*, 1992; Cloos, 1993; Shemenda, 1994). Only a few have been elaborated through the study of a single active margin evolution through a long period of time. The study of the Andean margin since earliest Mesozoic times, however, provides some geological constraints on the origin and nature of the tectonic phases of continental active margins.

Plate tectonic reconstructions are poorly constrained for the Late Cretaceous period (Pardo-Casas and Molnar, 1987), especially as regards subduction of ridges, dip of the subducting slab and direction of convergence. However, quantitative approximation of some parameters, such as the convergence velocity (Soler and Bonhomme, 1990) deduced from the global spreading rates (Larson, 1991), the absolute motion of the South American Plate driven by the opening and ridge activity of the South Atlantic Ocean (Nürnberg and Müller, 1991) and the age of the oceanic slab while subducted, calculated by Soler *et al.* (1989), allow us to analyze them in relation to the early tectonic evolution of the northern Central Andean margin.

Age of the subducted slab

Classical models assume that the subduction of a young, buoyant oceanic lithosphere induces a contractional strain in the overriding continental plate (Molnar and Atwater, 1978; Cross and Pilger, 1982; Sacks, 1983). According to Soler *et al.* (1989), the beginning of the contractional period (Albian) and the late Oligocene to Recent contractional phases roughly coincide with the rejuvenation of the oceanic plate subducting at that time. However, the Late Cretaceous and major Paleogene shortening phases occurred during a continuous increase in the relative age of the subducted slab (Fig. 49). Therefore, the lithosphere age of the subducted slab may contribute to the appearance of a long-termed contractional regime, but cannot account for short-termed shortening phases.

Absolute trenchward motion of the overriding plate

As noted by many authors, the opening of the South Atlantic Ocean at the equatorial latitudes during Albian times, which provoked the beginning of the westward shift of the South American Plate, roughly coincides with the initiation of the contractional deformation along the Peruvian-Ecuadorian margin. Thus, this parameter seems to control the long-termed contractional regime of the continental active margin.

As emphasized by Sébrier and Soler (1991) for the late Tertiary Andean contractional phases, only a slight shortening occurs in the Andean retro-arc foreland during the periods of tectonic quiescence, and then most of the westward drift of the South American Plate should be accommodated by an absolute westward overriding of the continental plate over a retreating oceanic slab. On the other hand, the amount of tectonic shortening observed during the contractional phases implies that virtually all the westward drift of the South American Plate is accommodated by the shortening. Therefore, during the contractional phases, the western continental margin of the South American Plate is virtually motionless in an absolute reference frame (*i.e.*, there is a stopping of the slab retreat). This recurrent stopping of the slab retreat, the mechanical origins of which are unclear, might be one of the driving phenomenon of the short-lived contractional tectonic crisis.

Collision of continental or oceanic obstacles

It has been proposed that the arrival in the subduction trench of oceanic or continental obstacles (aseismic ridges, sea-mounts, continental microplates) will lead to blocking of subduction, contractional deformation of the continental margin and plate reorganisation (Scholl *et al.*, 1980; Cross and Pilger, 1982; Ben-Avraham and Nur, 1987). According to Cloos (1993), only continental blocks and oceanic island arcs with a crust more than 15 to 20 km thick or basaltic plateaux of more than 30 km of crustal thickness will provoke a jam in the subduction zone. The current subduction of the 15 km-thick inactive Nazca Ridge results in the extensive subduction erosion of the fore-arc, and local uplift of *c.* 900 m associated with only moderate horizontal compressional stress (Couch and Whitsett, 1981; Macharé and Ortlieb, 1992). Thus, the arrival of moderately high obstacles in the trench seems to have moderate deformational effects on the upper active margin. On the other hand, the accretions of oceanic island arc terranes of Coastal Ecuador (Santonian, late Paleocene, late Eocene) are coeval with contractional phases observed in Bolivia,

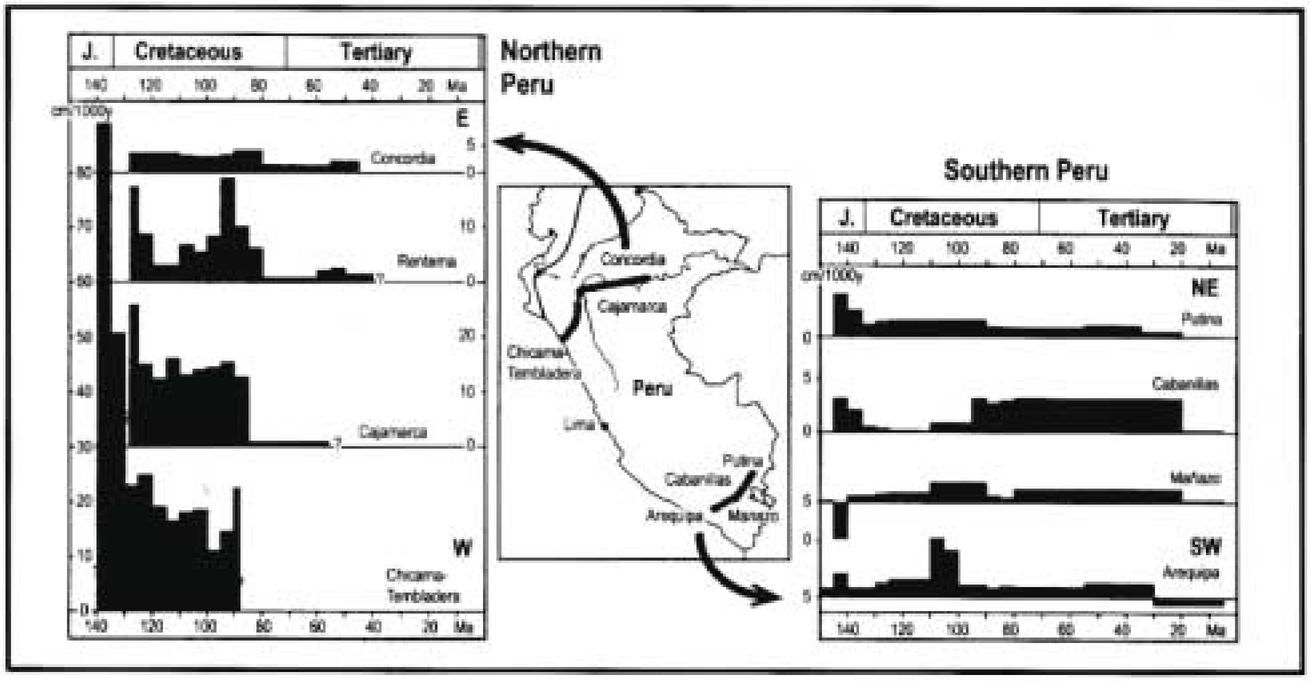


FIGURE 50 - Tectonic subsidence of representative areas of the Peruvian margin (after Jaillard and Soler, 1996).

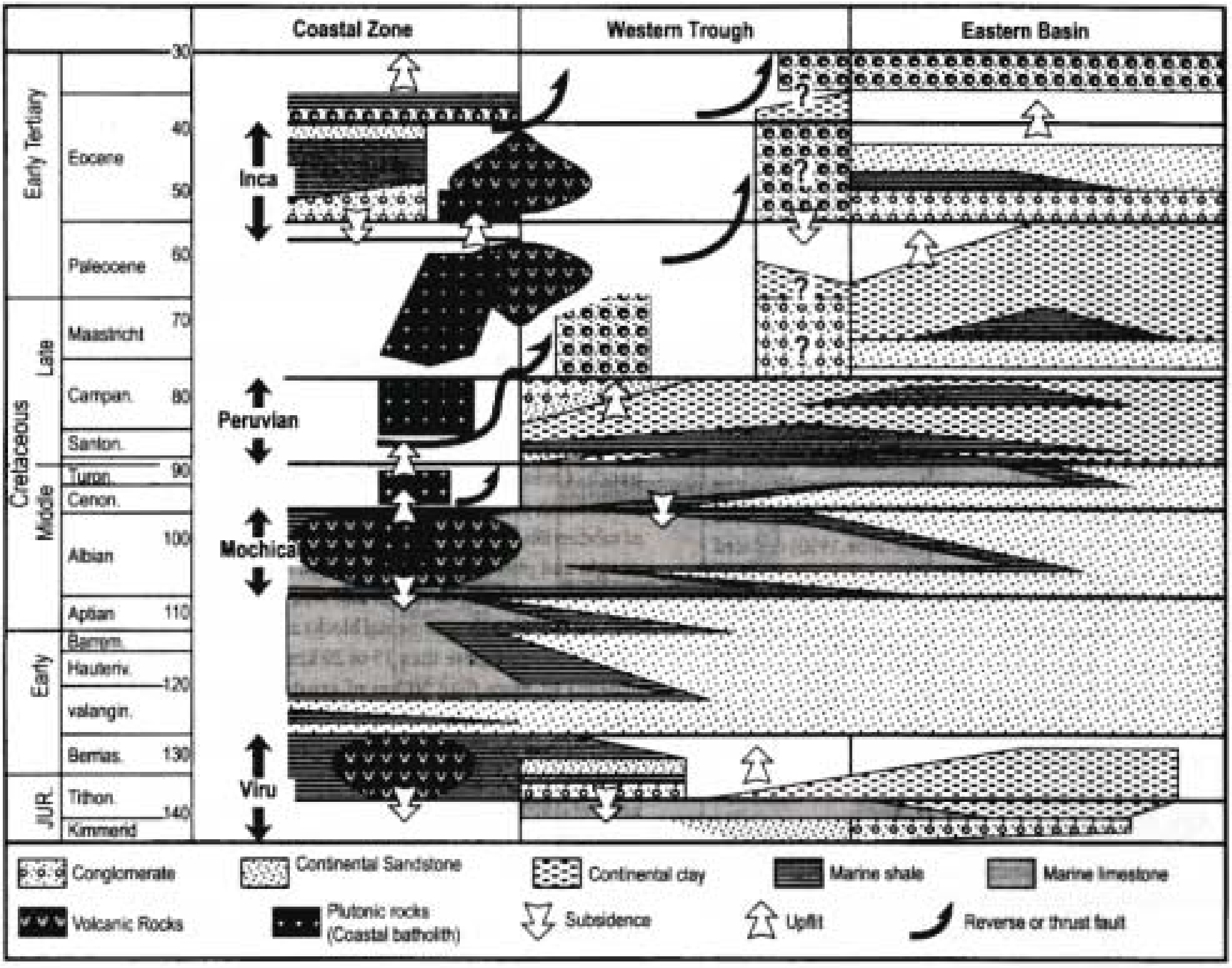


FIGURE 51 - Tectonic, magmatic and sedimentary evolution of the north Peruvian margin between latest Jurassic and Oligocene times (after Jaillard and Soler, 1996).

southern Peru and northern Peru where no collisions are known to have occurred. Thus, in this case, it seems that accretions or collisions of terranes cannot be the cause of regional contractional phases. In the studied area, the fact that contraction took place in non-accretionary settings at the same time as accretions occurred suggests that the accretion events and the coeval contractional phases are consequences of a same global geodynamic mechanism.

Convergence rate

According to Uyeda and Kanamori (1979), Cross and Pilger (1982); Pardo-Casas and Molnar (1987), a rapid convergence between the oceanic and continental plates provokes a compressional stress in the latter. According to Soler and Bonhomme (1990), periods of high convergence rates along the Peruvian margin occurred in Albian-Campanian and late Eocene-early Oligocene times, which coincide roughly with mainly contractional tectonic periods (Fig.49). However, rather than with the convergence velocity itself, the short-lived tectonic events seem to correlate better with changes in the convergence velocity, whichever their sign, either positive (acceleration) or negative (deceleration). If the reconstruction of Soler and Bonhomme (1990) is correct, acceleration occurred in Late Aptian (110 Ma), Late Campanian (75 Ma), early to middle Eocene (50 Ma), latest Eocene (38 Ma), and late Oligocene-early Miocene times (25 - 20 Ma), whereas deceleration occurred in the Late Albian (100 - 95 Ma), Santonian (85 Ma), middle-late Eocene (42 Ma), and Pliocene (4 Ma). All these periods coincide with apparently extensional (Late Aptian, early-middle Eocene boundary) or important contractional tectonic phases. Therefore, short-lived contractional phases as well as extensional tectonic events seem to be mainly controlled by changes in the convergence velocity.

Direction of convergence

The geometry of the geodynamic reconstructions are too poorly constrained to allow a valuable discussion for the Late Cretaceous. The Incaic contractional tectonic phases of late Paleocene (58 - 55 Ma), late middle Eocene (43 - 42 Ma) and late Oligocene age (26 Ma) coincide with successive clockwise rotation in the direction of convergence (Pilger, 1984; Pardo-Casas and Molnar, 1987; Mayes *et al.*, 1990; Tebbens and Cande, 1997). These changes in the convergence direction, which caused successive significant increases of the normal convergence rates seem to have the same effects as those assumed for a convergence acceleration.

Moreover, the important changes in the convergence direction from NNE to ENE by late Paleocene must have induced drastic changes in subduction geometry. The NNE trending Ecuadorian margin changed from a mainly transform to a chiefly convergent regime. This must have induced the eastward drift and accretion of oceanic island arcs along the Ecuadorian margin and the birth of new subduction zones to the W of them (Jaillard *et al.*, 1995). The change in the convergence direction of late middle Eocene age also resulted in a new event of collision of island arcs along the Ecuadorian margin (Bourgois *et al.*, 1990; Hughes and Pilatasig, 1999). Thus, changes in the convergence direction not only control the normal convergence rate, but also play a part in the regional subduction pattern that could in turn influence the tectonic regime. Such changes in the convergence direction during the Paleogene can explain the contemporaneity of the contractional events in non-accretionary settings of the Central Andes and the collisions of island arcs.

Relation convergence rates - subsidence

In northern Peru, periods of slow plate convergence correlate with low subsidence rates (130 to 110 Ma, 75 to 45 Ma, 35 to 25 Ma). Conversely, periods of high convergence velocity are coeval with periods of increased subsidence rate (110 to 85 Ma, 50 to 40 Ma; figs. 49 and 50). This cannot be explained by increased subduction erosion of the deep continental margin (Von Huene and Scholl, 1991), because this latter model is only proved to account for the subsidence of the fore-arc or arc zones, whereas increased subsidence is observed as far as the eastern domain between 110 and 85 Ma (Contreras *et al.*, 1996; figs. 39 and 50). In contrast, these observations are consistent with the thermal model of Mitrovica *et al.* (1989), that assumes that a fast convergence provokes an increase of the subsidence rates along the whole continental margin, through mantle convection (Gurnis, 1992; Stern and Holt, 1994). The lack of such correlation in southern Peru is most probably due to the fact that contractional tectonic events occurred earlier and were stronger than in northern Peru. There, tectonic uplift of the margin by crustal shortening and thickening, and overload tectonic subsidence of the foreland would have prevailed since Senonian times (Sempere, 1994).

Dip of subduction and subduction erosion

The continentward shift of the volcanic front is interpreted classically as a result of the shortening of the continental margin, either by compressive tectonic shortening, or by subduction erosion (Scholl *et al.*, 1980). During Albian and early Late Cretaceous times, the location of the magmatic arc of Peru was stable, indicating that neither significant shortening nor subduction erosion occurred at this time (figs. 20 and 51). The eastward shift of the magmatic arc in the Late Campanian can be explained mainly by the tectonic shortening related to the major Peruvian phase. As a consequence, it seems that no significant subduction erosion took place in Peru before latest Cretaceous, and possibly before Paleocene times, as indicated by the relative stability of the magmatic arc location before this period.

In Eocene times, the ongoing eastward shift of the magmatic belt is associated with its abrupt widening interpreted as the result of a widening of the melting zone in the asthenosphere wedge linked to a decrease of the dip of the Benioff Zone, in turn controlled by the normal convergence velocity (Soler, 1991). The coeval rapid extensional subsidence observed in most of the fore-arc regions by early middle Eocene times is too widespread to result from local tectonic events or paleogeographic effects.

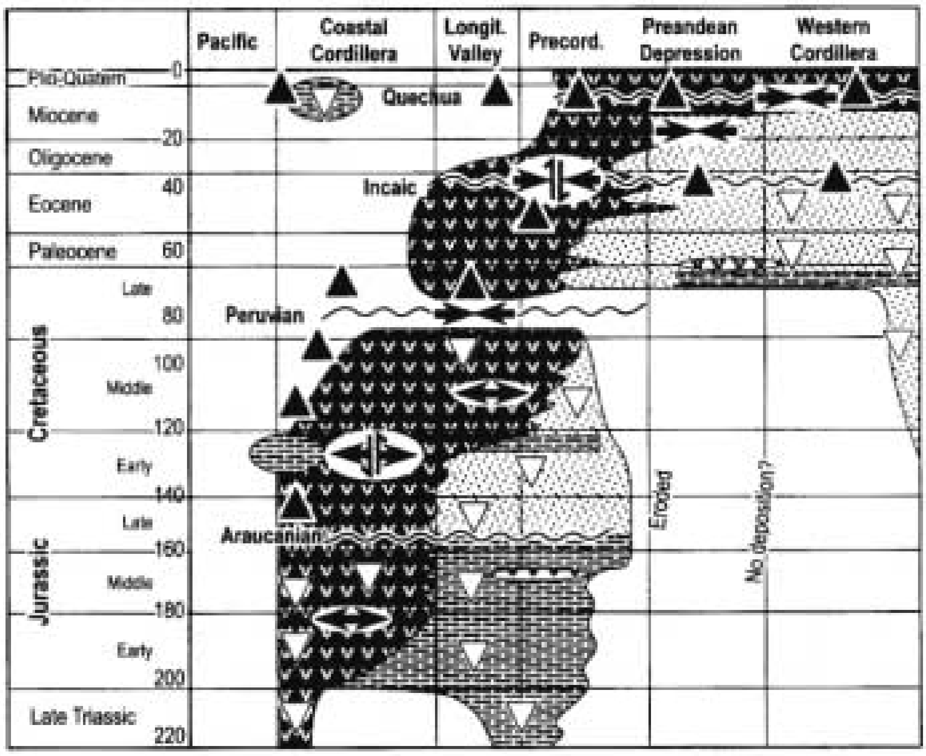


FIGURE 52 - Tectonic, magmatic and sedimentary evolution of the north Chilean margin since Late Triassic times (after Scheuber et al., 1994).

Since it is associated with the ongoing eastward shift of the magmatic front, we propose that both phenomena are due to the subduction erosion of the margin edge (Von Huene and Lallemand, 1990; Von Huene and Scholl, 1991). Because the widening of the magmatic arc coincides grossly with the assumed initiation of the subduction erosion process, we think that a low-angle subduction plan was a necessary condition for the subduction erosion of the Central Andean margin edge. A low dipping subduction zone would result in increased coupling and shear stress at the contact between the plates, inducing a greater potential of abrasive removal at the base of the overriding plate.

Compressional deformation and subduction erosion

Late early Eocene to early middle Eocene times (50 - 45 Ma) are characterized by the creation of subsiding fore-arc basins along the Andean margin of Peru and Ecuador (Figs. 25, 29, and 51). Such widespread phenomena can be regarded as a consequence of the tectonic erosion of the Andean margin, which is well documented for late Tertiary times (Von Huene and Lallemand 1990; Von Huene and Scholl, 1991). The Peruvian tectonic phase (Steinmann, 1929) is followed by the creation of the Campanian-Maastrichtian Paita-Yunguilla fore-arc basin; the creation of the middle Eocene fore-arc basins is subsequent to the late Paleocene contractional deformation, and the creation of the widespread Miocene fore-arc basins is subsequent to the late Oligocene compressional event. These examples suggest that the creation of fore-arc basins frequently occurs soon after contractional deformational events. Therefore, we propose that, due to increased coupling, tectonic erosion is favoured during contractional deformation periods, whereas the subsequent creation and subsidence of fore-arc basins occur only after the compressional strain has been released. The creation of the latest Cretaceous, Eocene and Miocene fore-arc basins is interpreted, therefore, as a delayed consequence of the tectonic erosion caused by the Peruvian, late Paleocene and late Oligocene contractional phases, respectively.

REFERENCES

Aceñolaza, F.G., Miller, H. and Toselli, A.J. (1990). *El Ciclo Pampeano en el noroeste argentino*. Univ. Nac. de Tucumán, Serie Correlación Geológica, vol. 4, 227 p.

Aguirre, L. and Offler, R. (1985). Burial metamorphism in the Western Peruvian Trough: its relation to Andean magmatism and tectonics. In: *Magmatism at a Plate Edge: the Peruvian Andes*, eds. Pitcher, M.P. et al., pp. 59-71, Blackie Halsted Press, London.

Alemán, A. (1980). Early Cretaceous arc sedimentation and volcanism in coastal ranges, central Peru. *American Association of Petroleum Geologists Bulletin*, **64**, 668 (Abstract).

Alemán, A. (1996). Stratigraphy, sedimentology and tectonic evolution of the Río Cañete Basin, central coastal range of Peru. In: *Andean Geodynamics II*, pp. 261-264, 3rd Int. Symp. And. Geodyn.-ISAG, Saint-Malo, Orstom publ., Paris.

Allmendinger, R.W. and Zapata, T.R. (1996). The Andean structure of the Cordillera Oriental from reprocessed YPF seismic reflection data. *3rd Int. Symp. And. Geodyn.-ISAG, Saint-Malo*, 265-268, Orstom publ., Paris.

Allmendinger, R.W., Ramos, V.A., Jordan, T.E., Palma, M. and Isacks, B.L., (1983). Paleogeography and Andean structural geometry, northwest Argentina. *Tectonics*, **2**, 1-16.

Allmendinger, R.W., Jordan, T.E., Kay, S.M. and Isacks, B.L (1997). The Evolution of the Altiplano-Puna plateau of the Central Andes. *Annual Review of Earth and Planetary Science*, **25**, 139-174.

Ángeles, C. (1987). *Les chevauchements de la Cordillère Occidentale par 12°15'S (Andes du Pérou central).* Thesis 3rd cycle, Univ. Montpellier II, 1 map. h.t., 184 p. (unpublished).

Ángeles, C. (1999). Los sedimentos cenozoicos de Cerro de Pasco: Estratigrafía, sedimentación y tectónica. *Soc. Geol. Perú*, Lima, Vol. jubilar **5**, 103-118.

Ardill, J., Flint, S., Chong, G. and Wilke, H. (1998). Sequence stratigraphy of the Domeyko Basin, northern Chile. *Journal of the Geological Society of London*, **155**, 71-88.

Aspden, J.A. and Litherland, M. (1992). The geology and Mesozoic collisional history of the Cordillera Real, Ecuador. *Tectonophysics*, **205**, 187-204.

Aspden, J.A. and McCourt, W.J. (1986). Mesozoic oceanic terrane in the Central Andes of Colombia. *Geology*, **14**, 415-418.

Aspden, J.A., Bonilla, W. and Duque, P. (1995). The El Oro metamorphic complex, Ecuador: geology and economic mineral deposits. *Overseas Geology and Mineral Resources*, 67, British Geological Survey publ., Nottingham, 63 p.

Aspden, J.A., Harrison, S.H. and Rundle, C.C. (1992). New geochronological control for the tectono-magmatic evolution of the metamorphic basement, Cordillera Real and El Oro Province of Ecuador. *Journal South American Earth Sciences*, **6**, 77-96.

Aspden, J.A., McCourt, W.J. and Brook, M. (1987). Geometrical control of subduction-related magmatism: the Mesozoic and Cenozoic plutonic history of Western Colombia. *Journal of the Geological Society of London*, **144**, 893-905.

Assumpção, M. and Araujo, M. (1993). Effect of the Altiplano-Puna plateau, South America, on the regional intraplate stresses. *Tectonophysics*, **221**, 475-493.

Assumpçao, M. (1992). The regional intraplate stress field in South America. *Journal of Geophysical Research*, **97**(B8), 11,889-11,903.

Astini, R.A., Benedetto, J.L. and Vaccari, N.E. (1995). The early Paleozoic evolution of the Argentina Precordillera as a Laurentian rifted, drifted, and collided terrane: a geodynamic model. *Geological Society of America Bulletin*, **107**(3), 253-273.

Atherton, M.P., Pitcher, W.S. and Warden, V. (1983). The mesozoic marginal basin of Central Peru. *Nature*, **305**, 303-305.

Atherton, M.P. and Petford, N. (1991). Rifting, insertial volcanism, batholith formation and crustal growth, Peru. *Terra Abstracts*, 3(1), 37-38.

Atherton, M.P. and Webb, S. (1989). Volcanic facies, structure and geochemistry of the marginal basin rocks of Central Peru. *Journal South American Earth Sciences*, **2**, 241-261.

Atherton, M.P., Warden, V. and Sanderson, L.M. (1985). The mesozoic marginal basin of Central Peru: A geochemichal study of within-plate-edge volcanism. In: M.P. Pitcher, E.J. Cobbing and R.D. Beckinsale (eds.). *Magmatism at a Plate Edge: the Peruvian Andes*, pp. 47-58. Blackie Halsted Press, London.

Atwater, T. (1989). Plate tectonic history of the northeast Pacific and western North America. In: *The Eastern Pacific Ocean and Hawaii*, eds. Winterer, E.L., Hussong, E.L. and Decker, R.W., pp. 21-72. The Geology of North America, vol. N, Geol. Soc. Am.

Audebaud, E. (1971). Mise au point sur la stratigraphie et la tectonique des calcaires cénomaniens du Sud-Est péruvien (formation Ayavacas). *Comptes Rendus de l'Académie des Sciences, Paris*, (D), **272**, 1059-1062.

Audebaud, E., Laubacher, G. and Marocco, R. (1976). Coupe géologique des Andes du Sud du Pérou, de l'Océan Pacifique au Bouclier Brésilien. *Geologische Rundschau*, **65**, 223-264.

Augusto, M., Ardiles, C. and Orosco, C. (1990). Geología del yacimiento Shiviyacu. *Boletin de la Sociedad Geologica del Peru*, **81**, 63-80.

Avila-Salinas, W.A. (1990). Tin-bearing granites from the Cordillera Real, Bolivia; a petrological and geochemical review. In: Plutonism from Antarctica to Alaska, eds. Kay, S. and Rapela, C. *Geological Society of America, Special Paper*, **241**, 145-159.

Azálgara, C., Jiménez, W., Morales, W. and Vergara, J. (1991). Posibilidades petrolíferas de las cuencas del Antearco, Perú. *Petroperú*, report, 37 p. (unpublished).

Baby, P., Rivadeneira, M., Christophoul, F. and Barragan, R. (1999). Style and timing of deformation in the Ecuadorian Oriente Basin. *4th Int. Symp. And. Geodyn.-ISAG*, Göttingen, IRD publ., Paris.

Baby, P., Rochat, P., Mascle, G. and Hérail, G. (1997). Neogene shortening contribution to crustal thickening in the back arc system of the Bolivian Orocline (Central Andes). *Geology*, **25**, 883-886.

Baby, P., Colletta, B. and Zubieta, D. (1995a). Étude géométrique et expérimentale d'un bassin transporté: Exemple du bassin subandin de l'Alto Beni (Andes centrales). *Bulletin de la Société Géologique de France*, **166**, 797-811.

Baby, P., Hérail, G., López, J.M., López, O., Oller, J., Pareja, J., Sempere, T. and Tufiño, D. (1989). Structure de la zone Subandine de Bolivie: influence de la géométrie des séries sédimentaires antéorogéniques sur la propagation des chevauchements. *Comptes Rendus de l'Académie des Sciences, Paris*, **309**(11), 1717-1722.

Baby, P. Hérail, G., Salinas, R. and Sempere, T. (1992). Geometry and kinematic evolution of passive roof duplexes deduced from cross section balancing: example from the foreland thrust system of the Southern Bolivian subandean orocline. *Tectonics*, **11**(3), 523-536.

Baby, P., Moretti, I., Guillier, B., Limachi, R., Méndez, E., Oller, J. and Specht, M. (1995b). Petroleum System of the northern and central Bolivian Sub-Andean Zone. In: Petroleum Basins of South America American, eds. Tankard, A.J., Suárez, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 445-458.

Baby, P., Specht, M., Oller, J., Montemuro, G., Colletta, B. and Letouzey, J. (1996). The Boomerang-Chapare transfer zone (recent oil discovery trend in Bolivia): Structural interpretation and experimental approach. In: *Geodynamic evolution of sedimentary basins*, eds. Roure, F., Shein, V.S. and Skvortsov, I. pp. 203-218. Technip publ.

Baeza, J. and Quinzio, L.A. (1991). Transgresión marina en el Hettangiano basal de posada los Hidalgos, Sur de Taltal, región de Antofagasta, Chile. *6 Cong. Geol. Chile, Viña del Mar, Serv. Nac. Geol. Min. Chile ed., Santiago*, 436-439.

Bahlburg, H. and Breitkreuz, C. (1991). Paleozoic evolution of active margin basins in the southern Central Andes (northwestern Argentina and northern Chile). *Journal of South American Earth Sciences*, **4**(3), 171-188.

Bahlburg, H. and Hervé, F. (1997). Geodynamic evolution and tectonostratigraphic terranes of northwestern Argentina and northern Chile. *Geological Society of America Bulletin*, **109**(7), 869-884.

Bahlburg, H., Moya, M.C. and Zeil, W. (1994). Geodynamic evolution of the Early Paleozoic continental margin of Gondwana in the southern Central Andes of northwestern Argentina and northern Chile. In: *Tectonics of the Southern Central Andes*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 293-302. Springer-Verlag.

Baldock, J.W. (1982). Geología del Ecuador. Boletín de Explicación del Mapa geológico de la República del Ecuador. *Dirección General de Geología y Minas, Quito*, 70 p.

Ballesteros, M.W., Moore, G.F., Taylor, B. and Ruppert, S. (1988). Seismic stratigraphic framework of the Lima and Yaquina forearc basins, Peru. *Proceedings of the Ocean Drilling Program: Initial Reports*, **112**, College Station, TX, 77-90.

Barazangi, M. and Isacks, BL. (1976). Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America. *Geology*, **4**, 686-692.

Bard, J.P., Botello, R., Martínez, C. and Subieta, T. (1974). Relations entre tectonique, métamorphisme et mise en place d'un granite éohercynien à deux micas dans la Cordillère Real de Bolivie (massif de Zongo-Yani). *Cahiers ORSTOM*, sér. Géol., 6(1), 3-18.

Barragán, R., Baudino, R. and Marocco, R. (1996). Geodynamic evolution of the Neogene intermontane Chota basin, Northern Andes of Ecuador. *Journal of South American Earth Sciences*, **9**, 309-319.

Batty, M. and Jaillard, E. (1989). La sedimentación neocomiana (Jurásico terminal - Aptiano) en el Sur del Perú. In: L.A. Spalletti (ed.). *Contribuciones de los simposios sobre el Cretácico de América latina*, Buenos-Aires, A75-A88.

Baudino, R. (1995). *Evolution des Andes d'Équateur au Néogène: les enseignements de l'étude des bassins montagneux*. Thèse Univ. Pau, Appendix, 437 p. (unpublished).

Baudino, R., Lavenu, A., and Delfaud, J. (1994). Un événement tectonique néogène majeur dans les Andes du Sud de l'Equateur. Déformation synsédimentaire dans le bassin de Nabón. *Comptes Rendus de l'Académie des Sciences de Paris*, **319**(II), 127-133.

Becerra, A., Orosco, C. and Augusto, M. (1990). Fallamiento inverso en el Nor-Oeste Peruano. *Boletin de la Sociedad Geologica del Peru*, **81**, 107-119.

Beck, S., Zandt, G., Myers, S., Wallace, T., Silver, P., Drake, L. and Minaya, E. (1996). Anomalous crust in the Central Andes. *3rd Int. Symp. And. Geodyn. ISAG, St Malo, France, 13-16*, Orstom publ., Paris.

Beckinsale, R.D., Sánchez-Fernandez, A.W., Brook, M., Cobbing, E.J., Taylor, W.P. and Moore, N.D. (1985). Rb-Sr whole-rock isochron and K-Ar age determinations for the Coastal Batholith of Peru. In: *Magmatism at a Plate Edge, the Peruvian Andes*, eds. Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D., pp. 177-202. Blackie, Glasgow and Halsted Press, New York.

Bell, C.M. (1987). The Late Paleozoic evolution of the Gondwanaland continental margin in northern Chile. In: Gondwana six: structure, tectonics and geophysics, ed. McKenzie, G.D. *AGU Geophysical Monograph*, **40**, 261-270.

Bellido, E. (1956). Geología del curso medio del rio Huaytara, Huancavelica. *Boletin de la Sociedad Geologica del Peru*, **13**, 49-75.

Ben-Avraham, Z. and Nur, A. (1987). Effects of collisions at trenches on ocanic ridges and passive margins. In: Circum-Pacific orogenic belts and evolution of the Pacific Ocean basin, eds. Monger, J.W.H. and Franchetau. *Journal of American Geophysical Union, Geodynamic Series*, **18**, 9-18.

Benavídes, V. (1956). Cretaceous system in Northern Peru. *American Museum of Natural History, Bulletin*, New York, **108**, 352-494.

Benavídes, V. (1962). Estratigrafía pre-Terciaria de la región de Arequipa. *Boletín de la Sociedad geológica del Perú*, **38**, 5-63.

Bengtson, P. and Jaillard, E. (1997). Stratigraphic revision of the Upper Cretaceous of the Peruvian-Ecuadorian border region: preliminary data and tectonic consequences. 18th I.A.S. Region. Europ. Meet. Sedimentology - 2nd Europ. Meet. Palaeont. Stratig. South America. *Gaea heidelbergensis, Heidelberg*, **4**, 71-72.

Benítez, S. (1995). Evolution géodynamique de la province côtière sud-équatorienne au Crétacé supérieur-Tertiaire. *Géologie Alpine, Grenoble*, **71**, 3-163.

Benítez, S., Zambrano, I., Ordoñez, M. and Chiriboga, J. (1993). Contribución al conocimiento estratigráfico del Paleógeno de la cuenca oriental ecuatoriana. *Actas Meeting PICG 301*, resumen, Caracas, Mars 93.

Benjamin, M.T., Johnson, N.M., and Naeser, C.W. (1987). Recent rapid uplift in the Bolivian Andes: evidence from fission-track dating. *Geology*, **15**, 680-683.

Berg, K. and Baumann, A. (1985). Plutonic and metasedimentary rocks from the Coastal Range of northern Chile: Rb-Sr and U-Pb isotopic systematics. *Earth and Planetary Science Letters*, **75**, 101-115.

Bernoulli, D. and Lemoine, M. (1980). Birth and evolution of the Tethys: the overall situation. *26th Intern. Geol. Cong. Paris*, Col. C5, 167-179.

Berrones, G. and Cotrina, J. (1996). Estudios de subsidencia y decompactación en el Noreste de la cuenca oriental; implicaciones tectónicas. *Actas VII Cong. Ecuat. Geol. Min. Petrol., Quito*, 248-258.

Bijwaard H., Spakman, W. and Engdahl, E.R. (1998). Closing the gap between regional and global travel time tomography. *Journal of Geophysical Research*, **103**, 30055-30078.

Bogdanic, T. and Espinosa, S. (1994). Tectono-sedimentary evolution of the Cretaceous-early Tertiary and metallogenic scheme of northern Chile, between 20°S and 26°S. In: *Cretaceous tectonics in the Andes*, ed. Salfity, J.A., pp. 213-265. Earth Evolution Sciences, Fried. Vieweg and Sohn, Braunschweig/Wiesbaden.

Boily, M., Brooks, C. and James, D.E. (1984). Geochemical characteristics of the late mesozoic Andean volcanics. In: *Andean magmatism: Chemical and isotopic constraints*, eds. Harmon, R.S. and Barreiro, B.A., pp. 190-202. Shiva publ.

Boily, M., Ludden, J.N. and Brooks, C. (1990). Geochemical constraints on the magmatic evolution of the pre- and post-Oligocene volcanic suites of southern Peru: Implications for the tectonic evolution of the Central Volcanic Zone. *Geological Society of America Bulletin*, **102**, 1565-1579.

Bond, G.C., Nickeson, P.A. and Kominz, M.A. (1984). Breakup of a supercontinent between 625 Ma and 555 Ma: new evidence and implications for continental histories. *Earth and Planetary Science Letters*, **70**, 325-345.

Bonhomme, M. and Carlier, G. (1990). Relation entre magmatisme et minéralisation dans le batholite d'Andahuaylas-Yauri (Sud Pérou): données chronologiques. *1st International Symposium on Andean Geodynamics-ISAG*, Grenoble, Orstom publ., Paris, 329-331.

Bonhomme, M., Audebaud, E. and Vivier, G. (1985). Edades K/Ar de rocas hercínicas y neógenas de un perfil E-W en el Perú meridional. *Communicaciones, Univ. Chile, Dept. Geología*, **35**, 27-30.

Bott, M.H.P., Waghorn, G.D. and Whittaker, A. (1989). Plate boundary forces at subduction zones and trench-arc compression. *Tectonophysics*, **170**, 1-15.

Bourgois, J., Egüez, A., Butterlin, J. and De Wever, P. (1990). Evolution géodynamique de la Cordillère Occidentale des Andes d'Equateur: la découverte de la formation éocène d'Apagua. *Comptes Rendus de l'Académie des Sciences, Paris*, **311**(II), 173-180.

Breitkreuz, C. and Zeil, W. (1994). The Late Carboniferous to Triassic volcanic belt in northern Chile. In: *Tectonics of the Southern Central Andes*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 277-292. Springer-Verlag.

Breitkreuz, C. (1990). Late Carboniferous to Triassic magmatism in the Central and S. Andes: the change from an accretionary to an erosive plate margin mirrors the Pangea history. *I International Symposium on Andean Geodynamics, Grenoble, Résumés des Communications*, 359-362.

Breitkreuz, C., Bahlburg, H. and Zeil, W. (1988). The Paleozoic evolution of northern Chile: geotectonic implications. In: The southern Central Andes, eds. Bahlburg, H.; Breitkreuz, C. and Giese, P. *Lecture Notes in Earth Sciences*, **17**, 87-102

Breitkreuz, C., Bahlburg, H., Delakowitz, B. and Pichowiak, S. (1989). Paleozoic volcanic events in the Central Andes. *Journal of South American Earth Sciences*, 2, 171-189.

Bristow, C.R., and Hoffstetter, R. (1977). *Ecuador*. Lexique Stratigraphique International, V, 5a2, CNRS publ., Paris, 410 p.

Brown, M., Dallmeyer, R.D. and Grocott, J. (1993). Tectonic control on Mesozoic arc magmatism in North Chile. *2nd International Symposium on Andean Geodynamics-ISAG, Oxford, Ext. Abst.*, Orstom ed., Paris, 167-170.

Bussel, M.A. and Pitcher, W.S. (1985). The structural control of batholith emplacement. In: *Magmatism at a Plate Edge: the Peruvian Andes*, eds. Pitcher, M.P., Cobbing, E.J. and Beckinsale, R.D., pp. 167-176. Blackie Halsted Press, London.

Bussel, M.A. (1983). Timing of tectonic and magmatic events in the Central Andes of Peru. *Journal of the Geological Society, London*, **140**, 279-286.

Cahill, T. and Isacks, B. L. (1992). Seismicity and shape of the subducted Nazsca plate. *Journal of Geophysical Research*, **97**, 17,503-17,529.

Carlier, G., Grandin, G., Laubacher, G., Marocco, R. and Mégard, F. (1982). Present knowledge of the magmatic evolution of the Eastern Cordillera of Peru. *Earth Science Reviews*, **18**, 253-283.

Carlier, G., Lorand, J.-P., Bonhomme, M. and Carlotto, V. (1996). A reappraisal of the Cenozoic inner arc magmatism in Southern Peru: consequences for the evolution of the Central Andes for the past 50 Ma. *3rd Int. Symp. And. Geodyn.*-ISAG, Saint-Malo, Orstom publ., Paris, 551-554.

Carlotto, V. (1998). *Évolution andine et raccourcissement au niveau de Cusco (13-16°S), Pérou. Enregistrement sédimentaire, chronologie, contrôles paléogéographiques, évolution cinématique*. Thèse Université Grenoble I, 250 p.

Carlotto, V., Cardenas, J., Gil, W. and Chavez, R. (1995). Sedimentología de la Formación Huancané (Neocomiano) de la región de Cusco y su relación con las variaciones del nivel del mar. *Bull. Instit Français Études Andines*, **24**, 1-21.

Cembrano, J., Hervé, F. and Lavenu, A. (1996). The Liquiñe-Ofqui fault zone, southern Chile: a case of a long-lived intra-arc (strike-slip) fault system? *Tectonophysics*, **259**, 55-66.

Clark, A.H., Farrar, E., Kontak, D.J., Langridge, R.J., Arenas, M.J., France, L.J., McBride, S.L., Woodman, P.L., Wasteneys, H.A., Sandeman, H.A. and Douglas, D.A. (1990). Geologic and Geochronologic Constraints on the Metallogenic Evolution of the Andes of Southeastern Peru. *Economic Geology*, **85**, 1520-1583.

Cloos, M. and Shreve, R.L. (1996). Shear-zone thickness and the seismicity of Chilean- and Marianas-type subduction zones. *Geology*, **24**, 107-110.

Cloos, M. (1993). Lithospheric buoyancy and collisional orogenesis: Subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts. *Geological Society of America Bulletin*, **105**, 715-737.

Cobbing, E.J., Pitcher, W.S., Wilson, J.J., Baldock, J.W., Taylor, W.P., McCourt, W. and Snelling, N.J. (1981). The geology of the Western Cordillera of Northern Peru. Institute of Geological Sciences London, *Overseas Memoir*, **5**, 143 p.

Coira, B., Davidson, J., Mpodozis, C. and Ramos, V. (1982). Tectonic and magmatic evolution of the Andes of Northern Argentina and Chile. *Earth Science Review*, **18**, 303-332.

Comte, D., Dorbath, L., Ardo, M., Monfret, T., Haessler, H., Rivera, L., Frogneux, M., Glass, B. and Meneses, C. (1999). A double-layered seismic zone in Arica, northern Chile. *Geophysical Research Letters*, **26**, 1965-1968.

Comte D., Roecker S. and Suárez G. (1994). Velocity structure in northern Chile: evidence of subducted oceanic crust in the Nazca plate. *Geophysical Journal International*, **117**, 625-639.

Contreras, C., Jaillard, É. and Paz, M. (1996). Subsidence history of the north peruvian Oriente (Marañon basin) since the Cretaceous. *3rd International Symposium on Andean Geodynamics-ISAG*, Saint-Malo, Orstom ed., Paris, 327-330.

Córdova, E. (1986). *Un bassin intramontagneux andin péruvien. Les Couches Rouges du bassin de Cuzco (Maastrichtien-Paléocène)*. Thèse 3° cycle Université Pau, 272 p.

Cosma, L., Lapierre, H., Jaillard, É., Laubacher, G., Bosch, D., Desmet, A., Mamberti, M. and Gabriele, P. (1998). Pétrographie et géochimie de la Cordillère Occidentale du Nord de l'Équateur (0°30'S): Implications tectoniques. *Bulletin de la Société géologique de France*, Paris **169**, 739-751.

Couch, R. and Whitsett, R.M. (1981). Structures of the Nazca Ridge and continental shelf and slope of southern Peru. *Geological Society of America Memoir*, **154**, 569-586.

Courjault-Radé, P., Debrenne, F. and Gandin, A. (1992). Palaeogeographic and geodynamic evolution of the Gondwana continental margins during the Cambrian. *Terra Nova*, **4**, 657-667.

Cross, T.A. and Pilger, R.H. (1982). Controls of subduction, location of magmatic arcs, and tectonics of arc and back-arc regions. *Geological Society American Bulletin*, **93**, 545-562.

Chanove, G., Mattauer, M. and Mégard, F. (1969). Précisions sur la tectonique tangentielle des terrains secondaires du massif de Pirin (Nord-Ouest du lac Titicaca, Pérou). *Comptes Rendus de l'Académie des Sciences, Paris* (D), **268**, 1698-1701.

Charrier, R. and Muñoz, N. (1994). Jurassic-Cretaceous palaeogeographic evolution of the Chilean Andes at 23°-24°S Latitude and 34°-35°S Latitude: A comparative analysis. In: *Tectonics of the Southern Central Andes*, pp. 233-242. Springer-Verlag.

Charrier, R. and Reutter, K.-J. (1994). The Purilactis Group of northern Chile: Boundary between arc and backarc from Late Cretaceous to Eocene. In: *Tectonics of the southern Central Andes*, eds. Reutter, K.-J., Scheuber, E. and Wigger, P.J., pp. 189-202, Springer Verlag, Berlin.

Charrier, R., Hérail, G., Flynn, J., Riquelme, R., García, M., Croft, D. and Wyss, A. (1999). Opposite thrust-vergencies in the Pre-Cordillera and Western Cordillera of Northern Chile, and structurally linked Cenozoic paleo-environmental evolution. *4th International Symposium on Andean Geodynamics-ISAG*, Göttingen, IRD publ., Paris, 155-158.

Chávez, A. (1982). *Estratigrafía y tectónica del sector de Chapi*. Thesis Universidad Nacional San Agustin, Arequipa, 101 p. (unpublished).

Cheilletz, A., Giuliani, G., Branquet, Y., Laumonier, B., Sánchez, A.J., Féraud, G. and Arhan, T. (1997). Datation K-Ar et $^{40}Ar/^{39}Ar$ à 65 ± 3 Ma des gisements d'émeraude du district de Chivor-Macanal: argument en faveur d'une déformation précoce dans la Cordillère orientale de Colombie. *Comptes Rendus de l'Académie des Sciences, Paris*, **324**(IIa), 369-377.

Chong, G. (1976). El sistema Jurásico en la Cordillera de Domeyko (Chile) entre 24°30' y 25°30' de Latitud Sur. *Memoria 2do Cong. Latinoamer. Geol.*, Caracas **1973**(II), 765-785.

Christophoul, F. (1999). *Differentiation des influences tectoniques et eustatiques dans les bassins liés à des zones de convergence: exemple du bassin subandin d'Equateur*. Thèse de Doctorat Univ. Paul Sabatier, 184 p. + annexes.

Christophoul, F., Baby, P. and Dávila, C. (1999). Discrimination of eustatic and tectonic influences in the Ecuadorian Oriente Basin from Aptian to Oligocene times. *4th International Symposium on Andean Geodynamics-ISAG*, Göttingen, IRD Publ., Paris, 168-171.

Dalmayrac, B., Laubacher, G., Marocco, R., Martínez, C. and Tomasi, P. (1980). La chaine hercynienne d'Amerique du sud: Structure et evolution d'un orogene intracratonique. *Geologische Rundschau*, **69**(1), 1-21.

Dalziel, I.W.D. (1994). Precambrian Scotland as a Laurentia-Gondwana link: Origin and significance of cratonic promontories. *Geology*, **22**, 589-592.

Damm, K.W., Harmon, R.S. and Kellye, S. (1994). Some isotopic and geochemical constraints on the origin and evolution of the Central Andean basement (19-24°S). In: *Tectonics of the Southern Central Andes*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 263-276. Springer-Verlag.

Damm, K.W., Pichowiak, S., Harmon, R.S., Todt, W., Breitkreuz, C. and Buchelt, M. (1991). The Cordón de Lila Complex, Central Andes, N-Chile: an Ordovician continental volcanic province. *Geological Society of America, Special Paper*, **265**, 179-188.

Damm, K.W., Pichowiak, S., Harmon, R.S., Todt, W., Omarini, R. and Niemeyer, H. (1990). Pre-Mesozoic evolution of the Central Andes - the basement revisited. *Paper*, **241**, 101-126.

Dashwood, M.F. and Abbots, I.L. (1990). Aspects of the petroleum geology of the Oriente Basin, Ecuador. In: Classic Petroleum Provinces, ed. Brooks, J. *Geological Society of America, Special Publ.*, **50**, 89-117.

De Vries, T.J. (1998). Oligocene deposition and Cenozoic sequence boundaries in the Pisco Basin (Peru). *Journal of South American Earth Sciences*, **11**, 217-231.

Delfaud, J., Sabrier, R., Baudino, R., Lavenu, A. and Marocco, R. (1999). Reconstitution des étapes de la surrection des Andes d'Équateur à partir de l'interprétation des minéraux argileux contenus dans les bassins intramontagneux (Miocène à Actuel). *Bulletin de la Société Géologique de* France, **170**, 13-23.

Deniaud, Y., Baby P., Basile C., Ordoñez M., Montenegro G. and Mascle G. (1999a). Neogene evolution of the main Ecuadorian fore-arc sedimentary basins and sediment mass-balance inferences. *4th Int. Symp. And. Geodyn.*, Göttingen, Germany, IRD Publ., Paris.

Deniaud, Y., Baby, P., Basile, C., Ordoñez, M., Montenegro, G. and Mascle, G. (1999). Ouverture et évolution tectono-sédimentaire du golfe de Guayaquil: bassin d'avant-arc néogène et quaternaire du Sud des Andes équatoriennes. *Comptes Rendus de l'Académie des Sciences*, Paris, **328**, 181-187.

Díaz Martínez, E., Vavrdová, M., Bek, J. and Isaacson, P.E. (1999). Late Devonian (Famennian) glaciation in western Gondwana: evidence from the Central Andes. *Jahrbuch der Geologisches Bundesanstalt*, **54** (in press).

Díaz-Martínez, E. (1995). Evidencia de actividad volcánica en el registro sedimentario del Carbonífero inferior (Viseano superior) del Altiplano norte de Bolivia (16°S), y su relación con el arco magmático de los Andes Centrales. *Revista Técnica de YPFB*, **16**(1/2), 37-49.

Díaz-Martínez, E. (1996). Síntesis estratigráfica y geodinámica del Carbonífero de Bolivia. *XII Congreso Geológico de Bolivia*, Tarija. Memorias, **1**, 355-367.

Díaz-Martínez, E. (1998a). Silurian of Peru and Bolivia: recent advances and future research. *6th International Graptolite Conference and Field Meeting of the International Subcommission on Silurian Stratigraphy*, Madrid. Temas Geológico-Mineros ITGE, **23**, 69-75.

Díaz-Martínez, E. (1998b). Estratigrafía y paleogeografía del Paleozoico superior del norte de los Andes Centrales (Perú y Bolivia). *Boletín de la Sociedad Geológica del Perú*, **89** (in press).

Díaz-Martínez, E., Isaacson, P.E. and Sablock, P.E. (1993). Late Paleozoic latitudinal shift of Gondwana: stratigraphic/sedimentologic and biogeographic evidence from Bolivia. *Documents des Laboratoires de Géologie, Lyon*, **125**, 119-138.

Díaz-Martínez, E., Limachi, R., Goitia, V.H., Sarmiento, D., Arispe, O. and Montecinos, R. (1996). Tectonic instability related with the development of the Paleozoic foreland basin of the Central Andes of Bolivia. *3rd International Symposium on Andean Geodynamics, St.-Malo. Expanded abstracts*, 343-346.

Dorbath C, A. Paul and Lithoscope Andean Group (1996). Tomography of the Andean crust and mantle at 20°S: first results of the Lithoscope experiment, *Physics of the Earth and Planetary Interiors*, **97**, 133-144.

Dorbath, C., Granet, M., Poupinet, G. and Martinez, C. (1993). Teleseismic study of the Altiplano and the Eastern Cordillera in northern Bolivia: New constraints on a lithospheric model. *Journal of Geophysical Research*, **98**, 9,825-9,844.

Dumont, J.F. and Fournier, M. (1994). Geodynamic environment of Quaternary morphostructures of the Subandean foreland basins of Peru and Bolivia: Characteristics and study methods. *Quaternary International*, **21**, 129-142.

Dumont, J.F. (1993). Lake patterns as related to neotectonics in subsiding basins: the example of the Ucamara Depression, Peru. *Tectonophysics*, **222**, 69-78.

Dumont, J.F. (1996). Neotectonic of Subandes/Brazilian craton boundaries using geomorphological data: Marañon and Beni Basins. *Tectonophysics*, **257**, 137-151.

Dumont, J.F., Alvarado, A., Guillier, B., Lavenu, A., Martinez, C., Ortlieb, L., Poli, J.T., Labrousse, B. (1997). Coastal morphology as related to geodynamics in Western Ecuador: Preliminary results. *Abs. 16, Late Quaternary Coastal Tectonics*, London.

Duncan, R.A. and Hargraves, R.B. (1984). Plate Tectonic evolution of the Caribbean region in the mantle reference frame. *Geological Society of America, Memoir*, **162**, 81-93.

Dunkley, P.N. and Gaibor, A. (1998). *Mapa geológico de la Cordillera Occidental del Ecuador entre 2° - 3° S., escala 1/200.000.* CODIGEM-Min. Energ. Min.-BGS publs., Quito.

Dunn, J.F., Hartshorn, K.G. and Hartshorn, P.W. (1995). Structural styles and hydrocarbon potential of the Subandean belt of southern Bolivia. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez S., R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 523-543.

Eduardo, H. (1991). Paleogeografía del Paleozoico en el oriente peruano. *VII Congreso Peruano de Geología*, Lima, volumen de resúmenes extendidos, 269-275.

Ego, F. (1995). *Accomodation de la convergence oblique dans une chaîne de type cordillerain: les Andes d'Equateur*. Thesis, Univ. Paris-Sud - Orsay, 209 p. (unpublished).

Ego, F., Sébrier, M., Lavenu, A., Yepes, H. and Egüez, A (1996). Quaternary state of stress in northern Andes and the restraining bend model for the Ecuadorian Andes. *Tectonophysics*, **259**, 101-116.

Egüez, A. (1986). *Evolution Cénozoïque de la Cordillère Occidentale Septentrionale d'Equateur (0°15'S-1°10'S): les minéralisations associées*. Unpublished thesis University Paris VI, Paris, 116 p.

Egüez, A., Dugas, F. and Bonhomme, M. (1992). Las unidades Huigra y Alausi en la evolución geodinámica del valle interandino del Ecuador. *Boletín Geológico Ecuatoriano*, **3**, 47-55, Quito.

Emery, K.O. and Uchupi, E. (1984). *The geology of the Atlantic Ocean*. Springer Verlag, New York, 1050 p.

Engdahl, E.R., van der Hilts, R.D. and Buland, R. (1998). Global teleseismic earthquake relocation with improved travel times and produced for depth relocation. *Bullettin of Seismological Society of America*, **88**, 722-743.

Evans, C.D.R. and Whittaker, J.E. (1982). The geology of the western part of the Borbón Basin, North-west Ecuador. In: *Trench-Forearc Geology*, ed. Legget, J.K., pp. 191-198. Geol. Soc. London Spec. Publ., 10, Blackwell Scient. Publ., Oxford.

Farrar, E., Clark, A. and Heinrich, S. (1990). The age of the Zongo pluton and the tectonothermal evolution of the Zongo-San Gabán zone in the Cordillera Real, Bolivia. *Ist International Symposium on Andean Geodynamics-ISAG, Grenoble*, ORSTOM

Publ., Paris, 171-174.

Farrar, E., Clark, A.H., Kontak, D.J. and Archibald, D.A. (1988). Zongo-San Gabán zone : Eocene foreland boundary of the central Andean orogen, northwest Bolivia and southeast Peru. *Geology*, **16**, 55-58.

Faucher, B., Vernet, R., Bizon, G., Bizon, J.J., Grekoff, N., Lys, M. and Sigal, J. (1971). *Sedimentary Formations in Ecuador. A stratigraphic and micropaleontological survey*. Bureau Études Indust. Coop. Inst. Franç. Pétrole (BEICIP), 3 vol., 220 p.

Feininger, T. and Silberman, M.L. (1982). K-Ar geochronology of basement rocks on the Northern Flank of the Huancabamba deflection, Ecuador. *U.S. Geological Survey*, Open-file Report, 21 p., Menlo Park, 82-206.

Fisher, A.G. (1956). Desarollo geológico del Noroeste peruano durante el Mesozóico. *Boletín de la Sociedad geológica del Perú*, **30**, 177-190.

Flint, S., Turner, P., Jolley, E.J. and Hartley, A.J. (1993). Extensional tectonics in convergent margin basins: An example from the Salar de Atacama, Chilean Andes. *Geological Society of America Bulletin*, **105**, 603-617.

Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., eds. (1990). *Stratabound Ore Deposits in the Andes*. Springer Verlag, 815 p.

Fornari, M. and Hérail, G. (1991). Lower Paleozoic gold occurrences in the Eastern Cordillera of southern Peru and northern Bolivia: a genetic model. In: *Brazil Gold '91* (the economics, geology, geochemistry and genesis of gold deposits), ed. *Ladeira, E.A., pp. 135-142.*

Fornari, M., Hérail, G., Viscarra, G., Laubacher, G. and Argollo, J. (1987). Sédimentation et structure du bassin Tipuani-Mapiri: Un témoin de l'évolution du front amazonien des Andes du nord de la Bolivie. *Comptes Rendus à l'Académie des Sciences, Paris*, **304**, 1303-1309.

Forsythe, R.D., Davidson, J., Mpodozis, C. and Jesinkey, C. (1993). Lower Paleozoic relative motion of the Arequipa block and Gondwana: paleomagnetic evidence from Sierra de Almeida of northern Chile. *Tectonics*, **12**, 219-236.

Frutos, J. (1981). Andean tectonics as a consequence of sea-floor spreading. *Tectonophysics*, **72**, T21-T32.

Frutos, J., 1990. The Andes Cordillera: a synthesis of the geologic evolution. In: *Stratabound Ore Deposits in the Ande*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 3-35. Springer Verlag.

Gabriele, P., Ballèvre, M., Jaillard, É. and Henandez, J. (1999). Decompression at decreasing temperatures in eclogite-facies metapelites (El Oro Metatmorphic Complex, SW Ecuador): A record of fast exhumation rates. *4th International Symposium on Andean Geodynamics-ISAG*, Göttingen, IRD publ., Paris, 245-248.

Gana, P. (1991). Magmatismo bimodal del Triasico superior-Jurásico inferior en la Cordillera de la Costa, Provincias de Elqui y Limari, Chile. *Revista Geológica de Chile*, **18**, 55-67.

Garcia, M., Hérail, G. and Charrier, R. (1996). The Cenozoic forearc evolution in northern Chili: the western border of the Altiplano of Belén (Chile). *3rd Int. Symp. And. Geodyn.-ISAG*, St Malo, France, Orstom publ., Paris, 359-362.

Garcia, M., Hérail, G. and Gardeweg, M. (1999). Oligo-Miocene ignimbritic volcanism of northern Chile (Arica region): stratigraphy and geochronology. 4th Int. Symp. And. Geodyn.-ISAG, Göttingen, IRD publ., Paris, 253-256.

García, M. (1997). Le volcanisme calco-alcalin oligo-miocène des Andes centrales (Nord du Chili): implications sur l'orogenèse andine. *Mémoire de DEA, Université Grenoble I*, 30 p. + Annexes.

Gayet, M., Sempere, T., Cappetta, H., Jaillard, É. and Lévy, A. (1993). La présence de fossiles marins dans le Crétacé terminal des Andes centrales et ses conséquences paléogéographiques. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **102**, 283-319.

Geyer, O. (1979). Ammoniten aus dem tiefen Unterjura von Nord-Peru. *Paläontologische Zeitshrift, Stuttgart*, **53**, 198-213.

Geyer, O.F. (1982). Comparaciones estratigráficas y faciales en el Triásico norandino. *Geología Norandina, Bogota*, 5, 27-31.

Giese, P., Scheuber, E., Schilling, F., Schmitz, M. and Wigger, P. (1999). Crustal thickening processes in the Central Andes and the different natures of the Moho discontinuity. *Journal of South American Earth Sciences*, **12**, 201-220.

Gil, W., Baby, P. and Paz, M. (1996). Continum tectonic during Cretaceous-Paleocene times in the north-peruvian foreland Basin (Marañón Basin). *3rd Intern. Symp. And. Geodyn.-ISAG*, Saint-Malo, Orstom publ., Paris, 363-366.

Gohrbandt, K.H.A., 1992. Paleozoic paleogeographic and depositional developments on the central proto-Pacific margin of Gondwana: their importance to hydrocarbon accumulation. *Journal of South American Earth Sciences*, **6**(4), 267-287.

González, G. (1976). *Ciclos de sedimentación en el Eoceno de la* Cuenca de Talara. *Bol. Soc. geol. Perú*, **51**, 73-80.

González, M., Díaz-Martínez, E. and Ticlla, L. (1996). Comentarios sobre la estratigrafía del Silúrico y Devónico del norte y centro de la Cordillera Oriental y Altiplano de Bolivia. *Simpósio Sul Americano do Siluro-Devoniano, Ponta Grossa. Anais*, 117-130.

González-Bonorino, G. (1991). Late Paleozoic orogeny in the northwestern Gondwana continental margin, western Argentina and Chile. *Journal of South American Earth Sciences*, 4, 131-144.

Gordon, R.G. and Jurdy, D.M. (1986). Cenozoic global plate motions. *Journal of Geophysical Research*, **91**, 12389-12406.

Gregory, H.E. (1916). Geological reconnaissance of the Cuzco valley, Peru. *American Journal Science*, fourth serie, XLI, **241**, 1-121.

Gröschke, M., Hillebrandt, A. von, Prinz, P., Quinzio, L.A. and Wilke, H.-G. (1988). Marine Mesozoic paleogeography in Northern Chile between 21°-26°S. In: *The Southern Central Andes*, eds. Bahlburg, H. et al., pp. 105-117. Lecture Notes in Earth Sciences, Springer Verlag.

Gubbels, T.L., Isacks, B.L. and Farrar, E. (1993). High-level surfaces, plateau uplift, and foreland development, Bolivian central Andes. *Geology*, **21**, 695-698.

Gurnis, M. (1992). Rapid continental subsidence following the initiation and evolution of subduction. *Science*, **255**, 1556-1558.

Gutierrez, M. (1982). Zonación bioestratigráfica del intervalo Cretáceo superior - Terciario inferior. In: *Evaluación del potencial petrolífero de las Cuencas Huallaga, Ucayali y Madre de Dios*. Petroperu, Lima, report, 30 p. (unpublished).

Gutiérrez-Marco, J.C., Saavedra, J. and Rábano, I., eds. (1992). *Paleozoico Inferior de Iberoamérica*. Universidad de Extremadura, 630 p.

Gutscher M.A., W. Spakman and H. Bijwaard (1999). Flat subduction beneath the Andes: seismological and tomographic constraints, *Fourth International Symposium on Andean Geodynamics*, Göttingen, 311-314.

Hall, M.L and Calle, J. (1982). Geochronological Control for the main tectonic-magmatic events of Ecuador. *Earth Science Review*, **18**, 215-239.

Hammerschmidt, K., Döbel, R. and Friedrichsen, H. (1992). Implication of $^{40}Ar/^{39}Ar$ dating of Early Tertiary volcanic rocks from the north-Chilean Precordillera. *Tectonophysics, Amsterdam*, **202**, 55-81.

Hartenberger, J.-L., Mégard, F. and Sigé, B. (1984). Faunules à rongeurs de l'Oligocène inférieur à Lircay (Andes du Pérou Central): datation d'un épisode karstique; intérêt paléobiogéographique des remplissages tertiaires en Amérique du Sud. *Comptes Rendus de l'Académie des Sciences,* *Paris*, **299**, (II), 565-568.

Hartley, A.J., Flint, S., Turner, P. and Jolley, E.J. (1992). Tectonic controls on the development of a semi-arid, alluvial basin as reflected in the stratigraphy of the Purilactis Group (Upper Cretaceous-Eocene), northern Chile. *Journal of South American Earth Sciences*, 5, 275-296.

Heinrich, S.M., Farrar, E., Clark, A.H., Archibald, D.A. and Parrish, R.R. (1988). Age, uplift and thermal evolution of the Zongo pluton, Cordillera Oriental, Bolivia. *Eos (Abstracts)*, **698**, 16, 487.

Hérail, G., Oller, J., Baby, P., Bonhomme, M. and Soler, P. (1996). The Tupiza, Nazareno and Estarca basins (Bolivia): Strike-slip faulting and related basins in the Cenozoic evolution of the southern branch of the Bolivian Orocline. *Tectonophysics*, **259**, 201-212.

Hérail, G., Soler, P., Bonhomme, M.G. and Lizeca, J.L. (1993). Evolution géodynamique du contact Altiplano-Cordillère Orientale au Nord d'Oruro (Bolivie); implications sur le déroulement de l'orogénèse andine. *Comptes Rendus de l'Académie des Sciences, Paris*, **317**, 512-522.

Hoorn, C. (1993). Marine incursions and the influence of Andean tectonics on the Miocene depositional history of northwestern Amazonia: Results of a palynological study. *Palaeogeography, Palaeoclimatology, Palaeoecology*, Amsterdam, **105**, 267-309.

Hoorn, C., Guerrero, J., Sarmiento, G.A. and Lorente, M.-A. (1995). Andean tectonics as a cause for changing drainage patterns in Miocene northern South America. *Geology*, **23**, 237-240.

Hughes, R.A. and Pilatasig, L.F. (1999). Cretaceous and Tertiary terrane accretion in the Cordillera Occidental of the Ecuadorian Andes. *4th International Symposium on Andean Geodynamics-ISAG*, Göttingen, IRD Publ., Paris, 340-342.

Hungerbühler, D. (1997). *Neogene basins in the Andes of southern Ecuador: evolution, deformation and regional tectonic implications.* PhD thesis, ETH Zürich, n° 12371, 1 map h.t., 182 p.

Hungerbühler, D., Steinmann, M. Winkler, W., Seward, D., Egüez, A., Heller, F. and Ford, M. (1995). An integrated study of fill and deformation in the Andean intermontane basin of Nabón (Late Miocene), southern Ecuador. *Sedimentary Geology*, **96**, 257-279.

Iddings, A. and Olsson, A.A. (1928). Geology of Northwest Peru. *Bulletin of the American Association of Petroleum Geologists*, **12**, 1-39.

Irwin, J.J., Garcia, C., Hervé, F. and Brook, M. (1987). Geology of part of a long-lived dynamic plate margin: the coastal cordillera of north-central Chile, latitude 30°51-31°S. *Canadian Journal of Earth Sciences*, **25**, 603-624.

Isaacson, P.E. and Díaz-Martínez, E. (1995). Evidence for a Middle-Late Paleozoic foreland basin and significant paleolatitudinal shift, Central Andes. *AAPG Memoir*, **62**, 231-249.

Isacks, B.L. (1988). Uplift of the central Andean plateau and bending of the Bolivian orocline. *Journal of Geophysical Research*, **93**, 3211-3231.

Jacay, J. (1994). Evolution sédimentaire de la marge andine: le Crétacé supérieur des Andes du Pérou central et septentrional. *Mémoire de DEA*, Université de Grenoble I, 97 p.

Jaillard, E. and Jacay, J. (1989). Les "couches Chicama" du Nord du Pérou: Colmatage d'un bassin né d'une collision oblique au Tithonique. *Comptes Rendus de l'Académie des Sciences, Paris*, **308**(II), 1459-1465.

Jaillard, E. and Santander, G. (1992). La tectónica polifásica en escamas de la zona de Mañazo-Lagunillas (Puno, Sur del Perú). *Bull. Instit Français Études Andines*, Lima, **21**, 37-58.

Jaillard, E. and Soler, P. (1996). The Cretaceous to Early Paleogene tectonic evolution of the northern Central Andes and its relations to geodynamics. *Tectonophysics*, **259**, 41-53.

Jaillard, E. (1987). Sedimentary evolution of an active margin during middle and upper Cretaceous times: the North Peruvian margin from Late Aptian up to Senonian. *Geologische Rundschau*, **76**, 677-697.

Jaillard, E. (1994). Kimmeridgian to Paleocene tectonic and geodynamic evolution of the Peruvian (and Ecuadorian) margin. In: J.A. Salfity (ed.). *Cretaceous tectonics in the Andes*, pp. 101-167. Earth Evolution Sciences, Fried. Vieweg and Sohn, Braunschweig/Wiesbaden.

Jaillard, E. (1995). La sedimentación Albiana - Turoniana en el Sur del Perú (Arequipa-Puno-Putina). *Sociedade Geológica del Perú*, Vol. Jubilar Alberto Benavides, Lima, 135-157.

Jaillard, É. (1997). *Síntesis estratigráfica y sedimentológica del Cretáceo y Paleógeno de la cuenca oriental del Ecuador.* Orstom-Petroproducción publ., Quito, 164 p.

Jaillard, E., Cappetta, H., Ellenberger, P., Feist, M., Grambast-Fessard, N., Lefranc, J. P. and Sigé, B. (1993). The Late Cretaceous Vilquechico Group of Southern Peru. Sedimentology, Paleontology, Biostratigraphy, Correlations. *Cretaceous Research*, London, **14**, 623-661.

Jaillard, E., Ordoñez, M., Benítez, S., Berrones, G., Jiménez, N., Montenegro, G. and Zambrano, I. (1995). Basin development in an accretionary, oceanic-floored forearc setting: southern coastal Ecuador during late Cretaceous to late Eocene times. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 615-631.

Jaillard, E., Soler, P., Carlier, G. and Mourier, T. (1990). Geodynamic evolution of the northern and central Andes during early to middle Mesozoic times: a Tethyan model. *Journal of the Geological Society of London*, **147**, 1009-1022.

Jaillard, É., Laubacher, G., Bengtson, P., Dhondt, A. and Bulot, L. (1999). Stratigraphy and evolution of the Cretaceous forearc "Celica-Lancones Basin" of Southwestern Ecuador. *Journal of South American Earth Sciences*, 12, 51-68.

Jaillard, É., Ordoñez, M., Bengtson, P., Berrones, G., Bonhomme, M., Jiménez, N. and Zambrano, I. (1996). Sedimentary and tectonic evolution of the arc zone of southwestern Ecuador during Late Cretaceous and Early Tertiary times. *Journal of South American Earth Sciences*, **9**, 131-140.

James D.E. 1971, Andean crustal and upper mantle structure. *Journal of Geophysical Research*, **76**, 3246-3271.

Jarrard, R.D. (1986). Relations among Subduction parameters. *Reviews of Geophysics*, **24**, 217-284.

Jordan, T.E., Isacks, B.L., Allmendiger, R.W., Brewer, J.A., Ramos, V.A. and Ando, C.J. (1983). Andean tectonics related to geometry of subducted Nazca plate. *Geological Society of America Bulletin*, **94**, 341-361.

Juteau, T., Mégard, F., Raharison, L. and Whitechurch, H. (1977). Les assemblages ophiolitiques de l'Occident équatorien: nature pétrographique et position structurale. *Bulletin de la Société Géologique de France*, **19**(7), 1127-1132.

Kay, S.M., Ramos, V.A., Mpodozis, C. and Sruoga, P. (1989). Late Paleozoic to Jurassic silicic magmatism at the Gondwana margin: Analogy to the Middle Proterozoic in North America? *Geology*, **17**, 324-328.

Kehrer, W. and Kehrer, P. (1969). Die oberkretazische San Juan Formation der Westkordillere Ecuadors. *Neues Jahrbuch für Geologie und Paläontologie. Abhandlungen*, Stuttgart, **133**, 1-22.

Kehrer, W. and Van der Kaaden, G. (1979). Notes on the Geology of Ecuador with special reference to the Western Cordillera. *Geologische Jahrbuch*, B **35**, 5-57, Hannover.

Keller, G., Adatte, T., Hollis, C., Ordoñez, M., Zambrano, I., Jiménez, N., Stinnesbeck, W., Alemán, A., and Hale-Erlich, W. (1997). The Cretaceous-Tretiary boundary event in Ecuador: reduced biotic effects due to eastern boundary current setting. *Marine Micropaleontology*, **31**, 97-133.

Kendrick E.C., Bevis, M., Smalley Jr., O. R.F. (1999). Cifuentes and F.

Galban, Current rates of convergence aross the Central Andes: estimates from continuous GPS observations. *Geophysical Research Letters*, **26**, 541-544.

Kennan, L., Lamb, S. and Rundle, C. (1995). K-Ar dates from the Altiplano and Cordillera Oriental of Bolivia: implications for Cenozoic stratigraphy and tectonics. *Journal of South American Earth Sciences*, **8**, 163-186.

Kley, J. and Reinhardt, M. (1994). Geothermal and tectonic evolution of the Eastern Cordillera and the subandean ranges of southern Bolivia. In: *Tectonics of the southern Central Andes*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 155-170. Springer-Verlag, Berlin.

Kley, J. (1999). Geologic and geometric constraints on a kinematic model of the Bolivian orocline. *Journal of South American Earth Sciences*, **12**, 221-235.

Kley, J. (1996). Transition from basement-involved to thin-skinned thrusting in the Cordillera Oriental of southern Bolivia. *Tectonics*, **15**, 763-775.

Klink, B.A., Ellison, R.A. and Hawkins, M.P. (1986). *The geology of the Cordillera occidental and Altiplano, West of the Lake Titicaca, Southern Peru.* Instituto de Geología Minería y Metalurgía, Preliminary report, Lima, 353 p.

Klitgord, K.D. and Schouten, H (1986). Plate kinematics of the Central Atlantic. In: The Western North Atlantic regions, eds. Vogt, P.R. and Tucholke, B.E. *Geological Society of America, Memoir*, The Geology of North America, M, 351-378.

Koch, E. and Blissenbach, E. (1962). Las Capas Rojas del Cretáceo superior - Terciario en la región del curso medio del rio Ucayali, Oriente del Perú. *Boletín de la Sociedad geológica del Perú*, **39**, 7-141.

Kontak, D.J., Clark, A.H., Farrar, E. and Strong, D.F. (1985). The rift-associated Permo-Triassic magmatism of the Eastern Cordillera: a precursor of the Andean orogeny. In: *Magmatism at a Plate Edge, the Peruvian Andes*, eds. Pitcher, W.S. et al., pp. 36-44. Blackie and Halsted Press.

Kontak, D.J., Clark, A.H., Farrar, E., Archibald, D.A. and Baadsgaard, H. (1990). Late Paleozoic-early Mesozoic magmatism in the Cordillera de Carabaya, Puno, southeastern Peru: Geochronology and petrochemistry. *Journal of South American Earth Sciences*, **3**, 213-230.

Kontak, D.J., Clark, A.H., Farrar, E., Pearce, T.H., Strong, D.F. and Baadsgaard, H. (1986). Petrogenesis of a Neogene shoshonite suite, Cerro Moromoroni, Puno, SE Peru. *Canadian Mineralogist*, **24**, 117-135.

Kossler, A. (1997). Jurassic volcaniclastic and carbonate deposits in the Coastal Cordillera of Iquique, northern Chile. *Mem. 1st Cong. Latinoamer. Sediment., Soc. Venezol. Geol.*, **1**, 367-371.

Kulm, L.D., Resig, J.M., Thornburg, T.M. and Schrader, H.-J. (1982). Cenozoic structure, stratigraphy and tectonics of the central Peru forearc. In: Leggett, J.K. (ed.). Trench-Forearc Geology. *Geological Society, Special Paper*, London, **10**, 151-169.

Kummel, B. (1948). Geological reconnaissance of the Contamana region, Peru. *Geological Society of America Bulletin*, **59**, 1217-1266.

Kurth, D., Scheuber, E. and Reutter, K.-J. (1996). The Jurassic-Early Cretaceous arc of the Chilean Coastal Cordillera near Taltal: an assemblage of crustal slivers. Andean Geodynamics II, *3rd International Symposium on Andean Geodynamics-ISAG*, Saint-Malo, France, Orstom publ., Paris, 411-413.

Lamb, S. and Hoke, L. (1997). Origin of the high plateau in the Central Andes, Bolivia, South America. *Tectonics*, **16**, 623-649.

Lamb, S., Hoke, L., Kennan, L. and Dewey, J. (1997). Cenozoic evolution of the Central Andes in Bolivia and northern Chile. In: Orogeny through time, eds. Burg, J.-P. and Ford, M. *Geological Society, London, Special Publication*, **121**, 237-264.

Lapierre, H., Dupuis, V., Bosch, D., Polvé, M., Maury, R.C., Hernandez, J., Monié, P., Yéghicheyan, D., Jaillard, É., Tardy, M., Mercier de Lépinay, B., Mamberti, M., Desmet, A., Keller, F. and Sénebier, F. (1999). Multiple Plume Events in the genesis of the peri-Caribbean Cretaceous Oceanic Plateau Province. *Journal of Geophysical Research* (in press).

Larson, R.L. (1991). Latest pulse of Earth: Evidence for a mid-Cretaceous superplume. *Geology*, **19**, 547-550.

Laubacher, G. and Naeser, C.W. (1994). Fission-track dating of granitic rocks from the Eastern Cordillera of Peru : evidence for late Jurassic and Cenozoic cooling. *Journal of the Geological Society of London*, **151**, 473-483.

Laubacher, G. (1978). Géologie de la Cordillère orientale et de l'Altiplano au Nord et Nord-Ouest du Lac Titicaca (Pérou). *Travaux et Documents de l'ORSTOM*, 95, Paris, 217 p.

Laubacher, G. (1974). Le Paléozoique inférieur de la Cordillére Orientale du sud-est du Pérou. *Cahiers d'ORSTOM*, série Géologique, **6**, 29-40.

Laubacher, G., Sébrier, M., Fornari, M. and Carlier, G. (1988). Oligocene and Miocene continental sedimentation, tectonics and S-type magmatism in the southeastern Andes of Peru (Crucero Basin): geodynamic implications. *Journal of South American Earth Sciences*, **1**, 225-238.

Laughlin, A.W., Damon, P.E. and Watson, B.N. (1968). Potassium-Argon Dates from Toquepala and Michiquillay, Peru. *Economic Geology*, **63**, 166-168.

Laurent, H. and Pardo, A., 1975. Ensayo de interpretacion del basamento del nororiente peruano. *Boletin de la Sociedad Geologica del Peru*, 45, 25-48.

Laurent, H. (1985). El Pre-Cretáceo en el Oriente peruano: su distribución y sus rasgos estructurales. *Boletín de la Sociedad geológica del Perú*, **74**, 33-59.

Lavenu, A. and Cembrano, J., 1998. Active tectonism and neogene state of stress in forearc and intra-arc zones (Andes of Central and southern Chile). *Journal of Structural Geology* (in press).

Lavenu, A. and Mercier, J.L. (1991). Evolution du régime tectonique de l'Altiplano et de la Cordillère orientale des Andes de Bolivie du Miocène supérieur à l'Actuel. *Géodynamique*, **6**, 21-55.

Lavenu, A. (1995). Geodinamica Plio-Cuaternaria en los Andes Centrales: El Altiplano Norte de Bolivia. *Rev. Técnica de YPFB*, **16**, 79-96.

Lavenu, A., Baudino, R. and Égo, F. (1996). Stratigraphie des dépôts tertiaires et quaternaires de la dépression interandine d'Équateur (entre 0° et 2°15'S). *Bull. Instit Français Études Andines*, **25**, 1-15, Lima.

Lavenu, A., Noblet, C., Bonhomme, M., Egüez, A., Dugas, F. and Vivier, G. (1992). New K-Ar age dates of Neogene and Quaternary volcanic rocks from the Ecuadorian Andes: Implications for the relationship between sedimentation, volcanism and tectonics. *Journal of South American Earth Sciences*, 5(3/4), 309-320.

Lavenu, A., Winter, T. and Dávila, F. (1995). A Plio-Quaternary compressional basin in the interandean Depression. Central Ecuador. *Geophysical Journal International*, **121**, 279-300.

Lebrat, M., Mégard, F., Dupuy, C. and Dostal, J. (1987). Geochemistry and tectonic setting of pre-collision Cretaceous and Paleogene volcanic rocks of Ecuador. *Geological Society of America Bulletin*, **99**, 569-578.

Lefèvre, C. (1979). *Un exemple de volcanisme de marge active dans les Andes du Pérou Sud du Miocène à l'Actuel. Zonation et pétrogenèse des andésites et shoshonites.* Sci. Thesis, Univ. S.T. Languedoc, Montpellier, 555 pp., unpublished.

León, I. (1983). *Analyse séquentielle et évolution dynamique du Bassin Tumbes du Nord-Ouest du Pérou.* D.E.A. Univ. Pau, 111 p. (unpublished).

Litherland, M. and 10 more authors (1989). The Proterozoic of Eastern Bolivia and its relationship with the Andean Mobile Belt. *Precambrian Research*, **43**, 157-174.

Litherland, M., Aspden, J.A. and Jemielita, R.A. (1994). The metamorphic belts of Ecuador. British Geological Survey, Keyworth, UK. *Overseas Memoir*, 11, 2 maps. 147 p.

Litherland, M., Klinck, B.A., O'Connor, E.A. and Pitfield, P.E.J. (1985). Andean-trending mobile belts in the Brazilian shield. *Nature*, **314**, 345-348.

López-Gamundí, O.R., and Breitkreuz, C. (1997). Carboniferous-to-Triassic evolution of the Panthalassan margin in southern South America. In: J.M. Dickins (ed.). *Late Palaeozoic and Early Mesozoic Circum-Pacific Events and their Global Correlation*, pp. 8-19. Cambridge Univ. Press.

Lork, A., Miller, H., Kramm, V. and Gravert, B. (1989). U-Pb characteristics of discordant zircons and concordant monazites of Paleozoic granitoids in the Cordillera Oriental, northwest Argentina. *Terra Abstracts*, **1**(1), 351.

Loske, W.P. (1995). 1.1 Ga old zircons in W Argentina: implications for sedimentary provenance in the Paleozoic of Western Gondwana. *Neues Jahrbuch für Geologie und Paläontologie Monatschefte*, **1**, 51-64.

Loughmann, D.L. and Hallam, A. (1982). A facies analysis of the Pucará group (Norian to Toarcian carbonates, organic-rich shales and phosphates) of Central and Northern Peru. *Sedimentary Geology*, **32**, 161-194.

Lucassen, F. and Thirlwall, M.F. (1998). Sm-Nd ages of mafic rocks from the Coastal Cordillera at 24°S, northern Chile. *Geologische Rundschau*, **86**, 767-774.

Macharé, J. and Ortlieb, L. (1992). Plio-Quaternary vertical motions and the subduction of the Nazca Ridge, central coast of Peru. *Tectonophysics*, **205**, 97-108.

Macharé, J. and Ortlieb, L. (1993). Coastal neotectonics in Peru: subduction regime and Quaternary vertical motion. *2nd Int. Symp. And. Geodyn.-ISAG*, Oxford, England, 107-110.

Macharé, J., DeVries, T., Barron, J. and Fourtanier, E. (1988). Oligo-Miocene transgression along the Pacific margin of South America: new paleontological evidence from the Pisco basin (Peru). *Géodynamique*, Paris, 3, 25-37.

Macharé, J., Sébrier, M., Huamán, D. and Mercier, J.-L. (1986). Tectónica cenozoica de la margen continental peruana. *Boletin de la Sociedad Geologica del Peru*, **76**, 45-77.

Malfere, J.-L. (1999). Géochimie et géochronologie du Complexe Métamorphique de Raspas (SW Équateur). *Mémoire DEA Dynamique de la lithosphère*, Univ. Grenoble I, 29 p. + Annexes.

Malfere, J.-L., Bosch, D., Lapierre, H., Jaillard, É., Arculus, R. and Monié, P. (1999). The Raspas metamorphic complex (southern Ecuador): remnant of a late Jurassic-early Cretaceous accretionary prism. Geochemical and isotopic constraints. *4th International Symposium on Andean Geodynamics-ISAG*, Göttingen, IRD ed., sér. Colloques et Séminaires, Paris.

Mamet, B.L. (1996). Late Paleozoic small foraminifers (endothyrids) from South America (Ecuador and Bolivia). *Canadian Journal of Earth Sciences*, **33**, 452-459.

Marocco, R. and De Muizon, C. (1988). Le Bassin Pisco, bassin cénozoïque d'avant-arc de la côte du Pérou central: analyse géodynamique de son remplissage. *Géodynamique*, Paris, 3, 3-19.

Marocco, R. and Noblet, C. (1990). Sedimentation, tectonism and volcanism relationships in two Andean basins of southern Peru. *Geologische Rundschau*, **79**, 111-120.

Marocco, R. (1978). Un segment est-ouest de la chaîne des Andes péruviennes: la déflexion d'Abancay. *Travaux Doct. ORSTOM*, **94**, 195 p.

Marocco, R., Bolaños, R., Paz, M. and Tarazona, A. (1993). Informe preliminar sobre el estudio de la sedimentación neógena del Nororiente del Peru en base al análisis de 9 pozos. *Rapport interne Orstom-Petroperú, Lima*, 8 p. (unpublished).

Marocco, R., Delfaud, J. and Lavenu, A. (1985). Ambiente deposicional de una cuenca continental intramontañosa andina: el Grupo Moquegua (Sur del Perú). Primeros resultados. *Boletín de la Sociedad Geológica del Perú*, Lima, 75, 73-90.

Marocco, R., Baudino, R. and Lavenu, A. (1995). The intermontane Neogene continental Basins of the Central Andes of Ecuador and Peru: Sedimentologic, tectonic and geodynamic implications. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez, R. and Welsink H.J. *American Association of Petroleum Geologists Memoir*, **62**, 597-613.

Marocco, R., Sempere, T., Cirbian, M. and Oller, J. (1987). Mise en évidence d'une déformation paléocène en Bolivie du Sud. Sa place dans l'évolution géodynamique des Andes Centrales. *Comptes Rendus de l'Académie des Sciences, Paris*, **304**(D), 1139-1142.

Marshall, L.G., Sempere, T. and Gayet, M. (1993). The Petaca (late Oligocene-middle Miocene) and Yecua (late Miocene) formations of the Subandean-Chaco basin, Bolivia, and their tectonic significance. In: *Paléontologie et Stratigraphie d'Amérique latine*, ed. Gayet, M., pp. 291-301. Documents des Laboratoires de Géologie, Lyon, 125.

Mathalone, J.M.P. and Montoya, M. (1995). Petroleum geology of the Sub-andean Basins of Peru. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 423-444.

May, S.R. and Butler, R.F. (1985). Paleomagnetism of the Puente Piedra Formation, Central Peru. *Earth and Planetary Science Letters*, 72, 205-21

Mayes, C.L., Lawver, L.A. and Sandwell, D.T. (1990). Tectonic history and New Isochron Chart of the South Pacific. *Journal of Geophysical Research*, **95**, 8543-8567.

McBride, S.L., Robertson, C.R., Clark, A.H. and Farrar, E. (1983). Magmatic and metallogenetic episodes in the northern tin belt, Cordillera Real, Bolivia. *Geologische Rundschau*, **72**, 685-713.

McCourt, W.J., Duque, P., Pilatasig, L.F. and Villagómez, R. (1998). *Mapa geológico de la Cordillera Occidental del Ecuador entre 1° - 2° S., escala 1/200.000*. CODIGEM-Min. Energ. Min.-BGS publs., Quito.

McKee, E.H. and Noble, D.C. (1982). Miocene volcanism and deformatio in the western Cordillera and High Plateaus of south-central Peru. *Geological Society of America Bulletin*, **93**, 657-662.

Mégard, F. (1978). Étude géologique des Andes du Pérou central. *Mémoire ORSTOM*, Paris, **86**, 310 p..

Mégard, F. (1984). The Andean orogenic period and its major structures in Central and Northern Peru. *Journal of the Geological Society, London*, **141**, 893-900.

Mégard, F. (1987). Cordilleran and marginal Andes: a review of Andean geology North of the Arica elbow (18°S). In: J.W.H. Monger and J. Francheteau (eds.). Circum-Pacific orogenic belts and evolution of the Pacific ocean basin. *American Geophysical Union*, Geodynamic series, **18**, 71-95.

Mégard, F., Caldas, J., Paredes , J. and De La Cruz, N. (1996). Geología de los cuadrángulos de Tarma, La Oroya y Yauyos. *Bol. INGEMMET*, Lima, serie A, 69, 3 maps, 279 p.

Mégard, F., Dalmayrac, B., Laubacher, G., Marocco, R., Martinez, C., Paredes, J. and Tomasi, P. (1971). La chaine hercynienne au Pérou et en Bolivie: Premiers résultats. *Cahiers de l'ORSTOM*, série Géologie, **3**, 5-44.

Mégard, F., Noble, D.C., McKee, E.H. and Bellon, H. (1984). Multiple pulses of Neogene compressive deformation in the Ayacucho intermontane basin, Andes of Central Peru. *Geological Society of America Bulletin*, **95**, *1108-1117*.

Mercier, J.L., Sébrier, M., Lavenu, A., Cabrera, J., Bellier, O., Dumont, J.F. and Macharé, J. (1992). Changes in the tectonic regime above a subduction zone of Andean type: the Andes of Peru and Bolivia during the Pliocene-Pleistocene. *Journal of Geophysical Research*, **97**(B8), 11945-11982.

Mitrovica, J.X., Beaumont, C. and Jarvis, G.T. (1989). Tilting of continental interiors by the dynamical effects of subduction. *Tectonics*, **8**, 1079-1094.

Mojica, J. and Dorado, J. (1987). El Jurásico anterior a los movimientos intermálmicos en los Andes Colombianos. In: *Bioestratigrafía de los sistemas Regionales del Jurásico y Cretácico de América del Sur, 1: El Jurásico anterior a los movimientos intermálmicos*, eds. Volkheimer, W. and Musacchio, E., pp. 49-110. Mendoza.

Molnar, P. and Atwater, T. (1978). Interarc spreading and Cordilleran tectonics as alternates related to the age of the subducted oceanic lithosphere. *Earth and Planetary Science Letters*, **41**, 330-340.

Montemurro, G. (1994). Estratigrafía y ambiente sedimentario del Silúrico y Devónico en la cuenca del Chaco boliviano. *11° Congr. Geol. Boliviano, Santa Cruz, Memorias*, 151-160.

Morales, W. (1993). Reinterpretación geológica del area de Lagunitos (NW Perú) en base a sísmica reflexión. *3rd INGEPET*, INGP-055, Lima, 10 fig., 1-19.

Moretti, I., Baby, P., Mendez, E. and Zubieta, D. 1996). Hydrocarbon generation in relation to thrusting in the sub-Andean zone from 18 to 22°S - Bolivia. *Petroleum Geoscience*, **2**, 17-28.

Moretti, I., Díaz-Martínez, E., Montemurro, G., Aguilera, E. and Pérez, M., 1995. The Bolivian source rocks: Sub-Andean, Madre de Dios, Chaco. *Revue de l'Institut Français du Pétrole*, **50**, 753-777.

Morris, R.C. and Alemán, A.R. (1975). Sedimentation and tectonics of middle Cretaceous Copa Sombrero formation in Northwest Peru. *Boletín de la Sociedad geológica del Perú*, **48**, 49-64, Lima

Moulin, N. (1989). *Facies et séquences de dépôt de la plate-forme du Jurassique moyen à l'Albien, et une coupe structurale des Andes du Pérou central*. Thesis Univ. Montpellier, 287 p.

Mourier, T. (1988). *La transition entre Andes marginales et Andes cordillèraines à ophiolites. Evolution sédimentaire, magmatique et structurale du relai de Huancabamba (3°-8°S, Nord Pérou-Sud Equateur)*. Dr. Thesis, Université Paris XI, 275 p. (unpublished).

Mourier, T., Bengtson, P., Bonhomme, M., Buge, E., Cappetta, H., Crochet, J.-Y., Feist, M., Hirsch, K., Jaillard, E., Laubacher, G., Lefranc, J.-P., Moullade, M., Noblet, C., Pons, D., Rey, J., Sigé, B., Tambareau, Y. and Taquet, P. (1988). The Upper Cretaceous-Lower Tertiary marine to continental transition in the Bagua basin, Northern Peru. *Newsletters on Stratigraphy*, **19**, 143-177.

Mpodozis, C. and Ramos, V. (1989). The Andes of Chile and Argentina. In: Geology of the Andes and its relations to energy and mineral resources, eds. Ericksen, G.E., Cañas Pinochetand, M.T. and Reinemund, J.A. *Circum-Pacific Council for Energy and Mineral Resources Earth Science Series, Houston, Texas*, **11**, 59-90.

Mukasa, S.B. and Henry, D.J. (1990). The San Nicolás batholith of coastal Peru: early Palaeozoic continental arc or continental rift magmatism? *Journal of the Geological Society of London*, **147**, 27-39.

Mukasa, S.B. and Tilton, G.R. (1985). Zircon U-Pb ages of super-units in the Coastal Batholith, Peru. In: *Magmatism at a Plate Edge, the Peruvian Andes*, eds. Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D., pp. 203-205. Blackie, Glasgow and Halsted press, New York.

Mukasa, S.B. (1986). Zircon U-Pb ages of superunits in the Coastal Boatholith of Peru: Implications for magmatic and tectonic processes. *Geological Society of America Bulletin*, **97**, 241-254.

Müller, H. and Aliaga, E. (1981). Estudio bioestratigráfico del Cretáceo de la cuenca Marañon.- Petroperu, Lima, report, 57 p. (unpublished).

Muñoz, N. and Charrier, R. (1993). Jurassic-early Cretaceous facies distribution in the Western Altiplano (18°-21°30'S Lat.). Implications for hydrocarbon exploration. *2nd International Symposium on Andean Geodynamics-ISAG, Oxford, Ext. Abst.*, Orstom ed., Paris, 307-310.

Muñoz, N. and Charrier, R. (1996). Uplift of the western border of the Altiplano on a west-vergent thrust system, northern Chile. *Journal of South American Earth Sciences*, **9**, 171-181.

Muñoz, N. and Fuenzalida, R. (1997). La cuenca Arica: un relleno de hemigraben. *VIII Congreso Geológico Chileno, Antofagasta-Chile*, 1830-1833.

Muñoz, N., Venegas, R. and Tellez, C. (1988). La Formación La Negra: Nuevos antecedentes estratigráficos en la Cordillera de la Costa de Antofagasta. *V Congreso Geológico Chileno*, Santiago, 1, 283-311.

Myers, C.M., Beck S., Zandt, G. and Wallace, T. (1998). Lithospheric-scale structure across the Bolivia Andes from tomographic images of velocity and attenuation for P and S waves. *Journal of Geophysical Research*, **103**, 21233-21252.

Myers, J.S. (1975). Vertical crustal movements of the Andes in Peru. *Nature*, **254**, 672-674.

Naeser, C.W., Crochet, J.-Y., Jaillard, E., Laubacher, G., Mourier, T. and Sigé, B. (1991). Tertiary Fission-Track ages from the Bagua syncline (Northern Peru). Stratigraphic and tectonic implications. *Journal of South American Earth Sciences*, **4**, 61-71.

Nakanishi, M., Tamaki, K. and Kobayashi, K. (1989). Mesozoic magnetic anomaly lineations and seafloor spreading history of Northwestern Pacific *Journal of Geophysical Research*, **94**, 15,437-15,462.

Newell, ND., Chronic, J. and Roberts, T. (1953) Upper Paleozoic of Peru. *Geological Society of America*, **58**, 276 p.

Newell, N. (1956). Reconocimiento geológico de la región Pisco-Nazca. *Boletín de la Sociedad Geológica del Perú*, **30**, 261-295.

Noble, D.C., McKee, E.H. and Mégard, F. (1979). Early Tertiary "Incaic" tectonism, uplift and volcanic activity, Andes of Central Peru. *Geological Society of America Bulletin*, Boulder, **90**, 903-907.

Noble, D.C., McKee, E.H., Farrar, E. and Petersen, U. (1974). Episodic Cenozoic volcanism and tectonism in the Andes of Peru. *Earth and Planetary Science Letters*, Amsterdam, **21**, 213-221.

Noble, D.C., McKee, E.H., Mourier, T. and Mégard, F. (1990). Cenozoic stratigraphy, magmatic activity, compressive deformation, and uplift in Northern Peru. *Geological Society of America Bulletin*, **102**, 1105-1113.

Noble, D.C., Sébrier, M., Mégard, F. and McKee, E.H. (1985). Demonstration of two pulses of Paleogene deformation in the Andes of Peru. *Earth and Planetary Science Letters*, **73**, 345-349.

Noble, D.C., Silbermann, M.L., Mégard, F. and Bowman, H.R. (1978). Comendite (peralkaline rhyolites) in the Mitú group, Central Peru: Evidence of Permian-Triassic crustal extension in the Central Andes. *U.S. Geological Survey Journal Research*, **6**, 453-457.

Noble, D.C., Wise, J.M. and Vidal, C.E. (1999). Episodes of Cenozoic extension in the Andean orogen of Peru and their relations to compression, magmatic activity and mineralization. *Sociedade Geológica del Perú*, Lima, vol. Jubilar, 5, 45-66.

Noblet, C., Lavenu, A. and Schneider, F. (1988). Étude géodynamique d'un bassin intramontagneux tertiaire sur décrochements dans les Andes du Sud de l'Équateur: l'exemple du bassin de Cuenca. *Géodynamique*, **3**(1-2), 117-138, Paris.

Noblet, C., Leonardi, G., Taquet, P., Marocco, R. and Córdova, E. (1995). Nouvelle découverte d'empreintes de dinosaures dans la Formation des Couches Rouges (bassin de Cuzco-Sicuani, Sud du Pérou): conséquences stratigraphiques et tectoniques. *Comptes Rendus de l'Académie des Sciences, Paris*, **320**(IIa), 785-791.

Noblet, C., Marocco, R. and Delfaud, J. (1987). Analyse sédimentologique des "Couches Rouges" du bassin intramontagneux de Sicuani (Sud du Pérou). *Bulletin de l'Institut Français d'Études Andines, Lima*, **16**, 55-78.

Norabuena E., Leffler-Griffin, L., Mao, A., Dixon, T., Stein, S., Sacks, I., Ocola, L. and Ellis, M. (1998). Space geodetic observations of the Nazca-South America convergence across the Central Andes. *Science*, **279**, 358-362.

Nürnberg, D. and Müller, R.D. (1991). The tectonic evolution of the South Atlantic from Late Jurassic to present. *Tectonophysics*, **191**, 27-53.

Olivet, J.-L., Bonnin, J., Beuzart, P. and Auzende, J.-M. (1984). Cinématique de l'Atlantique Nord et Central. Publications du Centre National pour l'Exploitation des Océans. *Publications du Centre National pour l'Exploitation des Océans, Série Rapports Scientifiques et Techniques*, **54**, 108 p.

Olsson, A.A. (1944). Contributions to the Paleontology of Northern Peru. Part VII: The Cretaceous of the Paita region. *Bulletin of American Paleontology*, New-York, **28**, 113 p.

Oller, J. and Sempere, T. (1990). A fluvio-eolian sequence of probable middle Triassic-Jurassic age in both Andean and Subandean Bolivia. *1st International Symposium on Andean Geodynamics-ISAG, Grenoble*, Orstom ed., Paris, 237-240.

Ordoñez, M. (1996). Aplicaciones del estudio de microfósiles en la industria petrolera ecuatoriana. *Actas VII Cong. Ecuat. Geol. Min. Petrol.*, Quito, 38-52.

Ortlieb, L., Goy, J.L., Zazo, C., Hillaire-Marcel, C., Ghaleb, B., Guzman, N. and Thiele, R. (1996). Quaternary morphostratigraphy and vertical deformation in Mejillones peninsula, Northern Chile. *3rd Int. Symp. And. Geodyn.*, St Malo, France, IRD publ., Paris, 211-214.

Palacios, O. (1991). El Silúrico-Devónico en el sur del Perú. *Revista Técnica de YPFB*, **12**, 113-117.

Pardo, A. and Sanz, V. (1979). Estratigrafía del curso medio del rio la Leche. *Boletin de la Sociedad Geologica del Peru*, **60**, 251-266.

Pardo, A. (1982). Características estructurales de la faja subandina del Norte del Perú. In: *Simp. Exploración petrolera en las cuencas subandinas de Venezuela, Colombia, Ecuador y Perú*. Assoc. Colomb. Geol. Geofís. del Petróleo, Bogotá.

Pardo-Casas, F. and Molnar, P. (1987). Relative motion of the Nazca (Farallón) and South America plate since late Cretaceous times. *Tectonics*, **6**, 233-248.

Paredes, M. (1958). *Terciario de La Brea y Pariñas y area de Lobitos*. Thesis Univ. Nac. San Agustin, Arequipa, 35 p.

Parraguez, G., Hérail, G., Roperch, P. and Lavenu, A. (1997). Significado tectónico de la geomorfología del margen continental del Norte de Chile (región de Arica). *VIII Congreso Geológico Chileno*, Antofagasta-Chile, 205-209.

Paton, S.McN., 1990. Palaeozoic arc-related volcanism on the Bolivian Altiplano. Pacific Rim 90 Congress, Queensland (Australia). *Australasian Institute of Mining and Metallurgy*, **3**, 565-573.

Pécora, L., Jaillard, É. and Lapierre, H. (1999). Accrétion paléogène et décrochement dextre d'un terrain océanique dans le Nord du Pérou. *C. R. Acad. Sciences, Paris, Earth and Planetary Science Letters*, **329**, 389-396.

Petersen, G. (1949). Condiciones geográficas y geológicas de la Cuenca del río Zarumilla. *Sociedad Geológica del Perú, Lima*, Vol. Jubilar, **7**, 1-40.

Pichowiak, S., Buchelt, M. and Damm, K.-W. (1990). Magmatic activity and tectonic setting of the early stages of the Andean cycle in northern Chile. In: S. Kay and C. Rapela (eds.). Plutonism from Antarctica to Alaska. *Geological Society of America, Special Paper*, **241**, 127-144.

Pilger, R.H. Jr. (1984). Cenozoic plate kinematics, subduction and magmatism. *Journal of the geological Society of London*, **141**, 793-802.

Pitcher, W.S. (1978). The anatomy of a Batholith. *Journal of the Geological Society of London*, **135**, 157-182.

Powell, C.McA., Li, Z.X., McElhinny, M.W., Meert, J.G. and Park, J.K. (1993). Paleomagnetic constraints of timing of the Neoproterozoic breakup of Rodinia and the Cambrian formation of Gondwana. *Geology*, **21**, 889-892.

Pratt, W.T., Figueroa, J.F. and Flores, B.G. (1998). *Mapa geológico de la Cordillera Occidental del Ecuador entre 3° - 4° S, escala 1/200.000*. CODIGEM-Min. Energ. Min.-BGS publs., Quito.

Prévot, R., Chatelain, J-L., Guillier, B. and Yepes, H. (1996). Tomographie des andes équatoriennes: évidence d'une continuité des Andes Centrales. *Comptes Rendus de l'Académie des Sciences, Paris*, **323**(IIa), 833-840.

Prinz, P. (1985). Stratigraphie und Ammonitenfauna der Pucara-Gruppe (Obertrias-Unterjura) von Nord-Peru. *Palaeontographica A*, **188**, 153-197.

Ramos, V.A. (1988). Late Proterozoic-Early Paleozoic of South America - a collisional history. *Episodes*, **11**, 168-174.

Ramos, V.A., Jordan, T.E., Allmendinger, R.W., Mpodozis, C., Kay, S.M., Cortés, J.M. and Palma, M. (1986). Paleozoic terranes of the central Argentine-Chilean Andes. *Tectonics*, 5, 855-880.

Rapela, C.W., Pankhurst, R.J., Casquet, C., Baldo, E., Saavedra, J. and Galindo, C. (1998). Early evolution of the Proto-Andean margin of Gondwana. *Geology*, **26**, 707-710.

Reyes, F.C., Viramonte, J.G., Salfity, J.A. and Gutierrez, W. (1976). Consideraciones sobre el vulcanismo del Subgrupo Pirgua (Cretácico) en el Norte Argentino. *VI Congreso Geológico Argentino, Bahía Blanca*, 1, 205-223.

Reyes, L. and Caldas, J. (1987). Geología de los cuadrángulos de Las Playas, La Tina, Las Lomas, Ayabaca, San Antonio, Chulucanas, Morropon, Huancabamba, Olmos y Pomahuaca. *Boletín del Instituto Geología Minería Metalurgia*, Lima, **39**(A), 83 p.

Reynaud, C., Jaillard, É., Lapierre, H., Mamberti, M. and Mascle, G.H. (1999). Oceanic plateau and island arcs of Southwestern Ecuador: their place in the geodynamic evolution of northwestern South America. *Tectonophysics*, **307**, 235-254.

Reynaud, C., Lapierre, H., Jaillard, É., Benítez, S., Berrones, G. and Mascle, G. (1996). Mineralogical and geochemical characterization of middle Cretaceous to Paleocene oceanic and continental volcanic rocks from southwestern Ecuador. *3rd International Symposium on Andean Geodynamics-ISAG, Saint-Malo, 633-636, ORSTOM ed.*, sér. Colloques et Séminaires, Paris.

Riquelme, R. and Hérail, G. (1997). Discordancias progresivas en el Cenozoico superior del borde occidental del Altiplano de Arica: implicancias tectónicas: *VIII Congreso Geológico Chileno*, Antofagasta-Chile, 231-235.

Rivadeneira, M. and Baby, P. (1999). *La Cuenca Oriente: Estilo tectónico, etapas de deformación y características geológicas de los principales campos de Petroproducción*. Petroproducción-IRD publ., Quito, 88 p.

Rivadeneira, M. and Sánchez, H. (1991). Consideraciones geológicas del PreAptense en la cuenca Oriente. *6to Cong. Ecuat. de Ingen. en Geol., Min., Petrol. y Geotecn., Guayaquil*, 41 p. (inédito).

Rivera, R., Petersen, G. and Rivera, M. (1975). Estratigrafía de la Costa de Lima. *Boletín de la Sociedad geológica del Perú*, **45**, 159-196.

Robert, E., Bulot, L.G., Dhondt, A., Jaillard, É., Villagómez, R., Rivadeneira, M. and Paz, M. (1998). La transgresión del Cretáceo inferior en el margen andino (Perú y Ecuador): datos preliminares. *Boletín de la Sociedad geológica del Perú*, **88**, 73-86, Lima.

Robertson Research (1990). Palynological zonation of the Oriente Basin. Petroperú, Robertson Research, Lima, *Intern. Report*, 78-111.

Rochat, P., Hérail, G., Baby, P. and Mascle, G. (1999). Bilan crustal et contrôle de la dynamique érosive et sédimentaire sur les mécanismes de formation de l'Altiplano. *Comptes Rendus de l'Académie des Sciences, Paris*, **328**, 189-195.

Rochat, P., Hérail, G., Baby, P., Mascle, G. and Aranibar, O. (1998). Analyse géométrique et modèle tectono-sédimentaire de l'Altiplano nord-bolivien. *Comptes Rendus de l'Académie des Sciences, Paris*, **327**, 769-775.

Rodríguez, A. and Chalco, A. (1975). Cuenca Huallaga, Resena geológica y posibilidades petrolíferas. *Boletín de la Sociedad Geológica del Perú*, **45**, 187-212.

Roeder, D. (1988). Andean-age structure of Eastern Cordillera (Province of La Paz, Bolivia). *Tectonics*, **7**, 23-39.

Romeuf N., Aguirre, L., Soler, P., Féraud, G., Jaillard, E. and Ruffet, G. (1995). Middle Jurassic volcanism in the Northern and Central Andes. *Revista geológica de Chile*, Santiago, **22**, 245-259.

Romeuf, N., Aguirre, L., Carlier, G., Soler, P., Bonhomme, M., Elmi, S. and Salas, G. (1993). Present knowledge of the Jurassic volcanogenic formations of the southern Coastal Peru. *2nd International Symposium on Andean Geodynamics-ISAG*, Oxford 1993, Orstom ed., Paris, 437-440.

Romeuf, N., Münch, P., Soler, P., Jaillard, É. and Aguirre, L. (1997). Le volcanisme jurassique inférieur en zone subandine équatorienne. *Comptes Rendus de l'Académie des Sciences, Paris*, **324**(IIa), 361-368.

Roperch, P. and Carlier, G. (1992). Paleomagnetism of Mesozoic rocks from the Central Andes of Southern Peru: Importance of rotations in the development of the Bolivian orocline. *Journal of Geophysical Research*, **97**(B12), 17.233-17.249.

Roperch, P., G. Hérail and M. Fornari (1999a). Magnetostratigraphy of the Miocene Corque basin (Bolivia) Implications for the Geodynamic Evolution of the Altiplano during the late Tertiary. *Journal of Geophysical Research.*, **104**(B9), 20415-20429.

Roperch, P., Fornari, M., Hérail, G. and Parraguez, G. (1999b). Tectonic rotations within the Bolivian Altiplano: implications for the geodynamic evolution of the Central Andes during the late Tertiary. *Journal of Geophysical Research* (in press).

Rosas, S. and Fontboté, L. (1995). Evolución sedimentológica del Grupo Pucara (Triásico superior-Jurásico inferior) en un perfil SW-NE en el Centro del Perú. *Sociedad Geológica del Perú*, Vol. jubilar A. Benavides, Lima, 279-309.

Rosas, S., Fontboté, L. and Morche, W. (1997). Vulcanismo de tipo intraplaca en los carbonatos del Grupo Pucara (Triásico superior-Jurásico inferior, Perú central) y su relación con el vulcanismo del Grupo Mitu (*Permico superior-Triásico*). *IX* Congreso Peruano de Geología. *Sociedad Geológica del Perú*, Lima, Vol. Esp., 1, 393-396.

Rouchy, J.-M., Camoin, G., Casanova, J. and Deconinck, J.-F. (1993). The central paleo-Andean basin of Bolivia (Potosi area) during late Cretaceous and early Tertiary: reconstruction of ancient saline lakes using sedimentological, paleoecological and stable isotope records. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **105**, 179-198

Rüegg, W. (1956). Geologie zwischen Cañete-San Juan, 13°00'-15°24', Süd-Peru. *Geologische Rundschau*, **45**, 775-858.

Rüegg, W. (1961). Hallazgo y posición estratigráfico-tectónica del Titoniano en la costa sur del Perú. *Boletín de la Sociedad geológica del Perú*, **36**, 203-208.

Rutland, R. (1971). Andean orogeny and sea floor spreading. *Nature*, **233**, 252-255.

Sacks, I.S. (1983). The subduction of young lithosphere. *Journal of Geophysical Research*, **88**, 3355-3366.

Sadowsky, G.R. and Bettencourt, J.S. (1996). Mesoproterozoic tectonic correlations between eastern Laurentia and the western border of the Amazonan Craton. *Precambrian Research*, **76**, 213-227.

Salas, G. (1991). Factores geológicos de control de acumulación de hidrocarburos en las cuencas del Oriente peruano. *IV Simposio Bolivariano "Exploración petrolera en las cuencas subandinas"*, Ed. por Asoc. Colomb. Geol. Geofis. Petrol., Bogotá, t. 2, trabajo 29, 14 fig., 15 p.

Salfity, J.A. and Marquillas, R.A. (1994). Tectonic and sedimentary evolution of the Cretaceous-Eocene Salta Group Basin, Argentina. In: *Cretaceous tectonics in the Andes*, ed. Salfity, J.A., pp. 266-315. Earth Evolution Sciences, Fried. Vieweg and Sohn, Braunschweig/Wiesbaden.

Santos, M., Ramírez, F., Alvarado, G. and Salgado, S. (1986). Las calizas del Eoceno medio del occidente ecuatoriano y su paleogeografía. *Actas IV Cong. Ecuat. Ing. Geol. Min. y Petrol.*, Quito, tomo 1, 79-90.

Savostin, L.A., Sibuet, J.C., Zonenshain, L.P., Le Pichon, X. and Roulet, M.J. (1986). Kinematic evolution of the Tethys belt from the Atlantic ocean to the Pamirs since the Triassic. *Tectonophysics*, **123**, 1-35.

Scotese, C.R., Gahagan, L.M. and Larson, R.L. (1988). Plate tectonic reconstructions of the Cretaceous and Cenozoic ocean basins. In: Scoteseand, C.R. and Sager, W.W. (eds.). Mesozoic and Cenozoic Plate Reconstructions. *Tectonophysics*, **155**, 27-48.

Scheuber , E., Bogdanic, T., Jensen, A. and Reutter, K.-J. (1994). Tectonic development of the Northern Chilean Andes in relation to plate convergence and magmatism since the Jurassic. In: *Tectonics of the Southern Central Andes*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 121-139, Springer-Verlag, Berlin.

Schmitz, M. (1994). A balanced model of the southern Central Andes. *Tectonics*, **13**, 484-492.

Schneider, H.-J. (1990). Gold deposits in Lower Paleozoic sediments of the Cordillera Real, Bolivia. In: *Stratabound Ore Deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 137-146. Springer Verlag.

Scholl, D.W., Von Huene, R., Vallier, T.L. and Howell, D.G. (1980). Sedimentary masses and concepts about tectonic processes at underthrust ocean margins. *Geology*, **8**, 564-568.

Schumm, S., Dumont, J.F. and Holbrook, J. (1998). *The mobile Earth: syntectonics, alluvial rivers and sedimentology*. Cambridge University Press (in press).

Sébrier, M. and Soler, P. (1991). Tectonics and magmatism in the Peruvian Andes from late Oligocene time to the present. In: Andean magmatism and its tectinic setting, eds. Harmon, R.S. and Rapela, C.W. *Geological Society of America Bulletin Special Paper*, **265**, 259-278.

Sébrier, M., Lavenu, A., Fornari, M. and Soulas, J.-P. (1988). Tectonics and uplift in Central Andes (Peru, Bolivia and Northern Chile) from Eocene to present. *Géodynamique, Paris*, **3**, 85-106.

Sébrier, M., Mercier, J.L., Mégard, F., Laubacher, G., Carey-Gailhardis, E. (1985). Quaternary normal and reverse faulting and the state of stress in the Central Peru. *Tectonics*, **IV**(7), 739-780.

Seminario, F. and Guizado, J. (1976). Síntesis bioestratigráfica de la región de la Selva del Perú. *Actas del 2ndo Congreso Latinoamericano de geología, Caracas* 1973, **2**, 881-898.

Sempere, T. (1994). Kimmeridgian (?) to Paleocene tectonic evolution of Bolivia. In: *Cretaceous tectonics in the Andes*, ed. Salfity, J.A., pp. 168-212, Earth Evolution Sciences, Fried. Vieweg and Sohn, Braunschweig/Wiesbaden.

Sempere, T. (1995). Phanerozoic evolution of Bolivia and adjacent areas. In: Petroleum Basins of South America, eds. Tankard, A.J. Suárez, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 207-230.

Sempere, T. (1989). Paleozoic evolution of Central Andes (10-26°S). *28th International Geological Congress, Washington D.C., Abstracts*, **3**, 73.

Sempere, T. (1991). Evolución de la cuenca centro-andina (10-26°S) del Cámbrico superior al Silúrico inferior. *Revista Técnica de*

YPFB, **12**, 221-223.

Sempere, T. (1993). Paleozoic to Jurassic evolution of Bolivia. *2nd International Symposium on Andean Geodynamics-ISAG, Oxford*, ORSTOM publ., Paris, 547-550.

Sempere, T., Baby, P., Oller, J. and Herail, G. (1991). La nappe de Calazaya: une preuve de raccourcissements majeurs gouvernés par des éléments paléostructuraux dans les Andes boliviennes. *Comptes Rendus de l'Académie des Sciences, Paris*, **312**(II), 77-83

Sempere, T., Butler, R.F., Richards, D.R., Marshall, L.G., Sharp, W. and Swisher III, C.C. (1997). Stratigraphy and chronology of Upper Cretaceous-Lower Paleogene strata in Bolivia and northwest Argentina. *Geological Society of America Bulletin*, **109**, 709-726.

Sempere, T., Carlier, G., Carlotto, V. and Jacay, J. (1998). Rifting Pérmico superior-Jurásico medio en la Cordillera Oriental de Perú y Bolivia. XIII *Congreso Geológico Boliviano, Potosí*, **1**, 31-38.

Sempere, T., Carlier, G., Carlotto, V., Jacay, J., Jiménez, N., Rosas, S., Soler, P., Cárdenas, J. and Boudesseul, N. (1999). Late Permian-early Mesozoic rifts in Peru and Bolivia, and their bearing on Andean-age tectonics. *4th International Symposium on Andean Geodynamics-ISAG, Göttingen*, IRD (ORSTOM) publ., Paris, 680-685.

Sempere, T., Hérail, G., Oller, J. and Bonhomme, M. (1990). Late Oligocene-early Miocene major tectonic crisis and related basins in Bolivia. *Geology*, **18**, 946-949.

Séranne, M. (1987). Evolution tectono-sédimentaire du bassin de Talara (nord-ouest du Pérou). *Bulletin de l'Institut Français d'Études Andines*, **16**, 103-125, Lima-Paris.

Servant, M., Sempere, T., Argollo, J., Bernat, M., Feraud, G., Lo Bello, P. (1989). Morphogenèse et soulevement de la Cordillère Orientale des Andes de Bolivie au Cénozoïque. *Comptes Rendus de l'Académie des Sciences, Paris*, **309**(II), 417-422.

Shackleton, R.M., Ries, A.C., Coward, M.P. and Cobbold, P.R. (1979). Structure, metamorphism and geochronology of the Arequipa Massif of coastal Peru. *Journal of the Geological Society of London*, **136**, 195-214.

Sheffels, B. (1990). Lower bound on the amount of crustal shortening in the central Bolivian Andes. *Geology*, **18**, 812-815.

Shemenda, A.I. (1994). Subduction: insights from physical modelling. In: *Modern approaches in Geophysics*, Kluwer Academic Publ., Dordrecht, 215 p.

Soler, P. and Bonhomme, M. (1987). Données radiochronologiques K-Ar sur les granitoïdes de la Cordillère Orientale des Andes du Pérou central. Implications tectoniques. *Comptes Rendus de l'Académie des Sciences, Paris*, **304**(II), 841- 845.

Soler, P. and Bonhomme, M. (1990). Relations of magmatic activity to Plate dynamics in Central Peru from Late Cretaceous to Present. In: Plutonism from Antarctica to Alaska, eds. Kay, S. and Rapela, C. *Geological Society of America Memoir*, **241**, 173-191.

Soler, P. and Sempere, T. (1993). Stratigraphie, géochimie et significations paléotectoniques des roches volcaniques basiques mésozoïques des Andes boliviennes. *Comptes Rendus de l'Académie des Sciences, Paris*, **316**(II), 777-784.

Soler, P. (1989). Petrography and geochemistry of lower Cretaceous alkali basalts from the High Plateaus of central Peru and their tectonic significance. *Zentralblatt für Geologie und Paläontologie*, Stuttgart, **1989**(5/6), 1053-1064.

Soler, P. (1991). *Contribution à l'étude du magmatisme associé aux zones de subduction. Pétrographie, géochimie et géochimie isotopique des roches intrusives sur un transect des Andes du Pérou Central. Implications géodynamiques et métallogéniques.* Dr Thesis, Univ. Paris VI, 950 p.

Soler, P., Carlier, G. and Marocco, R. (1989). Evidence for the subduction and underplating of an oceanic plateau beneath the south Peruvian margin during the late Cretaceous: structural implications. *Tectonophysics*, **163**, 13-24.

Somoza, R. (1998). Updated Nazca (Farallon) - South America relative motions during the last 40 My: implications for mountain building in the Central Andean region. *Journal of South American Earth Sciences*, **11**, 211-215.

Starck, D. (1995). Silurian-Jurassic stratigraphy and basin evolution of northwestern Argentina. *AAPG Memoir*, **62**, 251-267.

Steinmann, G. (1929). *Geologie von Peru*. Karl Winter (ed.). Heidelberg, 448 p.

Steinmann, M. (1997). *The Cuenca Basin of southern Ecuador: tectono-sedimentary history and the tertiary Andean evolution*. PhD. thesis, ETH Zürich, n° 12297, 176 pp.

Steinmann, M., Hungerbühler, D., Seward, D. and Winkler, W. (1999). Neogene tectonic evolution and exhumation of the southern Ecuadorian Andes: a combined stratigraphy and fission-track approach. *Tectonophysics*, **307**, 255-276.

Stern, C.R. (1991). Role of subduction erosion in the generation of Andean magmas. *Geology*, **19**, 78-81.

Stern, T.A. and Holt, W.E. (1994). Platform subsidence behind an active subduction zone. *Nature*, **368**, 233-236.

Stewart, J.W., Evernden, J.F. and Snelling, N.J. (1974). Age determinations from Andean Peru: a reconnaissance survey. *Geological Society of America Bulletin*, **85**, 1107-1116.

Stipanicic, P.N. and Rodrigo, F. (1969). El diastrofismo jurásico en Argentina y Chile. *Actas 4° Jorn. Geológicas Argentina*, Buenos-Aires, **2**, 353-368.

Suárez, M. and Bell, C.M. (1991). Depósitos triásicos continentales y Jurásicos marinos en La Cordillera de la Costa, región de Atacama, Chile. *6 Cong. Geol. Chile, Viña del Mar*, Serv. Nac. Geol. Min. Chile ed., Santiago, 664-667.

Suárez, M. and Dalenz, A. (1993). Pteromorphia (bivalvia) noriano de la Formación Vitiacua del área de Villamonte (Tarija). In: Invertebrados y Paleobotánica, vol. II., ed. Suárez-Soruco, R. Fósiles y Facies de Bolivia. *Rev. Técn. YPFB*, Santa Cruz, **13-14**, 155-160.

Suárez-Soruco, R. (1989). El ciclo cordillerano (Silúrico-Carbonífero inferior) en Bolivia y su relación con países limítrofes. *Revista Técnica de YPFB*, **10**(3-4), 233-243.

Suess, E., Von Huene, R. and the Leg 112 Shipboard Scientific Party (1988). Introduction, objectives and principal results, Leg 112, Peru continental margin. In: Suess, E., Von Huene, R. et al. *Proceedings of the Ocean Drilling Program, Initial Reports*, **112**, 5-23.

Tafur, I. (1991). Estratigrafía geológica de la cuenca del Alto Marañon (Departamento de Amazonas). Con especial referencia Comaina-Cenepa-Santiago. *Boletin de la Sociedad Geologica del Peru*, Lima, **82**, 73-97.

Tankard, A.J., Suárez-Soruco, R. and Welsink, H.J., eds. (1995). *Petroleum Basins of South America. AAPG Memoir*, 62, 792 p.

Tarazona, A. (1985). Palinología de la Formación Cushabatay del Pongo de Tiraco, Oriente peruano. *Primer Simposium Nacional del Carbón*, 10-14 Junio 1985, Lima, 20 p.

Tebbens, S.F. and Cande, S.C. (1997). Southeast pacific tectonic evolution from early Oligocene to Present. *Journal of Geophysical Research*, **102**, 12,061-12,084.

Teixeira, W., Gaeta Tassinari, C.C., Cordani, U.G. and Kawashita, K. (1989). A review of the geochronology of the Amazonian craton: tectonic implications. *Precambrian Research*, **42**, 213-227.

Thomas, G., Lavenu, A. and Berrones, G. (1995). Evolution de la subsidence dans le Nord du bassin de l'Oriente équatorien (Crétacé supérieur à Actuel). *Comptes Rendus de l'Académie des Sciences, Paris*, **320**(IIa), 617-624.

Tibaldi, A. and Ferrari, L. (1992). Latest Pleistocene-Holocene tectonics of the Ecuadorian Andes. *Tectonophysics*, **205**, 109-125.

Tosdal, R.M. (1996). The Amazonan-Laurentian connection as viewed from the Middle Proterozoic rocks in the central Andes, western Bolivia and northern Chile. *Tectonics*, **15**(4), 827-842.

The Southern Central Andes

Victor A. Ramos

The Southern Central Andes, as delimited here, encompasses the orogenic belt developed along the Pacific margin of South America between the Arica Deflection (22°S), and the Gulf of Penas (46°S). This southern segment of the Central Andes, as defined by Gansser (1973), extends N-S for over 2500 km. In this region of Argentina and Chile there occur the Highest Andes, with elevations close to 7000m a.m.s.l. in the Cerro Aconcagua (6962 m), the next highest mountain chain after the Himalayas. There are numerous volcanoes, some of which are amongst the highest volcanoes in the world: Ojos del Salado (6887 m) and the Tupungato (6800 m)

The topographic expression of this sector of the Andes of Argentina and Chile as illustrated in Figure 1 shows an intricate structure that is the result of the interaction of a complex basement history with active subduction and changing geometry through time. As a result, different subordinate segments have been recognized, with distinctive characteristics as described by Mpodozis and Ramos (1990). This updated account of these segments is made with reference to four time-related divisions: (1) the Precambrian Basement; (2) the Famatinian Orogen (Early Paleozoic); (3) the Gondwanian Orogen (Late Paleozoic to Early Jurassic); and (4) the Andean Orogen (Jurassic to present).

The Precambrian basement

Although the Andes of Argentina and Chile are developed on a thick continental crust, Precambrian metamorphic rocks are generally poorly exposed under thick Phanerozoic volcano-sedimentary sequences. These isolated outcrops of Andean basement can be grouped in segments based on their lithological characteristics and ages.

Northern segment

The Precambrian-Early Cambrian rocks of the Andean basement between 18°S and 28°S can be subdivided in two different groups of outcrops. A low-grade metasedimentary sequence, known as the Puncoviscana Belt (Omarini, 1983) and a high-grade metamorphic assemblage.

The northwestern border of the Pampia Craton consists of greywacke and shale beds, locally with oceanic mafic rocks that were deposited in the Puncoviscana Basin (see review in Omarini *et al.*, 1999). Farther N, the Puncoviscana Formation continues into the Tucavaca Belt in Bolivia that separates the Arequipa basement of the western margin of the Amazonian Craton. The fossil record of the Puncoviscana Formation indicate a Vendian age based on the fauna and trace fossils (Durand, 1993). Some younger components such as *Oldhamia* would indicate a Tommotian age in the Lower Cambrian. However, the granitic intrusives emplaced in the Puncoviscana Formation yielded ages of 536 and 535 Ma (U/Pb, Bachmann *et al.*, 1987). The age of detrital zircon from the same unit ranges from 560 to 530 Ma (U/Pb, Lork *et al.*, 1989); and the regional metamorphism took place between 540 and 535 Ma (K/Ar, Adams *et al.*, 1990). Based on these data it is assumed that the basin was formed between 700 and 535 Ma.

The Puncoviscana Basin was considered to be either intracratonic by Jezek *et al.* (1985) or a western continental margin open to the ocean of the Pampia Craton (Ramos and Vujovich, 1993). This basin was interpreted as a peripheral foreland basin formed during the latest Proterozoic-Early Cambrian collision of the Pampia Terrane against the Río de La Plata Craton (Kraemer *et al.*, 1995). Strong deformation predated by subduction- related magmatism took place at the Precambrian-Cambrian boundary (Fig. 2). This deformation, known as the Pampean Orogeny, formed the Puncoviscana Belt on the northwestern border of the Pampia Craton.

The high-grade metamorphic basement of northern Chile and Argentina is mainly exposed along the Coastal Cordillera, and in isolated outcrops along the Cordillera de Domeyko and Puna (Hervé *et al.*, 1981; Pacci *et al.*, 1980; Segerstrom and Turner, 1972; Allmendinger *et al.*, 1982). The petrographic and geochemical characteristics of these exposures have been described by Damm *et al.* (1990) in Belén, Quebrada Choja (Sierra Moreno), Peninsula Mejillones, Limón Verde and Salar de Navidad (Fig. 2). The Puna outcrops have been recently described by Becchio *et al.* (1999).

The basement at Belén consists of amphibolite, gneiss, and mica schist preserved in the amphibolite grade. The metabasite rocks are interpreted as oceanic tholeiite of arc affinity and mature island arc rocks, with crystallization ages of 1.877 ± 0.139 and 1.745 ± 0.027 Ga (U/Pb in zircon, Lezaun *et al.*, 1997). Lower intercepts of 456 ± 4 and 366 ± 3 Ma, are consistent with 457 ± 7 and 358 ± 10 Ma K/Ar ages in hornblende (Wörner *et al.*, 1999). These rocks are quite different from the basement of Cerro Uyarani in the western Altiplano of Bolivia, and the Precambrian pebbles described by Tosdal (1996) in the Berenguela area. Zircon from the Uyarani charnockite has a crystallization age of 2.024 ± 0.133 Ga, and a lower intercept of 1.157 ± 0.062 Ga interpreted as a Grenville age of metamorphism (Lezaun *et al.*, 1997). Although Belén and Uyarani lie at the same latitude, major tectonic processes (Fig. 2) may explain their age differences.

Farther to the S at Caleta Loa, Sierra Moreno, and Mejillones, a high-grade metamorphic basement is exposed. Geneiss and migmatite with sedimentary and magmatic protoliths were indentified by

Lucassen and Franz (1996), who recognized high temperature (600 - 700°C) and low-pressure (4 - 6 kbar) metamorphic conditions. $^{143}Nd/^{144}Nd$ isochrons in Quebrada Choja at Sierra Moreno and in Mejillones yielded ages of 525 ± 10 and 509 ± 6 Ma (Becchio *et al.*, 1999), indicating a metamorphic peak between 525 and 509 Ma in the Early Cambrian (Bowring and Martin, 1999). The high-grade metamorphic basement of the Puna exposed in Salar Centenario, Filo de Oire Grande and Salar de Hombre Muerto, as well as the Salar de Antofalla area was recently studied by Becchio *et al.* (1999). The medium to high-grade basement is composed of mica-schist, gneiss and migmatite, which share a common crustal composition with the metamorphic rocks of northern Chile. The basement exposed at Salar de Hombre Muerto has a $^{143}Nd/^{144}Nd$ age of 509 ± 1 Ma, partially confirmed by U/Pb in zircon that yielded 508 ± 19 Ma (Becchio *et al.*, 1999). K/Ar ages between 426 and 393 Ma indicate older exhumation ages than the cooling ages of northern Chile. The protolith of these rocks has T_{DM} model ages of 1.84 ± 0.13 Ga, similar to the 1.65 ± 0.16 Ga of northern Chile or the T_{DM} ages of 1.75 - 1.54 Ga of Belén (Becchio *et al.*, 1999) that may indicate a relatively similar old component in the continental crust.

Based on these considerations the high-grade metamorphic basement of this segment could be considered as part of a major crustal block, the Arequipa-Antofalla Terrane as proposed by Ramos (1988a). Within this block different sectors can be identified as shown in Figure 2.

Sector *a* comprises the Arequipa Massif in southern Peru, where crystallization ages up to 1.9 Ga and metamorphic ages of 1.2Ga to 970 Ma were identified, with partial evidence in the western margin for a lower intercept at 330 Ma (Wastenys *et al.*, 1995). This continental basement with crystallization ages as old as *c.* 2 Ga (early Proterozoic), with a major metamorphic event close to 1.0 Ga (middle Proterozoic, Grenville Orogeny), is also seen at Berenguela, Uyarani and in the subsurface in the San Andrés well (Suárez Soruco, 1999). The 2.0 Ga basement with Grenville metamorphism, with only minor or non-existent Paleozoic reactivation is considered here as the Arequipa Craton (Fig. 2).

A second sector *b* comprises the basement exposures of Belén and Sierra Moreno, in addition to a large basement Bouguer-gravity anomaly described by Götze *et al.* (1994). This basement was strongly remobilized during the Pampean Orogeny (a late phase of the Brasiliano Cycle at these latitudes) at 529 - 509 Ma (Lucassen and Franz, 1996, 1997), interpreted as a metamorphic peak at the Early Cambrian. These rocks have T_{DM} model ages, which indicate an older protolith of near 1.84 ± 0.13 Ga, close to the 1.9 Ga age of the Arequipa protolith.

A third sector *c* corresponds to the basement of Antofalla, where the western magmatic belt of the Puna was emplaced. It is indirectly known through a series of gneiss xenoliths in Cenozoic volcanic rocks in the northern Puna (Coira and Caffe, 1995). This basement has similar characteristics to that of northern Chile, but is the country rock of an Ordovician magmatic arc, and therefore affected by an important Early Paleozoic resetting.

The basement of Mejillones and Caleta Loa, which have similar conditions, has been grouped in the Mejillonia Terrane, which probably amalgamated during the Pampean

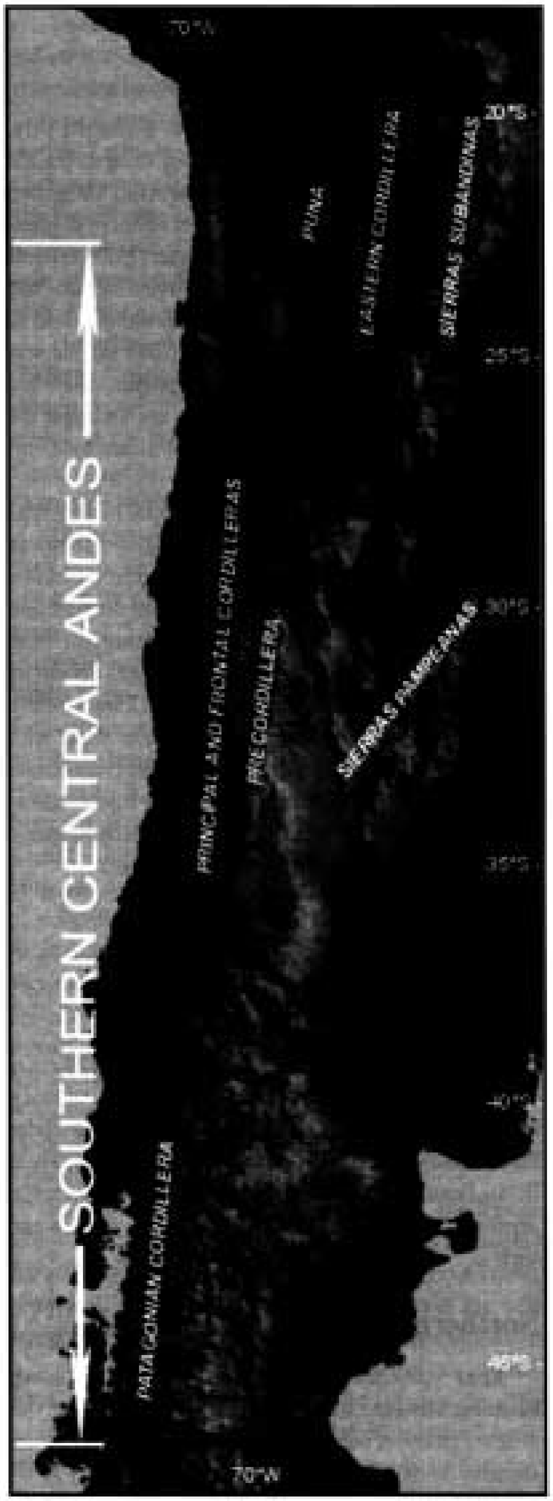


FIGURE 1 - Digital topography of the Southern Central Andes (based on U.S.G.S. data base).

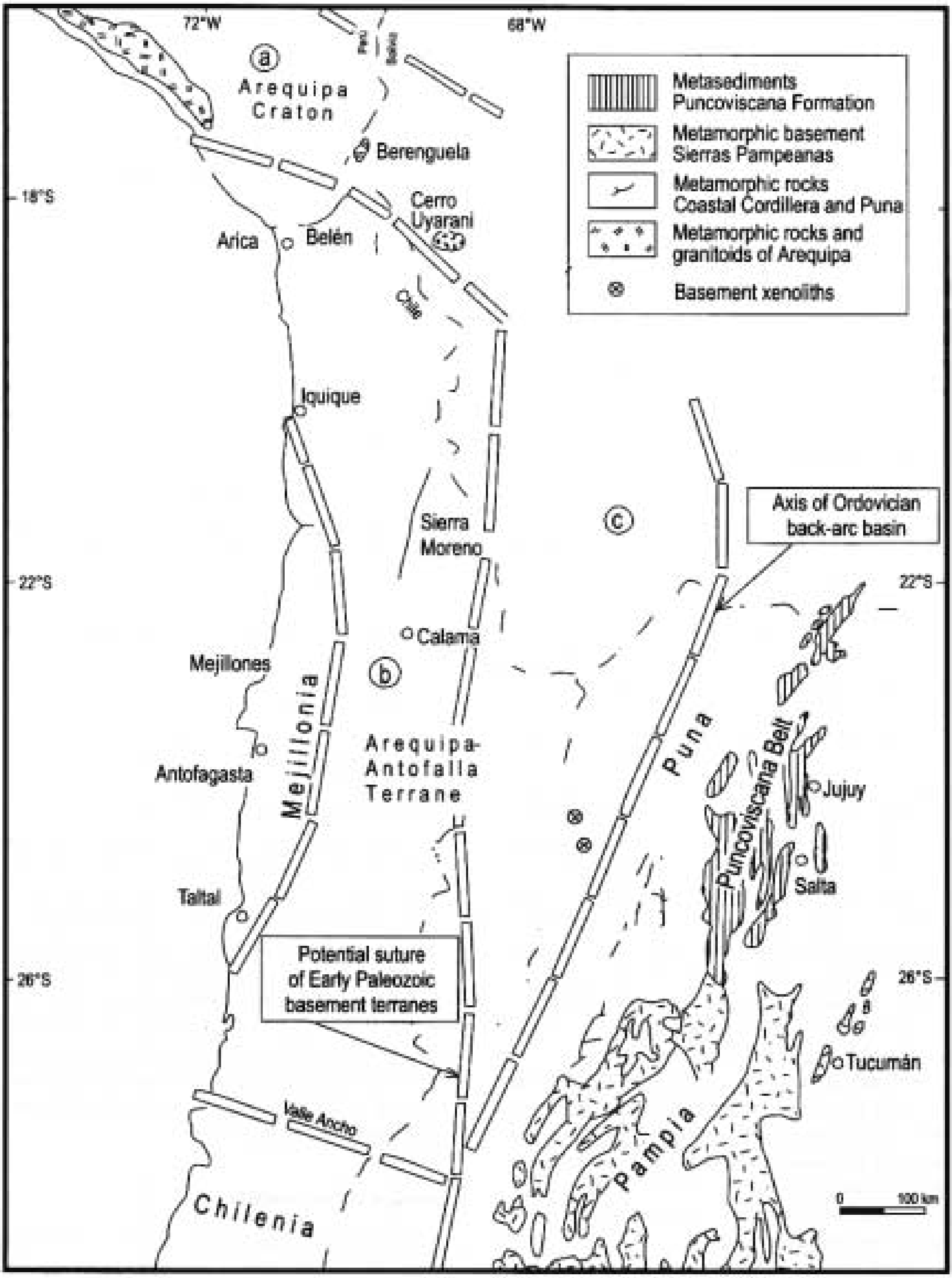


FIGURE 2 - Basement outcrops in the northern segment of Argentina and Chile (after Damm et al., 1990; Tosdal, 1996; Becchio et al., 1999 and Wörner et al., 1999). a, b and c are referred to in the text.

Orogeny to the Arequipa-Antofalla Terrane. The potential suture may coincide with some major crustal discontinuities such as the Atacama or the Domeyko fault systems.

The basement of eastern Puna shares the Pampean Orogeny with the Arequipa-Antofalla Terrane, that led Becchio *et al.* (1999) to postulate a 600 km wide mobile belt, with a N-S trend, with subduction situated on the Pacific border. The magmatic activity of this mobile belt (for discussion, see Rapela *et al.*, 1998) occurs more than 400 km from the continental margin at that time, if the Meso-Cenozoic continental erosion at these latitudes is taken into account (Mpodozis and Ramos, 1990). Therefore, to explain the Pampean magmatic and metamorphic activity as well as the Early Paleozoic history, a series of independent blocks are required.

Central segment

The basement of this segment is exposed between 28°S and 36°S, and corresponds to isolated exposures for which there is indirect evidence in large xenoliths in Miocene volcanic rocks. This basement is part of a composite terrane known as Cuyania (Ramos, 1995), which includes the Precordillera and the Pie de Palo terranes, amalgamated during the Grenville Orogeny, and Chilenia, the substratum of the main Andean Cordillera (Mpodozis and Ramos, 1990).

The Cuyania Terrane is well known in the Sierra de Pie de Palo, the westernmost sector of the Sierras Pampeanas (Fig. 3). Two metamorphic complexes compose this basement. A basic and ultrabasic assemblage with oceanic and island arc affinities, interpreted as a back-arc basin (Vujovich and Kay, 1998), typical of a suprasubduction ophiolite (Ramos *et al.*, 1999) of Grenville age. A second assemblage comprises orthogneiss and mica schist of middle to high metamorphic grade. This basement has been dated by U/Pb in zircon from gneiss and amphibolite between 1.1 and 1.0 Ga (McDonough *et al.*, 1993). It has been interpreted as having formed by island arc accretion between 1.05 Ga and 950 Ma (Vujovich, 1993).

Although the basement of the Precordillera fold and thrust belt is not exposed (Ramos, 1988b), it is well known from metamorphic xenoliths obtained from Miocene volcanic rocks (Leveratto, 1968; Abbruzzi *et al.*, 1993). These xenoliths have been dated by U/Pb in 1.1 Ga, and show geochemical and isotope characteristics similar to the Pie de Palo basement (Kay *et al.*, 1996).

Minor exposures of the basement farther S in Ponón Trehue (Fig. 3) have been dated by U/Pb as Grenville age (Ramos *et al.*, 1998). Even farther S at Las Matras, ages of 1.212 ± 0.047 Ga by Rb/Sr and $^{143}Nd/^{144}Nd$ ages of 1.178 ± 0.047 Ga may indicate the continuation of this basement (Sato *et al.*, 1999).

Based on these isotope and geochronological characteristics the basement of Cuyania has been correlated with the Grenville basement of eastern Laurentia (Ramos *et al.*, 1993).

The basement of the Main Andes is poorly known. It is composed of gneiss and amphibolite of medium metamorphic grade exposed in the Frontal Cordillera, where the geochemical and petrographic characteristics indicate a juvenile volcanic arc source and little transportation (Basei *et al.*, 1998). Preliminary U/Pb data from zircon from the Cordón del Portillo indicate a Grenville age (Ramos and Basei, 1997a, b).

The Precambrian outcrops of Cordillera Frontal (Fig. 4), together with some other metamorphic exposures of possible Precambrian-Early Paleozoic age, such as the La Pampa Gneiss in Chile (Mpodozis and Ramos, 1990), define a large fragment of continental basement known as the Chilenia Terrane. The nature of this basement is indirectly known through the extensive Late Paleozoic-Early Triassic melting related to the Choiyoi Province, composed of large batholiths of granitic composition and thick piles of rhyolite (Kay *et al.*, 1989; Mpodozis and Kay, 1992). The northern boundary of Chilenia is observed in the Valle Ancho Lineament (Fig. 2), where a major crustal boundary is defined by the chemical composition of the Cenozoic volcanic sequences.

Caminos (1993), who analyzed the composition of the gneiss and amphibolite between the Las Tunas and Tunuyán rivers (Fig. 4) made the most complete description of the Chilenia basement. The northern sector consists of quartz-muscovite schist, and marble that Bjerg *et al.* (1992) have considered to be a low-grade metamorphic facies. According to Caminos (1993) the metamorphism increases to the S and to the W, with sillimanite schist and hornblende-plagioclase amphibolite, and gneiss. This author recognized a polyphase deformation in these rocks.

Previous ages in these rocks were almost nonexistent. K/Ar ages indicated 430 ± 15 Ma and 362 ± 15 Ma, probably related to the last deformation episode in the Early Paleozoic. Some Rb/Sr ages of 500 ± 50 Ma, associated with K/Ar ages of 508 ± 30 Ma suggest a late Proterozoic-Early Cambrian age (Caminos, 1993). The U/Pb zircon ages from the western sector of outcrops of the Cordón del Portillo above 4000 m a.m.s.l. in the High Andes of Mendoza, showed 1.069 ± 0.036 Ga at Las Yaretas (Ramos and Basei, 1997a), and 1.081 ± 0.045 Ga at El Portillo (unpublished data). The zircon crystals were obtained from quartz-plagioclase gneiss, bearing biotite and hornblende as the main mafic minerals. The zircon is dominantly prismatic, with euhedral faces, transparent, almost without fractures and inclusions, suggesting an igneous protolith.

Southern segment

A high-grade metamorphic basement has been recognized in the northern sector of the Patagonian Cordillera between 40°S and 43°S since the pioneer work of Ljunger (1932) (Fig. 5). These outcrops were described by González Bonorino (1944) as gneiss and migmatite, as well as mica schist, widely distributed N and S of the City of Bariloche. Ectinite and phyllite of low metamorphic grade have also been described at Colohuincul by Turner (1965) and at Cushamen by Volkheimer (1964).

Previous dating by Rb/Sr presented by Linares *et al.* (1988) yielded Precambrian ages of 1.19 ± 0.016 Ga in the Río Limay basement, whereas K/Ar ages where mostly Paleozoic. Precambrian Rb/Sr ages were also obtained in the Colohuincul area, near San Martín de Los Andes in the metamorphic basement

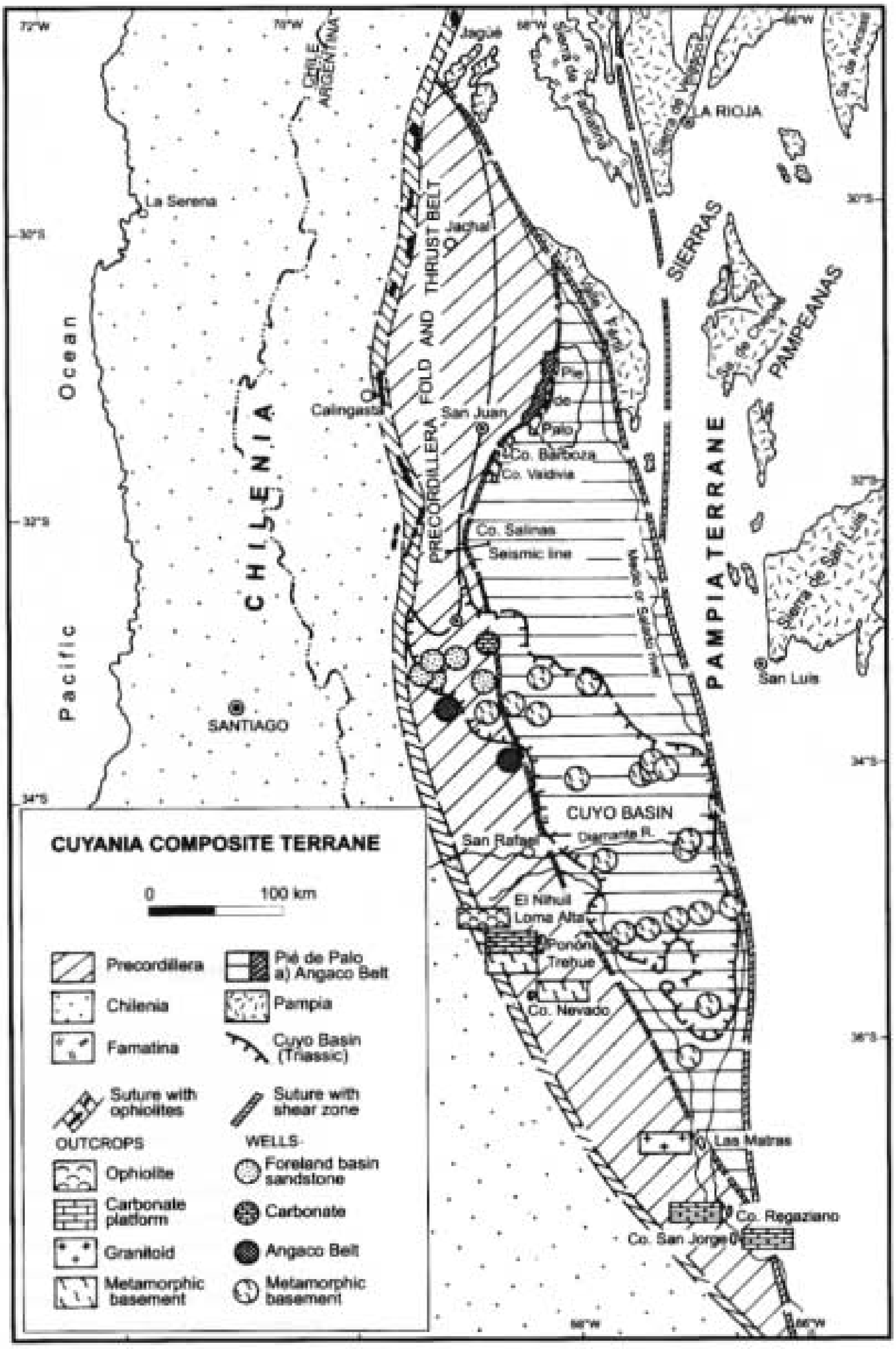


FIGURE 3 - Main basement exposures and subsurface data of the Cuyania Composite Terrane (after Ramos et al., 1998).

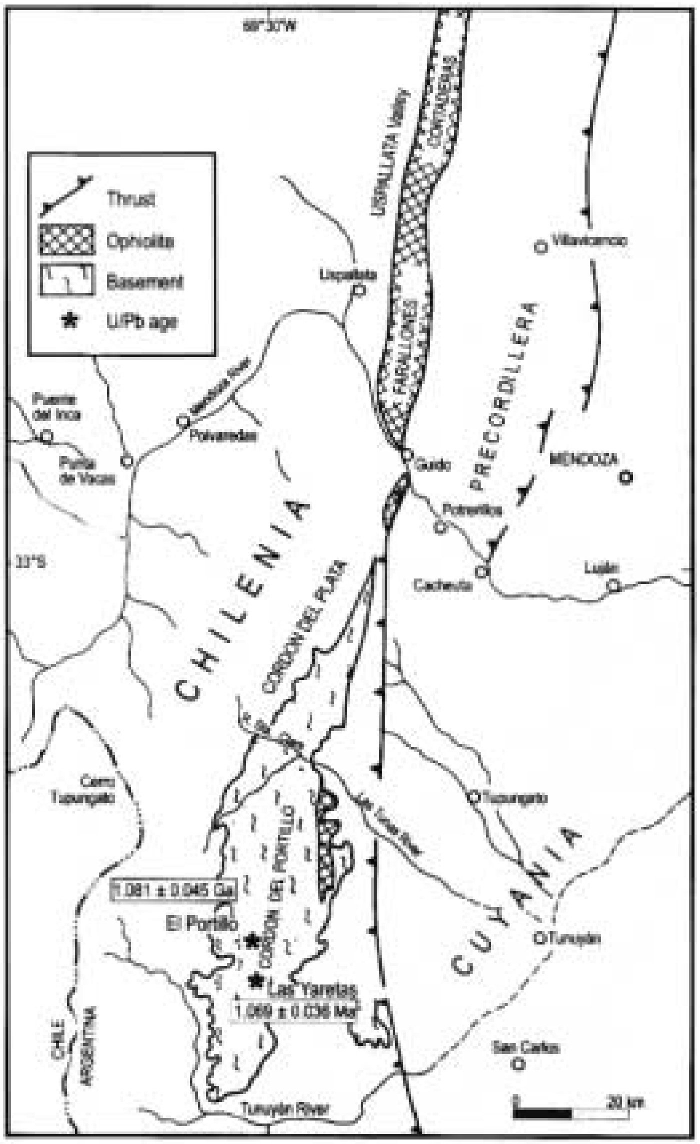


FIGURE 4 - Chilenia's metamorphic basement exposures in the Cordón del Portillo, Cordillera Frontal of Argentina, with indication of the main ophiolite outcrops (after Caminos, 1993).

and in the associated tonalite (Dalla Salda *et al.*, 1991).

Recent studies performed in foliated amphibolite of the basement of Lagos Mascardi and Gutiérrez indicate important Carboniferous magmatic and deformational processes, based on the upper intercept of these rocks (345 ± 4.4 Ma, Basei *et al.*, 1999). The lower intercept indicates 206 ± 40 Ma for the Lago Gutiérrez Granodiorite. The crystalline basement at Río Limay has also U/Pb ages of 334 ± 28 Ma to 269 ± 13 Ma that have been related to the crystallization ages of a magmatic arc (Varela *et al.*, 1999). These new geochronological data of this basement slice indicate that the main magmatic activity and metamorphism is Late Paleozoic. González Bonorino (1944) established that if the correlation with metamorphic rocks from the provinces of Arauco and Concepción in Chile is accepted, then the age of the basement near Bariloche should be considered as Late Paleozoic.

Although the crystallization ages are Paleozoic, most of the T_{DM} model ages indicate a protolith between 1.4 and 1.2 Ga (Basei *et al.*, 1999; Varela *et al.*, 1999), that may show an older continental crust remobilized during the Gondwanian Orogeny.

To the S of Bariloche the basement is exposed at Leleque and in the country rocks of the Cerro Mogote Granitoid (Fig. 5). U/Pb dates on zircon from these rocks yielded 1.5 Ga, showing again an old crustal participation in these rocks. Dalla Salda *et al.* (1994) studied the Cushamen metamorphic rocks in the Río Chico area. The gneiss and migmatite were intruded by foliated granitoid plutons, which yielded Early Paleozoic Rb/Sr ages.

It may be concluded that the southern segment has high and low-grade metamorphic rocks associated with the western extension of the Somuncura Massif (Fig. 5) into the Patagonian Cordillera. However, these rocks represent a Late Paleozoic remobilized Precambrian basement in the northern part, whereas in the southern part they may be related to an Early Paleozoic deformation.

The Early Paleozoic Famatinian Orogen

This orogenic cycle defined by Aceñolaza and Toselli (1976), comprises several collisional episodes that record the amalgamation of some continental terranes during Early Paleozoic times to the protomargin of Gondwana (Ramos, 1988a, b). These terranes are either exotic fragments derived from Laurentia and accreted during the Early Paleozoic or para-autochthonous terranes (Ramos and Basei, 1997b). These para-autochthonous blocks are interpreted as peri-gondwanian derived from western Gondwana that were previously amalgamated to the proto-margin during the Grenville and Brasiliano orogenies, prior to latest Proterozoic-Early Cambrian times.

The outstanding feature of this orogen is the existence of disrupted ophiolitic sequences developed from about 24°S to below 36°S (Borrello, 1969; Haller and Ramos, 1984). These ophiolite rocks have been interpreted as remnants of an oceanic realm (Kay *et al.*, 1984) that was situated between the Gondwana Supercontinent to the E and a series of microcontinental blocks to the W. Terrane accretion is also demonstrated by the development of some calc-alkaline magmatic arcs, presently exposed within the continent, more than 500 km from the present trench. These facts were partially corroborated by paleomagnetic data by Conti *et al.* (1996), Rapalini and Astini (1997), Rapalini *et al.* (1999).

The Famatinian Cycle comprises two main diastrophic events: the Ocloyic Orogeny during the Middle to Late Ordovician and the Chanic Orogeny of early-Middle Devonian times. In order to describe their relative importance three different segments will be considered, matching the description of the metamorphic basement.

The northern segment

This northern segment of Chile and Argentina shows the sedimentary and magmatic history of Cambrian and Ordovician times of the region situated between 22°S and 26°S latitudes. The basement of this region is quite homogeneous (Lucassen and Franz, 1997; Becchio *et al.*, 1999), and records an important collisional event in the Early Cambrian, discussed in the section on the Pampean Orogeny. This event seems to be related to the amalgamation of Mejillonia and Arequipa-Antofalla terranes to the proto-margin of Gondwana.

A major angular unconformity, known as the Tilcaric Event (Turner and Méndez, 1975) separates the intensively folded and deformed late Proterozoic-Infracambrian Puncoviscana Formation from the Cambrian sedimentary cover. Recent studies on Puncoviscana Formation showed medium-anchizonal to epizonal grade metamorphism that in the western border of the basin has prograde biotite indicating greenschist facies metamorphism with intermediate to high pressure (approximately 7 kbar, Do Campo *et al.*, 1998, 1999). K/Ar determinations on the metamorphic facies of Puncoviscana Formation indicate an age between 545 - 535 Ma for the main metamorphic peak (Adams *et al.*, 1990). New Ar/Ar ages on metamorphic mica of this unit yielded older ages for the metamorphism dated at 618 ± 25 Ma (Do Campo, 1999). However, this proposal is not supported by the presence of *Oldhamia* sp. in this unit. *Oldamia* sp. is a well-known trace fossil in the Tommotian. The Ar/Ar dating could reflect detrital mica derived from an eastern Sierras Pampeanas source associated with the Pampean deformation and uplift in a foreland basin setting as proposed by Kraemer *et al.* (1995).

Cambrian-Early Ordovician extension

As indicated in Figure 2, a large sector of northern Chile and Argentina was incorporated to Gondwana in Early Cambrian times. Soon after this amalgamation there is evidence for extension and subsidence. The basal quartzitic wedge of the Mesón Group and the platform deposits of Early to Middle Ordovician age (Fig. 6) unconformably overlie the older deformed rocks. In the Eastern Cordillera there are alkaline basaltic lava flows interbedded with quartzite of the Cambrian Mesón Group and with Early Ordovician rocks (Manca *et al.*, 1987). Sulfide mineralization occurring at the Mina Aguilar in the same region has been

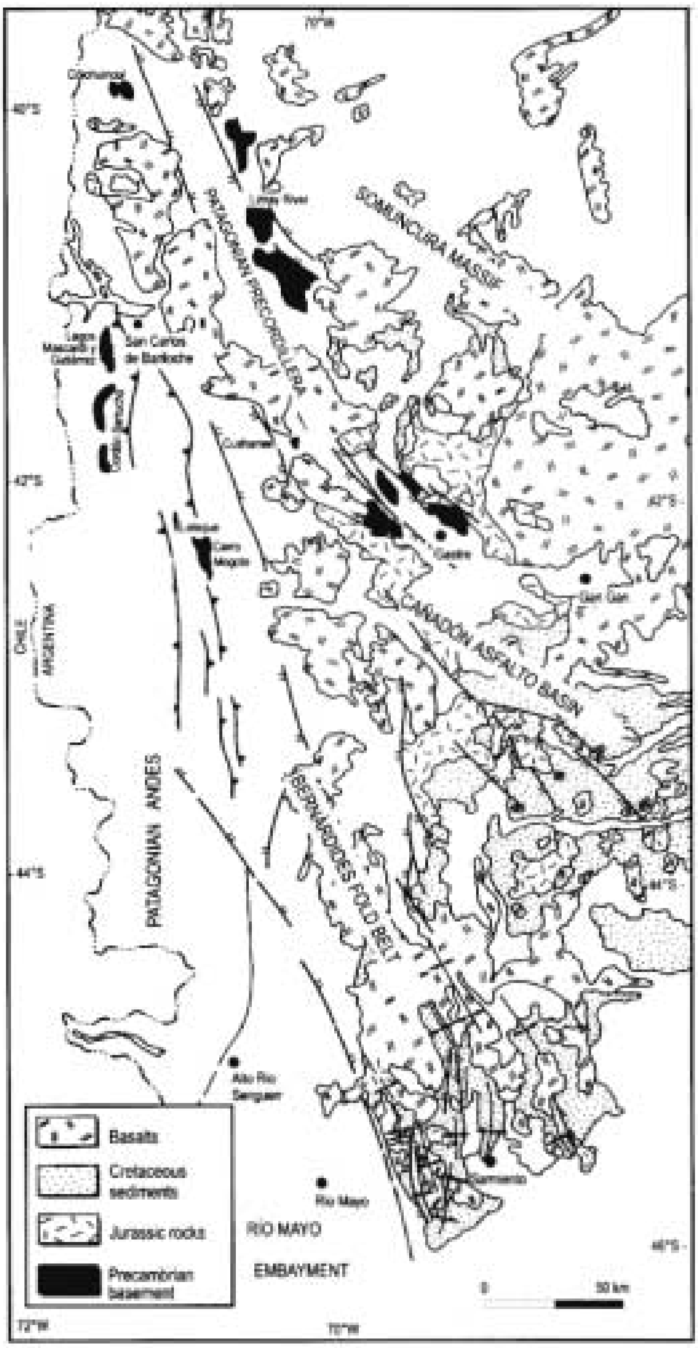


FIGURE 5 - Metamorphic basement exposures in the Patagonian Cordillera. Note the close relationship between the western extension of the Somuncura Massif and the basement outcrops in the extrandean region (after Ramos, 1999a).

interpreted as Early Ordovician Sedex deposits formed in an extensional regime (Pérez *et al.*, 1998).

Evidence of extension in the Puna is also based on the presence of oceanic rocks (Kay *et al.*, 1984; Ramos, 1988a; Coira *et al.*, 1999a, b). These Ordovician rocks, which have a geochemical oceanic signature, have been interpreted as oceanic back-arc rocks (Ramos, 1988a; Blasco *et al.*, 1996) or as island arc rocks (Coira *et al.*, 1999b).

Ordovician magmatic activity

The magmatic rocks in the Puna are associated with two distinct belts, the Faja Eruptiva Occidental (Palma *et al.*, 1986) and the Faja Eruptiva Oriental (Méndez *et al.*, 1973). The western magmatic belt is developed on the Arequipa-Antofalla basement (compare Figs. 2 and 6). The eastern magmatic belt is developed in the independent Puna Terrane of Conti *et al.* (1996), or on the western margin of the Pampia Terrane. Both regions have a common Pampean basement, according to Becchio *et al.* (1999).

Plutons and volcanic rocks characterize the western magmatic belt. The Cordón de Lila Complex and the Choschas plutons in Chile have arc affinities as well as the volcanic rocks interbedded with the sedimentary sequences at Huaytiquina and Guayaos (Coira *et al.*, 1999b). Volcanic rocks vary from low-K tholeiite to medium to high-K andesite from W to E. In the Sierra de Almeida, Late Ordovician granitoid plutons have collisional signatures.

The Ocloyic collisional episode (Ramos, 1986) is related to the closure of the Early Ordovician basin, and according to Bahlburg (1990) produced a thick turbidite sequence in a foreland basin setting. This episode could indicate the final amalgamation of the western Arequipa-Antofalla Terrane to the Gondwana proto-margin (Bahlburg and Hervé, 1997).

On the other hand, the eastern magmatic belt has some uncertainties. The bimodal volcanism is represented by dacite and basalt erupted contemporaneously with the emplacement of shallow level intrusive bodies and active subaqueous sedimentation. Geochemical data for these rocks suggest the melting of a homogeneous source over 200 km long. Radiometric and fossil constraints indicate that these magmas formed over a relatively restricted time span around 476 - 467 Ma in Arenig to Llanvirn times (Coira *et al.*, 1999b). Although some authors have previously assigned this belt to an arc setting (Coira *et al.*, 1982; Ramos, 1986), new data favor an extensional environment, probably related to transtension in an oblique subduction setting (Coira *et al.*, 1999b). This magmatic belt is the northern extension of the Famatinian Magmatic Arc, well-established farther S in the western Sierras Pampeanas (Ramos, 1988a; Pankhurst *et al.*, 1998).

Middle to Late Ordovician deformation

The dominantly penetrative westerly vergent Ocloyic deformation is widely observed in the Puna, which is compatible with previous subduction to the E of the western Arequipa-Antofalla Terrane (Ramos, 1988a; Hongn and Mon, 1999). Collisional granites were emplaced close to the potential suture on the Chilean side (Davidson *et al.*, 1983).

Time constraints for deformation were set in 465 Ma by Salfity *et al.* (1984) based on the fossil record of the Ordovician sedimentary sequence. This coincides with the setting of the foreland basin stage in western Puna (Bahlburg, 1990). However, the foreland basin sequences are also deformed, indicating the final amalgamation of the different terranes during Late Ordovician times. The development of ductile shear zones along the eastern magmatic belt with left-lateral kinematic indicators (Hongn and Mon, 1999) can be explained by oblique subduction.

Some authors explained the Western Magmatic Belt as having been produced from the present trench, in order to reject the existence of any allochthonous terrane (Bahlburg and Zimmermann, 1999). However, the present distance to the trench at 22°S is over 350 km, which is minimal. At least 50 to 80 km of crustal erosion would be required to be added (Stern, 1991), in addition to the Andean shortening that could be in the order of several tens of kilometres.

On the other hand, these simplistic models fail to explain the series of Ordovician granitoid and trondhjemite bodies of Ordovician age exposed to the N and S of Cachi (Fig. 6). This magmatic belt lies along the same trend and has the same age as the Famatinian Magmatic Arc of the Western Sierras Pampeanas.

The proposed subduction zone would have developed immediately to the W of the Cordón de Lila, where previous extension had split the Arequipa-Antofalla Terrane into two separate blocks. Precise location is not possible due to the thick Andean volcanic cover.

The central segment

The Early Paleozoic rocks of this segment between 28°S and 36°S are represented in several distinct structural units: the Precordillera, Sierras Pampeanas, and the San Rafael Block.

The Precordillera is an Andean thrust-and-fold belt that has been developed on Early to Late Paleozoic sedimentary rocks (Fig. 3). The western edge of the Precordillera was strongly deformed during the Famatinian Orogeny by the collision and accretion of an allochthonous terrane in Early to Late Devonian times (Ramos *et al.*, 1984, 1986; Hervé *et al.*, 1987; Mpodozis and Ramos, 1990).

The Sierras Pampeanas (Pampean Ranges) situated in Central Argentina (Figs. 3 and 6), are formed by a series of crystalline basement blocks of Precambrian-Early Paleozoic age that were uplifted and tilted during the Tertiary Andean compression (González Bonorino, 1950) in association with an episode of shallow subduction (Jordan *et al.*, 1983). The resulting structures closely resemble those of the Laramide region of the United States. The basement consists of metamorphic and igneous rocks, which correspond to two distinct orogenic cycles. The oldest Pampean (late Brasiliano) Cycle is preserved along the eastern Sierras Pampeanas, and the metamorphic facies and igneous rocks define a N-S trending belt of late Proterozoic-Early Cambrian age (600 - 520 Ma). The younger Famatinian Cycle rocks are exposed along the Western Sierras Pampeanas (Fig. 7) which define an Early Paleozoic magmatic belt that is associated with a pair of metamorphic belts (Dalla Salda and Varela, 1982). The low temperature belt consists of the highly deformed Caucete Group, and is situated W and

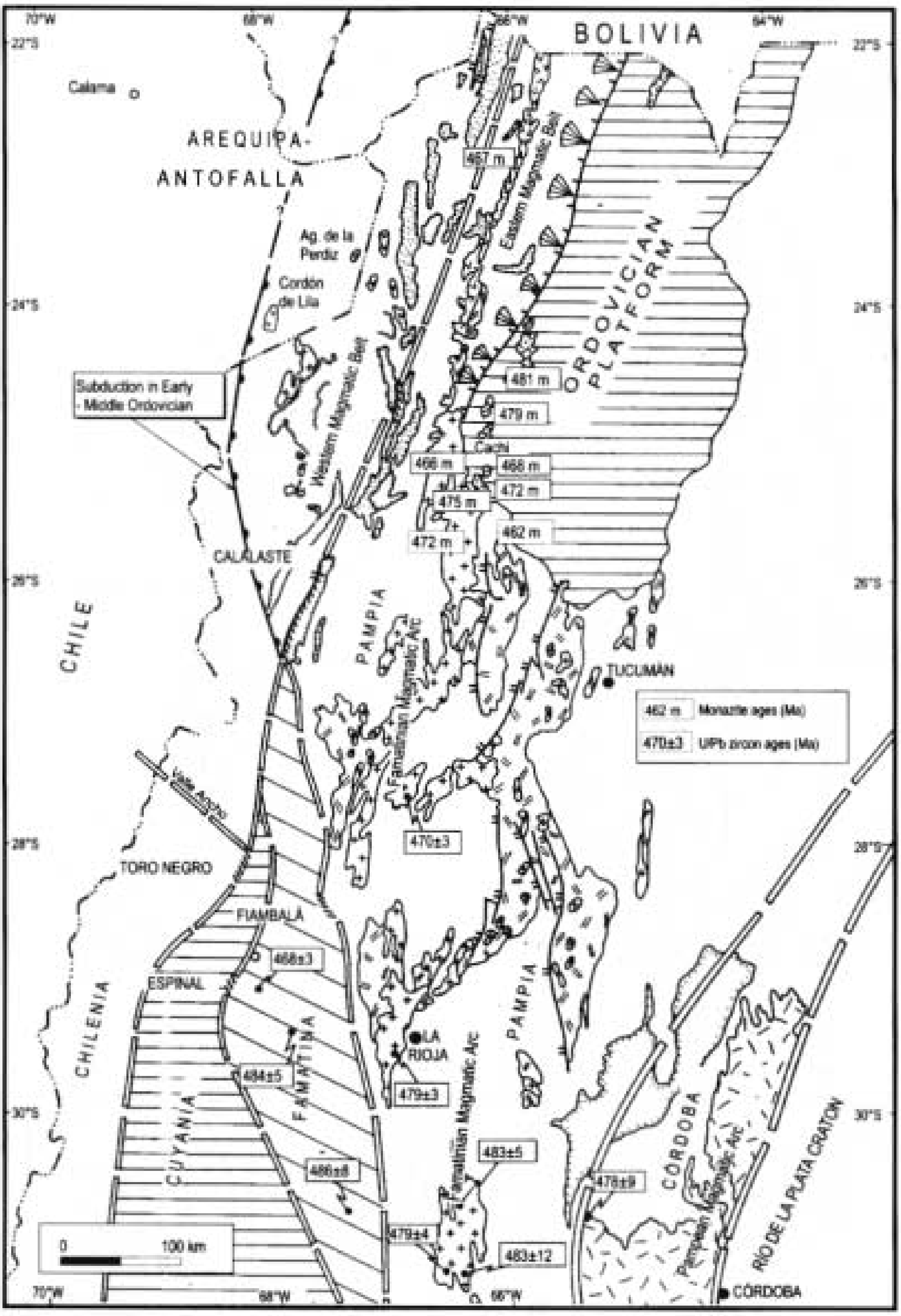


FIGURE 6 - Correlation between the western and eastern magmatic belts of northern Argentina and Chile with the Famatinian Arc further south in western Sierras Pampeanas and Sierra de Famatina (after Bahlburg and Hervé, 1997; Quenardelle and Ramos, 1999; Rapella et al., 1999).

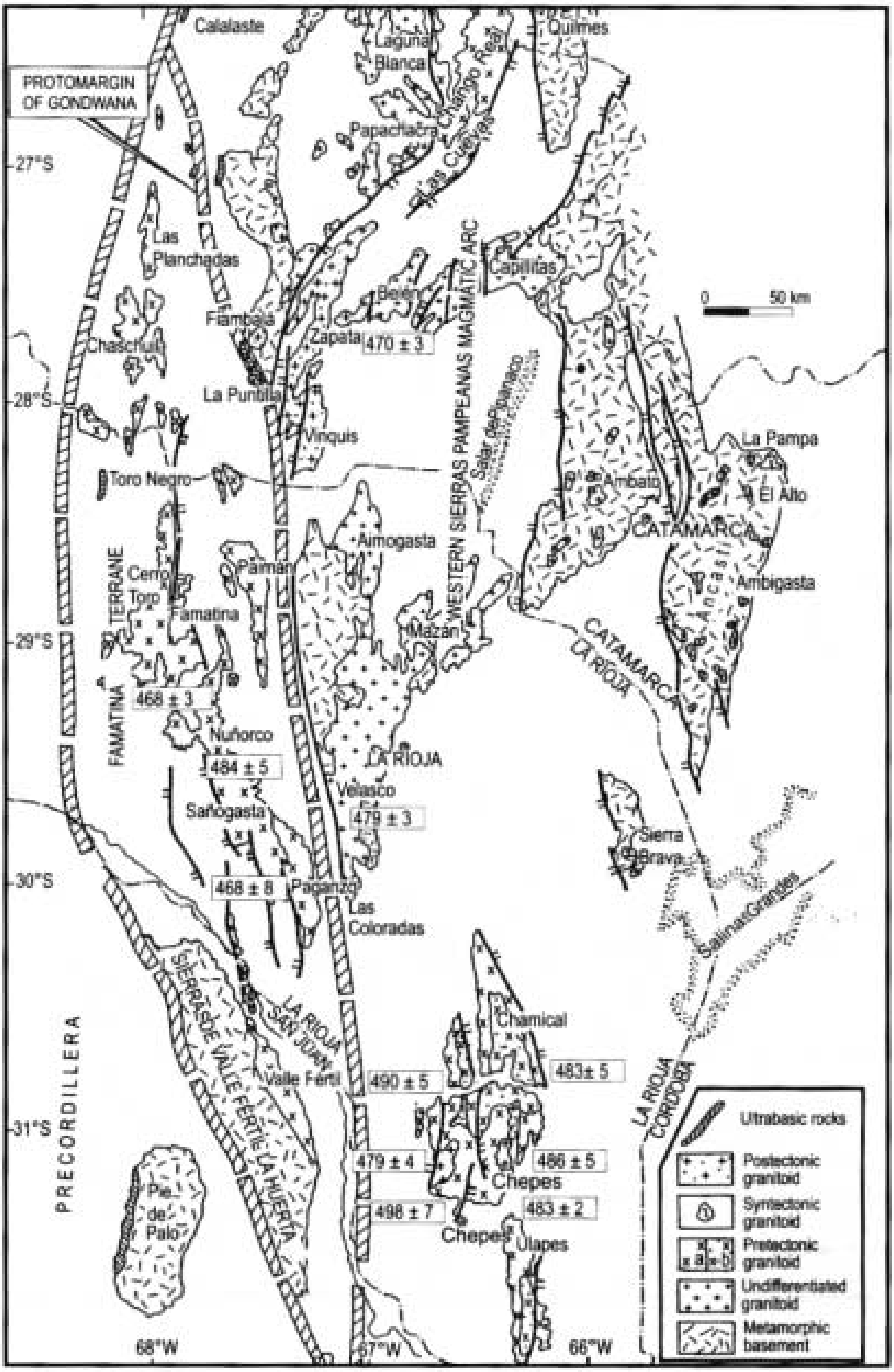


FIGURE 7 - Western Sierras Pampeanas and the Famatina magmatic arcs showing the different granitoid plutons of the Famatinian Orogeny.

oceanward of the middle to high temperature metamorphic belt that comprises the Sierra de Valle Fértil Complex (Ramos and Vujovich, 1995). The western belt shows Late Cambrian to Middle Ordovician magmatic activity, that reached its magmatic climax around 460 Ma (González *et al.*, 1985; Pankhurst *et al.*, 1998; Stuart-Smith *et al.*, 1999).

The San Rafael Bock is a Precambrian-Paleozoic basement peneplain, slightly uplifted during the Andean compression. It is exposed in some metamorphic windows such as at Ponón Trehue (Fig. 3) where it is possible to observe Ordovician platform limestone unconformably overlying the Grenville basement.

Taken as whole, these structural units permit the reconstruction of the Famatinian Orogen. From W to E the following units are identified.

The sedimentary platform

The Early Paleozoic rocks are represented by a carbonate platform of Early Cambrian to Middle Ordovician age. This carbonate platform defines a passive margin (González Bonorino, 1975), that in its northeastern sector preserves the syn-rift deposits (Astini and Vaccari, 1996). A detailed biostratigraphic zoning has been proposed for these highly fossiliferous deposits (Bordonaro, 1980; Baldis *et al.*, 1982; Benedetto, 1998; Benedetto *et al.*, 1999). The Cambrian carbonate rocks contain the typical *Ollenelus* trilobite that permits correlation with Laurentia (Borrello, 1963; Ramos *et al.*, 1986; Astini *et al.*, 1995). Clastic marine Middle to Upper Ordovician rocks cover the platform in the eastern and central sectors, whereas slope and oceanic facies occur to the W. This Early Paleozoic continental margin was deformed in middle Late Ordovician, giving place to Silurian to Devonian clastic foreland basin deposits (Astini *et al.*, 1996) that persisted locally until the Early Carboniferous. The marine sequences are mainly turbidite facies (González Bonorino, 1973), typical of a prodelta environment (Astini, 1990).

A major deformation, known as the Chanic, affected these Early Paleozoic rocks during the Middle Devonian, giving rise to a series of isolated hills known as the "Protoprecordillera" (Amos and Rolleri, 1965). Continental Early Carboniferous alluvial deposits in the western and eastern Precordillera covered the older slope facies.

The magmatic arc rocks

Subduction-related granitoid plutons and volcanic rocks were emplaced in the Western Sierras Pampeanas Belt between 510 to *c.* 470 Ma (Linares and González, 1990), although ages within the 490 - 470 Ma interval are abundant in the northern sector (Lazarte, 1992). The La Puntilla Orthogneiss, described by Grissom *et al.* (1991, 1998) in the Fiambalá area, with U/Pb-ages *c.* 550 Ma, may represent one of the oldest pre-tectonic granitoid plutons, probably related to the Pampean Cycle. In the Fiambalá area there occur gabbro-norite bodies (515 - 500 Ma) with metamorphic ages between 550 to 470 Ma. The age of the last emplacement of Sierra de Famatina granitoid intrusives seems to be slightly younger, with common values of 459 to 450 Ma. Some of these ages are confirmed by scarce U/Pb determinations in zircon (Loske and Miller, 1996).

The syn-collisional granite, identified mainly in the Sierras de San Luis and Ancasti are small bodies concordant with regional foliation and have ages varying from 490 to 470 Ma (Sims *et al.*, 1998; Stuart-Smith *et al.*, 1999). Granite contacts are sharp and are not accompanied by contact aureoles. Plutons are emplaced in migmatite and medium to high-grade schist, associated with pegmatoid facies. Composition varies from that of leucogranodiorite to monzogranite and syenogranite. In general, these granitoid plutons are peraluminous, high K, LILE enriched, and have well-defined crustal signatures (Quenardelle and Ramos, 1999).

The post-collisional granitoid plutons are well represented throughout the entire belt, although volumetrically they seem to be better exposed in the San Luis region. They constitute stocks and large batholiths, in general with circular plan. Locally, these bodies are associated with structural lineaments. Their composition varies from monzogranite to syenogranite with porphyritic textures. Peraluminous rocks are dominant, but metaluminous rocks are also common. Some of them have alkaline affinities. They are potassium-rich, and display characteristics of high temperature and low water content. They have high HFS elements, REE patterns with moderate to high slopes, and small Eu anomalies. These characteristics suggest a thickened crust, with slight arc influence that changed with time to an intraplate setting (Quenardelle and Ramos, 1999).

Available ages from these rocks vary from 440 to 360 Ma, although locally it is possible to recognize several pulses (Linares and González, 1990). The Rb/(Y+Nb) diagram of Pearce *et al.* (1984) summarizes the regional trend and tectonic settings of the different granitoid bodies of the Western Sierras Pampeanas and Famatina (Fig. 8). The analysis of granitoid rock from the northern sector showed that these fall mainly in the syn-collisional field, with a trend toward the within-plate granitoid field. The Famatina granitoid bodies fall within the volcanic arc granite field (VAG) with a trend toward the triple point of VAG-synCOLG (syn-collisional granite) and WPG (within-plate granite). The granitoid plutons of San Luis vary from VAG to WPG.

These chemical characteristics are consistent with post-collisional granite originated and evolved in a continental crust (Pearce, 1996). The granitoid bodies have more affinities with upper crustal granite, as indicated by the low Nb and Y content, controlled by the influence of older subduction processes (Pankhurst *et al.*, 1998; Quenardelle and Ramos, 1999).

Terrane accretion and deformation

The Cambro-Ordovician magmatic belt of the Western Sierras Pampeanas records a series of episodes that can be matched with the sedimentary evolution of the Precordillera Early Paleozoic sequences. A coherent evolution can be proposed when both systems are integrated in a single model (Table 1).

Early Cambrian (545 to 518 Ma)

In this period there is no evidence of subduction-related granitoid or volcanic rocks. It could be a period of collision as proposed by Aceñolaza and Toselli (1984) or a period

FIGURE 8 - Rb vs. (Y + Nb) diagram of Pearce et al. (1984) summarizing the tectonic settings of the different granitoid plutons of western Sierras Pampeanas and Famatina (after Quenardelle and Ramos, 1999).

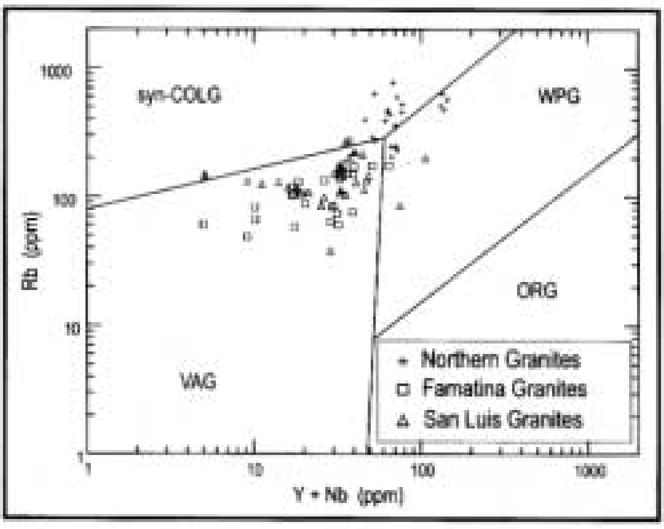


TABLE 1: Correlation chart between the Cuyania Terrane and western and eastern Sierras Pampeanas

	CUYANIA (PRECORDILLERA)	WESTERN SIERRAS PAMPEANAS	EASTERN SIERRAS PAMPEANAS
DEVINIAN	FORELAND DEPOSITS II Cillision of Chlenia 395	WITHIN PLATE GRANITOID AND	ANOROGENIC GRANITOID
SILURIAN	FORELAND DEPOSITS I	PSTCOLLISIONAL GRANITOID Cease of deformation and upflit	
ORDOVICIAN	Collision of Precordillera 460 CLASTIC DEPOSITS Flexural extension Platform drowing 475	SYNCOLLISIONAL GRANITE Collision and defprmation SUBDUCTION RELATED GRANITOID AND VOLCANIC ROCKS (prekinematic granitoid)	
CAMBRIAN	495 CARBONATE PLATAFORM Detachment from Laurentia 535	Brasiliano (Pampean) deformation	RHYOLITIC PLATEAU Final upflit POSTCOLLISIONAL GRANITOID Collision of Pampia with Gondwana
LATE PROTEROZOIC	SYN-RIFT DEPOSITS		SUBDUCTION RELATED GRANITOID

without subduction.

This period coincides with the detachment of Cuyania Composite Terrane from Laurentia, which started its drift towards Gondwana in the Middle to Late Cambrian (Astini *et al.*, 1995, 1996).

Middle Cambrian to Early Ordovician (515 to 470 Ma)

Evidence for the beginning of subduction and arc magmatism along the proto-margin of Gondwana is preserved in a series of volcanic and granitic rocks that were emplaced in the Western Sierras Pampeanas. This happened while the carbonate platform of Precordillera was thermally subsiding, and was isolated from Laurentia (Astini *et al.*, 1996).

Middle to Late Ordovician (470 to 450 Ma)

In this period syn-collisional granitoid plutons were emplaced synchronously with the drowning of the Precordilleran carbonate platform, and subsequent flexural extension. Deformation by ductile thrusting at depth was followed by general uplift of the areas close to the present suture (Ramos *et al.*, 1998).

The Northern Sierras Pampeanas may have had a more complex history. Here, a collision of the Famatina Terrane against the proto-margin of Gondwana was followed by the final collision of the Precordillera with the Famatina Terrane. This could explain the delay of subduction-related magmatism in comparison with the Ancasti area and other northern batholiths.

If correct, the Famatina Terrane was an independent terrane sutured to the proto-margin of Gondwana *c.* 470 - 460 Ma as proposed on paleomagnetic grounds by Conti *et al.* (1996). Quenardelle and Ramos (1999) favoured this alternative (Fig. 9) and disregard the hypothesis of Toselli *et al.* (1996) that proposed that Famatina was separated only by a back-arc basin from the Sierras Pampeanas.

Latest Ordovician to Late Devonian (440 - 360 Ma)

In this period intense anorogenic magmatism occurred in most of Sierras Pampeanas. This cratonization of the Sierras Pampeanas basement was followed in the Early *Carboniferous by the beginning of rifting and sedimentation* of the conglomerate and red beds of the Paganzo Group.

The southern segment

There are only a few Early Paleozoic exposures S of 36°S latitude. Most of the Famatinian Orogeny was preserved in northern Patagonia and in the Ventania region in the Province of Buenos Aires, as the Early Paleozoic trends turned from NNE to WNW. The basement exposed in the Cushamen-Gastre area consists of some metamorphic and magmatic rocks related to the Famatinian Orogen.

The Cushamen region is situated E of the foothills of the Patagonian Andes at 41°45'S (Fig. 5). The Río Chico Complex, which includes the Cushamen metamorphic rocks and the Mamil Choique granitoid plutons (Ordovician, 439 ± 10 Ma by Rb/Sr), represent the southwestern border of the Somuncura Massif. The metamorphic rocks as well as the associated migmatite resulted from a medium to high-grade tectonothermal event that affected a sedimentary protolith of greywacke, pelite, and some quartz-rich sandstone units. The regional schistosity has a NNW trend that controls most of the post-Paleozoic structures. A calc-alkaline post-tectonic granite emplaced in the Río Chico Complex has a Devonian age (387 ± 17 Ma, Rb/Sr), and contrasts with the intraplate leucocratic granite bodies of Permian age (260 ± 5 Ma; Dalla Salda *et al.*, 1994).

These tectonomagmatic events have been related to the Famatinian Orogen, but the paleogeographic setting is uncertain. It may be related to the accretion of the Deseado Massif against the Somuncura basement block as proposed by Ramos and Palma (1996).

The Late Paleozoic-Triassic Gondwanan Orogen

The Gondwanides Orogeny was proposed at an early date by Keidel (1921), to encompass the mountain system formed during the Late Paleozoic deformation observed in the Ventania System in the southern part of the Province of Buenos, and its extension in the Cape Province of South Africa (Keidel, 1917). Du Toit (1937) established the importance of this orogeny in the Southern Hemisphere. The time span of the Gondwanan Cycle is from Early to Middle Carboniferous to Late Triassic-Early Jurassic.

The Late Paleozoic tectonic cycle began with the start of subduction along the present continental margin. Simple subduction of an oceanic plate beneath the continental margin was responsible for sedimentary and tectonic accretion on the W-facing continental edge. A subduction complex developed along the western slope of the Coastal Cordillera from the Late Carboniferous to the Late Triassic (Thiele and Hervé, 1984). Paired metamorphic belts of blue schist rocks and high to medium P/T metamorphic facies that crop out along the Pacific margin were associated with metamorphosed pillow basalt and metachert, which occurred in discontinuous boudinage bands and rootless intrafolial folds (Hervé *et al.*, 1984; Hervé, 1988).

The Meso-Cenozoic continental erosion by subduction has eliminated great parts of the Gondwana subduction complex N of 32°S as suggested by Rutland (1971) and Schweller and Kulm (1978). Stern (1991) described the influence of this tectonic erosion in the magmatism. Von Huene *et al.* (1999) recently described the morphotectonic units along the inner slope of the trench, and the crustal erosion along the continental margin of northern Chile. In spite of this erosion, a reconstruction of the Gondwanan Orogen transverse to the main structural trend is still possible (Mpodozis and Ramos, 1990). Most of the Gondwana sedimentary and metamorphic rocks indicate a significant degree of lateral accretion of the continental margin, probably in sectors associated with discontinuous and isolated pre-Carboniferous basement blocks which constitute the substratum of the Frontal Cordillera.

The Gondwanan Orogeny includes an important diastrophic event: the Sanrafaelic Orogenic Phase (Ramos, 1988b). This compressional phase produced the intense folding and thrusting responsible for the strong angular unconformity between the Permo-Triassic volcanic rocks and the Late Carboniferous to Early Permian turbidite beds that crop out in the Principal Cordillera. This phase of deformation was followed by the resumption of subduction and an important phase of extension that characterized the final stage of the Gondwanan Orogeny, which is coeval with the beginning of the break-up of the Pangea Supercontinent.

The extension that followed the Gondwanan Orogeny was associated with high thermal gradient and widespread melting. However, the forces originated by gravitational instability, failed to reproduce the timing and amount of extension as demonstrated in the Variscides by Henk (1999). The potential source for this generalized rifting could be related to plate reorganization between Laurentia and Gondwana as proposed by Ziegler (1993). This change in the stress regime controlled the magmatic and sedimentary evolution of the Gondwanan Cycle.

Gondwanian magmatism

One of the most outstanding characteristics of the Gondwanan Cycle is the important and extensive felsic magmatism. There is a widespread volcanic cover of rhyolitic composition known as the Choiyoi Volcanics (Groeber, 1946), that since the pioneer work of Zeil (1981) were interpreted as being related to a generalized rifting in the basement of the Andes. However, Polanski (1964, 1972) in his detailed studies of the Late Paleozoic, interpreted these rocks as evidence for typical orogenic magmatism. Years later, Kay *et al.* (1989), Mpodozis and Ramos (1990), Llambías and Sato (1990), and Mpodozis and Kay (1990, 1992) demonstrated that these magmatic rocks represent an early cycle of subduction followed by acid non-orogenic magmatism. Recently, various authors have shown that this last stage is associated with active extensional faulting (Ragona, 1993; Rodríguez Fernández *et al.*, 1995) that preceded the active rifting that characterizes the Middle-Late Triassic sedimentation.

Thick volcanic sequences are dominant in the Frontal Cordillera of Argentina, whereas large batholiths are widely exposed in Chile, although both rock-types are associated in some regions.

Volcanic activity

Late Paleozoic to Early Triassic volcanic rocks are exposed from northern Chile (20°S) to southern Neuquén (42°S) as shown in Figure 10 (after Kay *et al.*, 1989). The solid parts show the present outcrop of Carboniferous and Late Permian-Triassic granitoid plutons and rhyolite flows, and the circles represent oil wells and boreholes in which the Choiyoi Rhyolite has been intersected in the Neuquén Basin. The stripped part shows the inferred original extent of the Province based on the interpretations of Digregorio and Uliana (1980) that a continuous blanket of rhyolite covers the floor of the Neuquén Basin. The stripped part also includes the area indicated by Llambías and Leveratto (1975) and Llambías and Sruoga (1992) in which there occur isolated exposures in the eastern part of the region that are remnants of a larger rhyolite plateau. However, it seems equally likely that the coverage in the eastern part of the region was never continuous. The later Andean Orogeny has exposed rocks at various levels along the Chilean-Argentine border (Kay *et al.*, 1989).

The Choiyoi volcanics consist of basalt, andesite, dacite and rhyolite, in volcanic piles 2 to 4 km thick (Cortés, 1985) in some key areas of the Frontal Cordillera of Mendoza. Subsequent petrological studies in the Cordón del Portillo have demonstrated that the lower section represents a typical calc-alkaline orogenic suite related to subduction, whereas the upper part is of rhyolitic composition and is linked to a generalized extension (Poma and Ramos, 1994). The late Choiyoi volcanic rocks are predominantly rhyolitic to high-Si rhyolitic ignimbrite. In regions where the rocks are unaltered, cooling units can be discerned and several calderas have been recognized (Davidson *et al.* 1985; Pérez *et al.* 1997). The structural evidence of extension is seen along the valley of the Río Mendoza between Uspallata and Polvaredas, as syn-tectonic normal faults affecting the rhyolitic upper section of Cordillera Frontal. In some other areas, conglomerate as initial graben fill are associated with these acid volcanic rocks, as seen in the Cordón del Portillo, where rapid thickness changes are fault controlled.

Main batholiths

The Choiyoi volcanic rocks of rhyolitic and dacitic composition are associated with shallow level plutons of similar composition (Mpodozis *et al.* 1976; Caminos *et al.* 1982). The plutons are often fine-grained, pink to red in color, and display granophyric textures consistent with intrusion at a shallow level. The plutons are typified by the El Leon and El Colorado units of the Elqui Batholith (Nasi *et al.* 1985; Mpodozis and Kay, 1992). Compared to the pre-Choiyoi plutons, the Choiyoi-age plutons are more siliceous and are composed of syenogranite and monzogranite, rather than of granodiorite (Kay *et al.*, 1989; Llambías and Sato, 1990; Mpodozis and Kay, 1992).

The pre-Choiyoi plutonic rocks of the Elqui region in Central Chile are mid-crustal, calc-alkaline granitoid, which compositionally plots in the granodiorite and tonalite fields. They were deformed prior to the intrusion of the late Choiyoi age granite (Las Ingaguas sequence of Nasi *et al.*, 1985). These younger undeformed, shallower level granitoid plutons are progressively more alkaline and siliceous and plot in the monzogranite and syenogranite fields (Mpodozis and Kay, 1992). Similar suites have been recognized in the Argentine slope of the Andean Cordillera (Fig. 11) and have been described in Colangüil (29°S - 30°S) by Llambías and Sato (1990, 1995); in the La Ramada and Aconcagua region (32°S - 33°S) by Pérez and Ramos (1996); and in the Cordón del Portillo by Gregori *et al.* (1996).

These two different suites of granitoid plutons shown in Figure 12 are separated by the Sanrafaelic Orogenic Phase. This orogenic phase has been interpreted either as the result of and accretion and collision of an enigmatic terrane (Mpodozis and Kay, 1992) or to an increase of the convergence rate (Ramos, 1988b) before the cessation of subduction during Late Permian-Early Triassic times (Kay, 1993). This orogenic phase is responsible for a strong

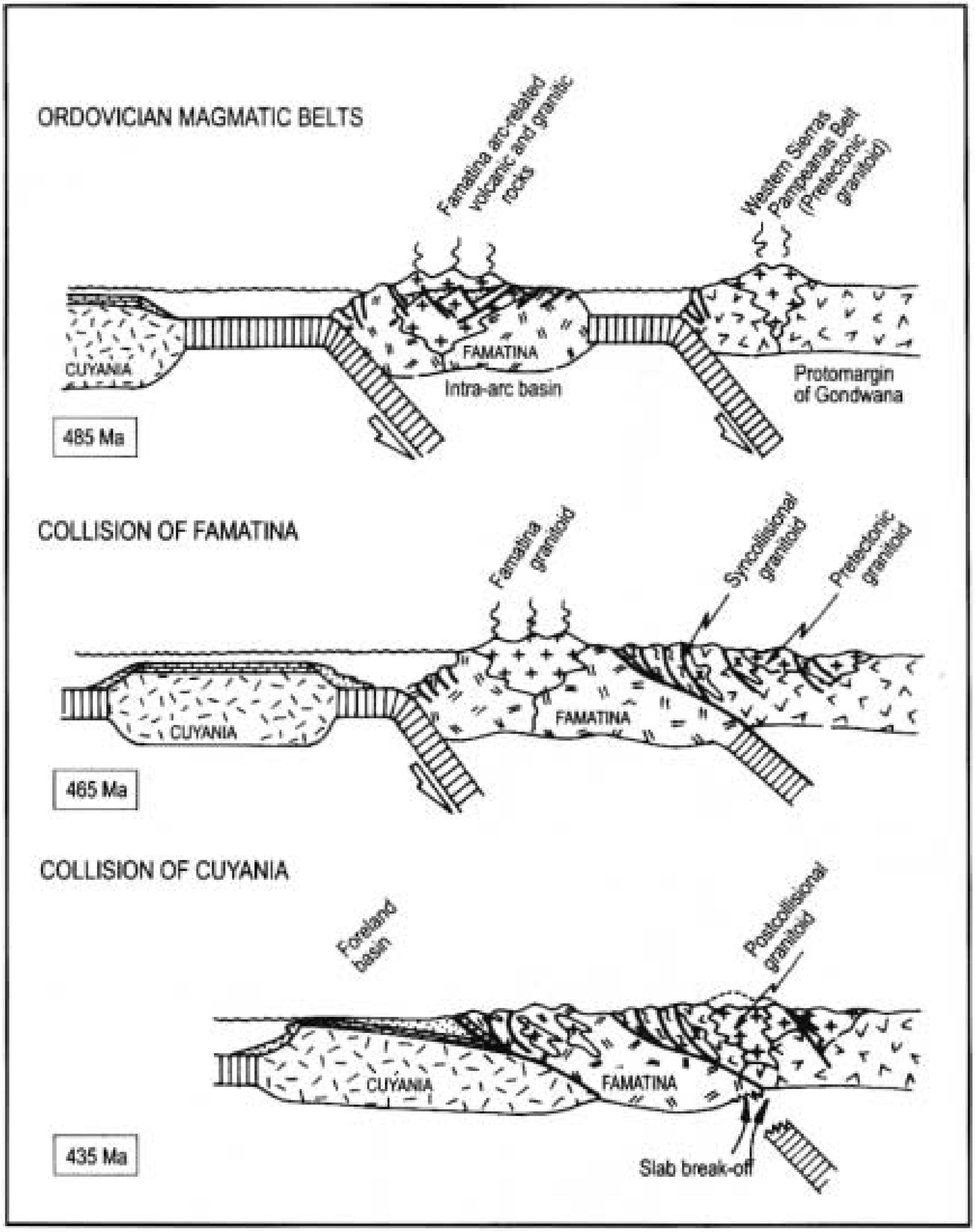


FIGURE 9 - Schematic tectonic evolution of the western proto-margin of Gondwana showing the relationships between Cuyania and Famatina terranes. Cross-cutting relationships between those terranes required that the Famatina Magmatic Arc collided prior to the docking of Cuyania. Precordillera is included as part of the Cuyania Terrane.

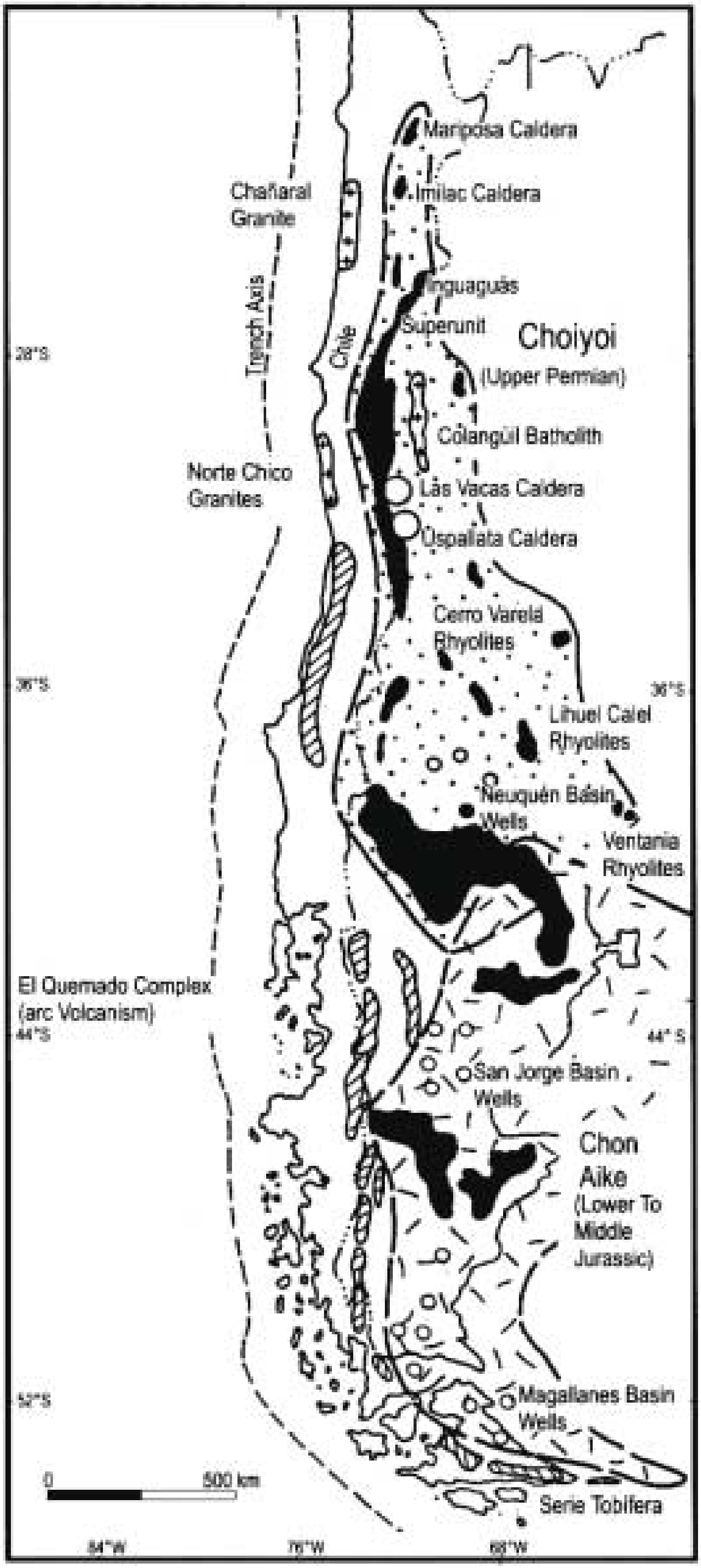


FIGURE 10 - Rhyolitic provinces of Southern South America (after Kay et al., 1989). Black areas are exposures; hatched areas, equivalent volcanic arc rocks; pattern areas, subexposures of rhyolite; and small circles, oil exploration wells.

angular unconformity between Late Carboniferous-Early Permian sedimentary rocks and Late Permian volcaniclastic sequences (Ramos, 1988b).

Sedimentary basins

The Gondwanan Orogen has preserved two successive sedimentary cycles with different tectonic settings. They are represented by two unconformity-bounded sequences. The first was deposited in a series of retro-arc basins developed during active subduction on the Pacific margin in Carboniferous-Early Permian times (Ramos *et al.*, 1986). The second cycle is represented by an apron of rift basins developed along the continental margin over the older accreted terrane during the Triassic (Uliana and Biddle, 1988; Ramos and Kay, 1991).

The basins of the first cycle are widely represented in northern Chile and central Argentina and delineate the eastern border of the Late Paleozoic magmatic arc. These retro-arc basins have links with the Paganzo and Chacoparaná intracratonic basins, and farther S, with the Claromecó Foreland Basin (Fig. 13).

The basal sedimentary sequence lies unconformably on Early Paleozoic deposits deformed during the Chañic Phase of the Famatinian Orogeny. The Visean (350 Ma) to earliest Permian (275 Ma) sequence began with valley fill deposits in the Calingasta-Uspallata Basin (López-Gamundí *et al.*, 1994). By the Namurian, alpine glaciation related to the Protocordillera, fed sediment to the marine basin to the W, and to the nonmarine western Paganzo Basin to the E. The Paganzo Basin has also received important glacial deposits from the cratonic highlands situated farther to the E (González Bonorino, 1991). These Late Paleozoic sequences prograde in the last stage the Protocordillera relief, with the accumulation in the W of more that 2000 m of turbidite sediments that are covered by the early Choiyoi volcanics. Tectonic subsidence has been related either to transtensional deformation (Fernández-Seveso *et al.*, 1993), or to foreland basin development (Limarino *et al.*, 1999a, López-Gamundí *et al.*, 1994). Paleomagnetic data indicate a significant degree of rotation that has been interpreted as evidence of oblique subduction during this cycle (Rapalini and Vilas, 1991; Ramos *et al.*, 1996a).

Farther S, in the northern Patagonia area, the Tepuel-Genoa Basin contains more than 1500 m of sediments deposited between Carboniferous and Early Permian times (González, 1985). Important glacial deposits are well exposed in this basin, in which there also occur shallow to deep platform marine deposits showing an increase in water depth to the W (González Bonorino, 1991).

The second sedimentary cycle of Middle to Late Triassic age was unconformably deposited on the underlying Paleozoic and Precambrian rocks. The angular unconformity is assigned to the Sanrafaelic Orogenic Phase of Early to Late Permian age (Ramos *et al.*, 1996a).

The Triassic basins of central Argentina and northern Chile formed a series of fault-bounded troughs filled with up to 1500 to 2000 m of continental deposits (Charrier, 1979; Uliana and Biddle, 1988) such as the Sierra Exploradora, La Ramada, Ischigualasto, Marayes, Beazley, and Cuyo Basins (Fig. 14). Each basin, as clearly seen in the case of the Cuyo

Basin, has a series of unconnected depocenters, filled with early syn-rift deposits. These independent troughs are linked by the more widespread and extended facies of the sag-fill deposits. The palinspastic restorations of these depocenters along the northwestern trend of the basin show an asymmetric structural control, where the thickest part of the fill repeatedly shifts from the eastern to the western side (Ramos and Kay, 1991). These variations of the depocenters are interpreted as related to the dip of the master normal fault that controls the tensional structure of the basin with a series of alternating detachments as described elsewhere by Bosworth *et al.* (1986), Rosendahl (1987), and Scott and Rosendahl (1989).

The inception of these basins matches a first order boundary among different crustal blocks. For example the Cuyo Basin was developed for more than 700 km along the contact between the Precordillera and Chilenia terranes which were probably amalgamated during the Late Devonian to the Early Carboniferous (Ramos, 1994). The Cuyo Basin was situated to the E of an extensive sequence of mafic oceanic rocks (Kay *et al.*, 1984) interpreted as an ophiolite sequence by Haller and Ramos (1984), Ramos *et al.* (1999). This Triassic basin was therefore developed in the upper block of the proposed suture. The Ischigualasto and Beazley basins show a similar setting (Milana and Alcober, 1994), and they are developed along the border between the Pampia (central Sierras Pampeanas) and the Cuyania (Precordillera) terranes. This boundary corresponds to a slightly older suture developed during Late Ordovician-Early Silurian times. These crustal boundaries also enhanced the Andean deformation during middle to late Tertiary times when a complete tectonic inversion of the basin occurred.

These Triassic basins of Argentina and Chile with an *en échelon* array were probably related to a transtensional regime as proposed by Charrier (1979). This author correlated the continental basins such as Cuyo and the others shown on Figure 14 a and b, with their isolated counterparts exposed along the Chilean margin beneath thick Andean sequences. If we accept a transtensional origin as proposed by Criado Roque *et al.* (1981), and Uliana and Biddle (1988); or a pure extensional regime (Rolleri and Criado Roque, 1968), there is no doubt that during the Triassic a series of rift basins were formed and concentrated along the inter-terrane boundaries.

The generalized extension that began with the Choiyoi Rhyolite continued up to the final rupture of the crust, reactivating ancient weakness zones. The Cuyo and La Ramada basins were developed E and W of the main Choiyoi magmatic focus, and partially coeval with the termination of the acid magmatism (Ramos, 1994; Alvarez and Ramos, 1999).

The late syn-rift deposits are interfingered with alkali basalt that post-dates the main faulting period of the rift. During most of the uppermost Triassic there is no evidence of volcanic activity along these rifts that were dominated by a generalized subsidence related to the thermal decay and sedimentary load that characterize the sag phases (Ramos and Kay, 1992).

Some of the rift systems such as the Sierra Exploradora in northern Chile (25°S) are associated with basalt and acid domes, where the syn-rift deposits have an important volcanic component (Mpodozis and Cornejo, 1997). The rift activity extended until Sinemurian times. This structure imposed an important control in the Andean trend at these latitudes.

At the same time, along the Pacific margin of Chile, near Los Vilos (32°S latitude) there occur isolated outcrops of marine sediments intercalated with low-K tholeiitic basalt to basaltic andesite and rhyolite. This succession has been interpreted as an early primitive stage of an Andean margin volcanic arc (Vergara *et al.*, 1991). However, there are some reservations in this regard. According to Forsythe *et al.* (1987) these rocks may not have been formed on what was part of the continent at that time but rather formed on what was an allochthonous oceanic-floored island-arc terrane that was subsequently accreted to the continent.

A typical Andean arc was only developed later during the Early Jurassic (Mpodozis and Ramos, 1990; Ramos and Alvarez, 1996), although generalized extension continued during most of the Jurassic and Early Cretaceous. Several periods of reactivation have been observed in some of the rift systems, as for example in the Cuyo Basin, in which rifting is associated with new sedimentary sequences and basaltic magmatism that shifted eastward towards the craton.

The Jurassic to Cenozoic Andean Orogen

The Southern Central Andes show important sedimentation and magmatism from Early Jurassic times, developed in close association with the magmatic arc. Since the pioneer work of Groeber (1946, 1947a, b), the Mesozoic and Cenozoic sequences were divided into a series of unconformity-bounded sequences that could be identified along this segment of the Andes. The geology of the region will be described in terms of two major cycles: the Jurassic to early Paleocene cycle, and the Eocene to present cycle.

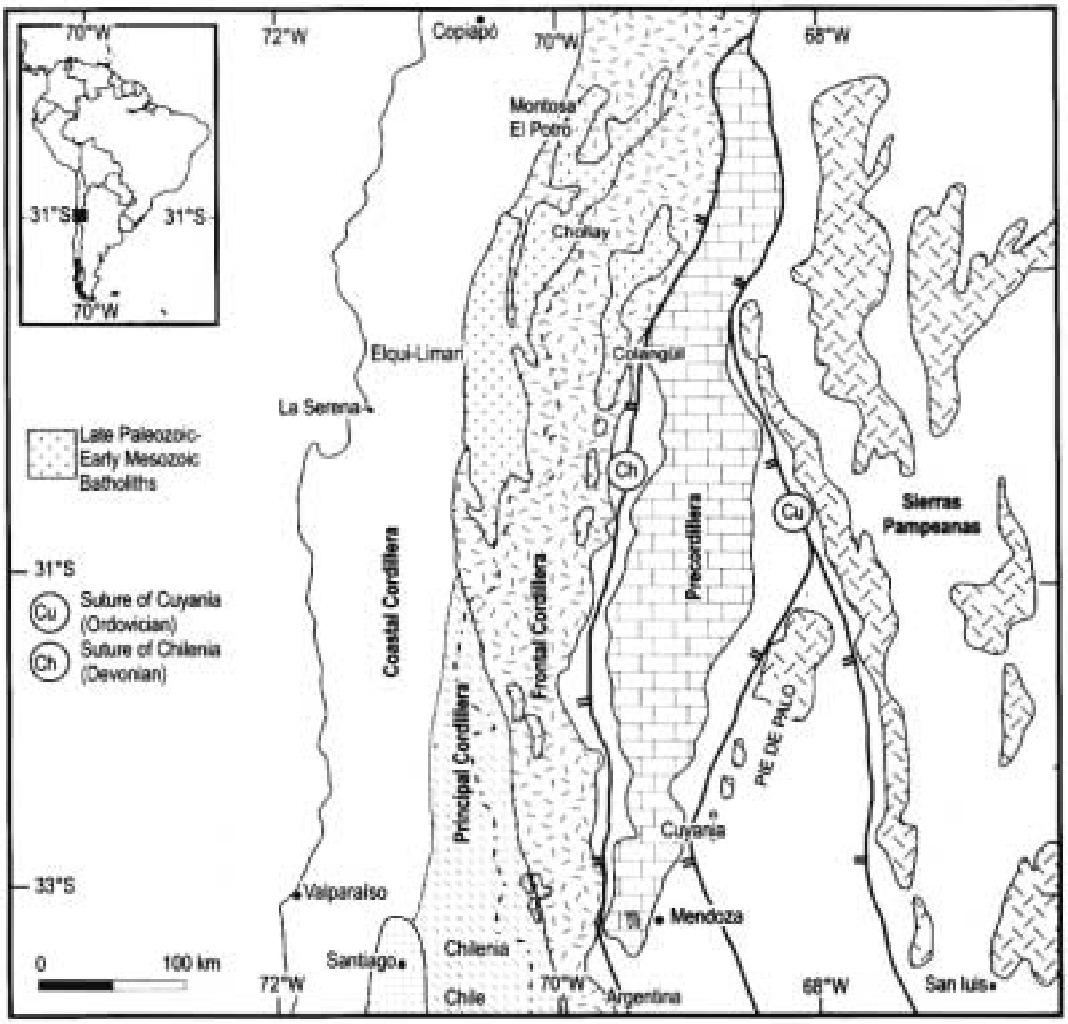
Jurassic to early Paleocene evolution

Based on Groeber's framework, several authors have reviewed the paleogeographical, sedimentological, magmatic and tectonic evolution of the region (Legarreta and Uliana, 1996; Riccardi *et al.*, 1992; Gulisano *et al.*, 1984). In order to describe the varying tectonic settings, several segments have been recognized following Mpodozis and Ramos (1990) (Fig. 15).

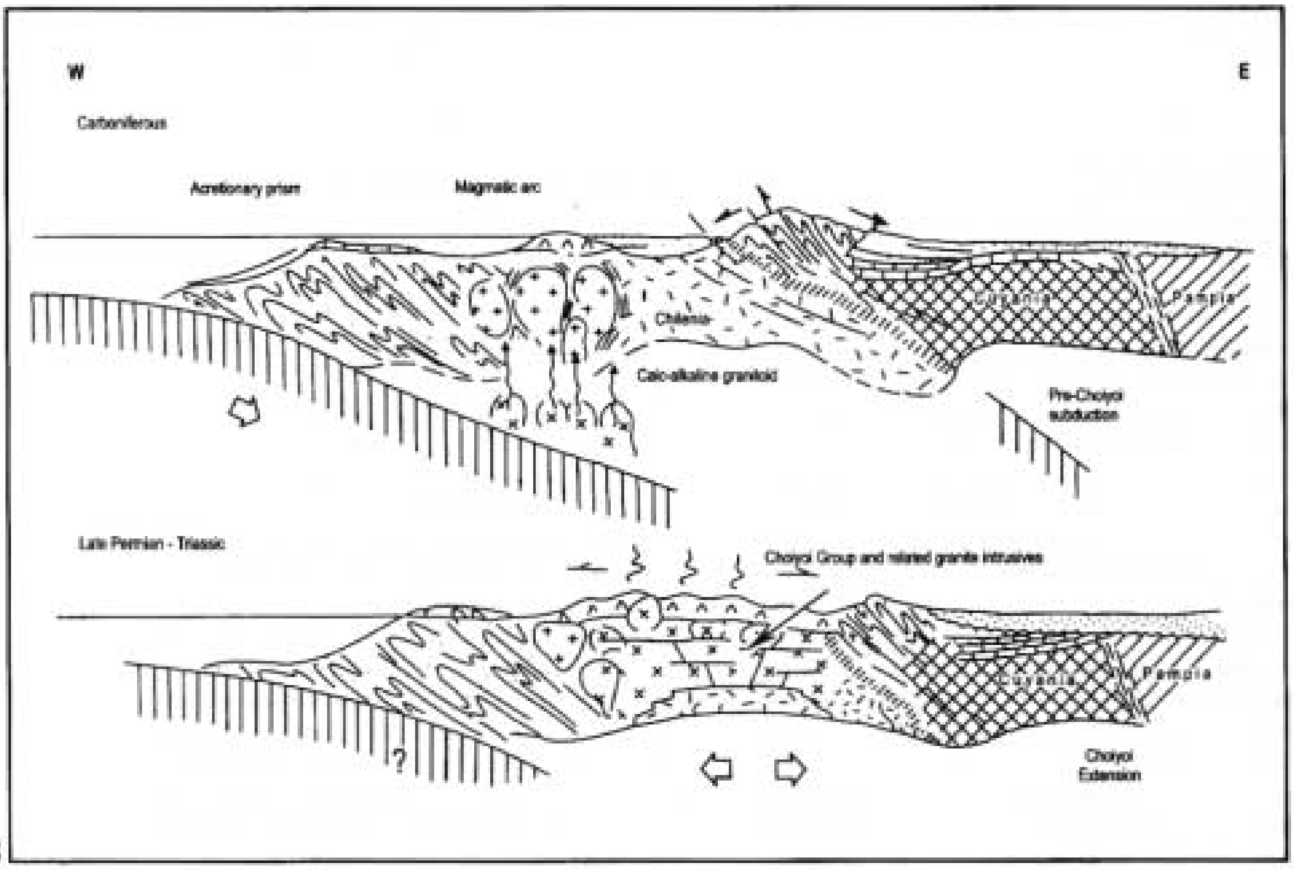
Northern segment: 22°S to 27°S

Most of the Jurassic succession along this segment is preserved on the Chilean side, along the Coastal Cordillera and in the western foothills of Cordillera de Domeyko (Fig. 16).

The Tarapacá Basin developed in the Early Mesozoic from 21°S to about 28°S in a retro-arc setting closely associated with La Negra Magmatic Arc (Naranjo and Puig, 1984; Prinz *et al.*, 1994). The earliest record in this basin is a Late Triassic marine transgression known in the Cordillera de Domeyko associated with syn-rift facies present along the coast, S of Taltal (Chong and von Hillebrandt, 1985; Suárez *et al.*, 1985; Suárez and Bell, 1993).



70°W
72°W
Copiapó
70°W
68°W
Montosa
El Potro
31°S
31°S
70°W
Chollay
Elqui-Limarí
Colangüil
La Serena
Late Paleozoic-Early Mesozoic Batholiths
Ch
Cu
Sierras Pampeanas
Coastal Cordillera
Frontal Cordillera
Precordillera
31°S
Suture of Cuyania (Ordovician)
Suture of Chilenia (Devonian)
Principal Cordillera
PIE DE PALO
Cuyania
33°S
Valparaíso
Mendoza
0
100 km
Santiago
Chilenia
72°W
70°W
68°W
San luis
Chile
Argentina



W
E
Carboniferous
Acretionary prism
Magmatic arc
Chilenia
Cuyania
Pampia
Calc-alkaline granitoid
Pre-Choiyoi subduction
Late Permian - Triassic
Choiyoi Group and related granite intrusives
Cuyania
Pampia
?
Choiyoi Extension

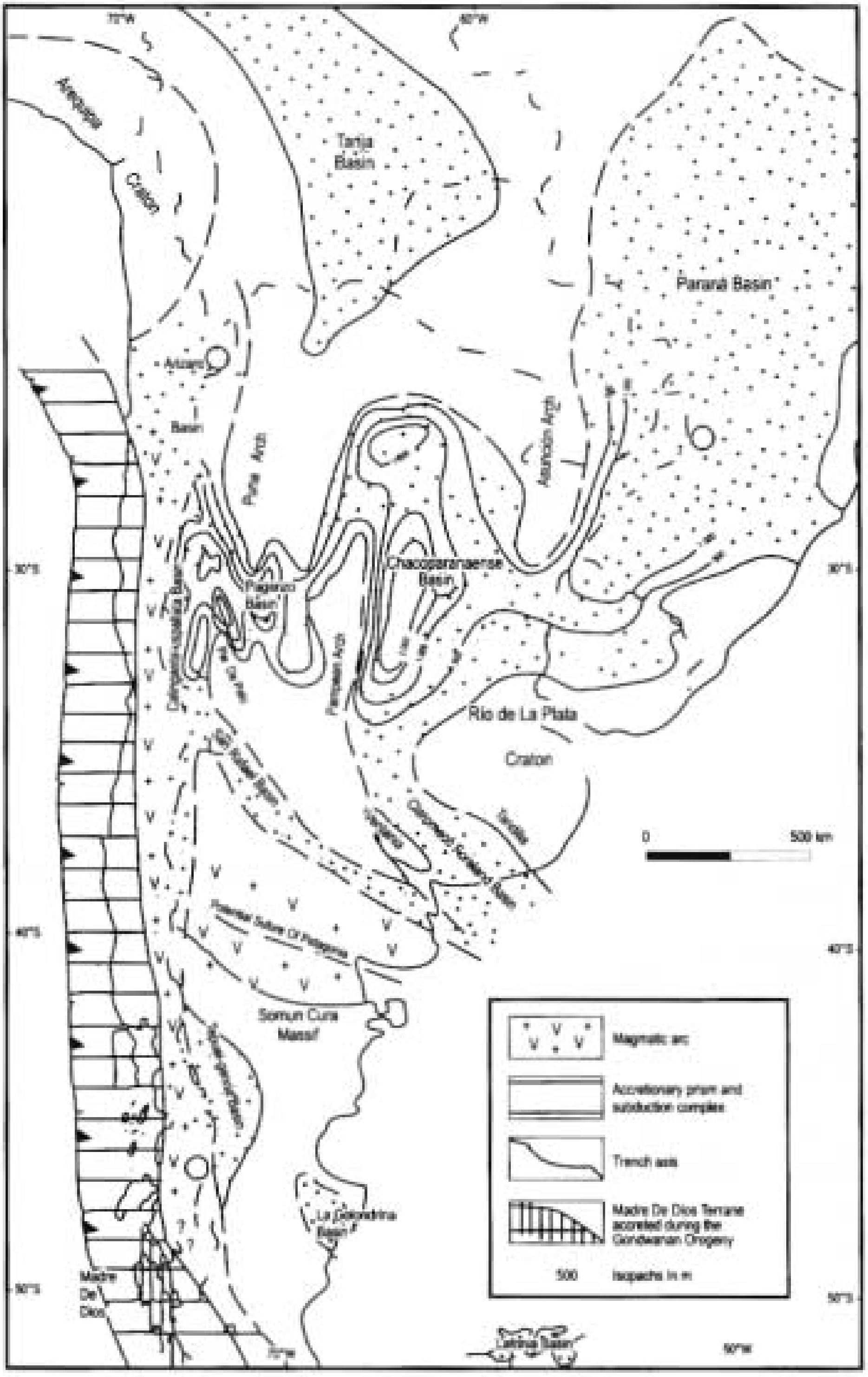


FIGURE 13 - Late Paleozoic basins of the Southern Central Andes (after Azcuy, 1985; Ramos et al., 1986; López Gamundí et al., 1994).

FIGURE 11 - Main batholiths and stocks of Gondwanan granitoids in Argentina and Chile (after Mpodozis and Kay, 1992, Pérez and Ramos, 1996).

FIGURE 12 - Scheme of the tectonic processes associated with the Gondwanian magmatism before and after the Sanrafaelic phase (after Mpodozis and Ramos, 1990).

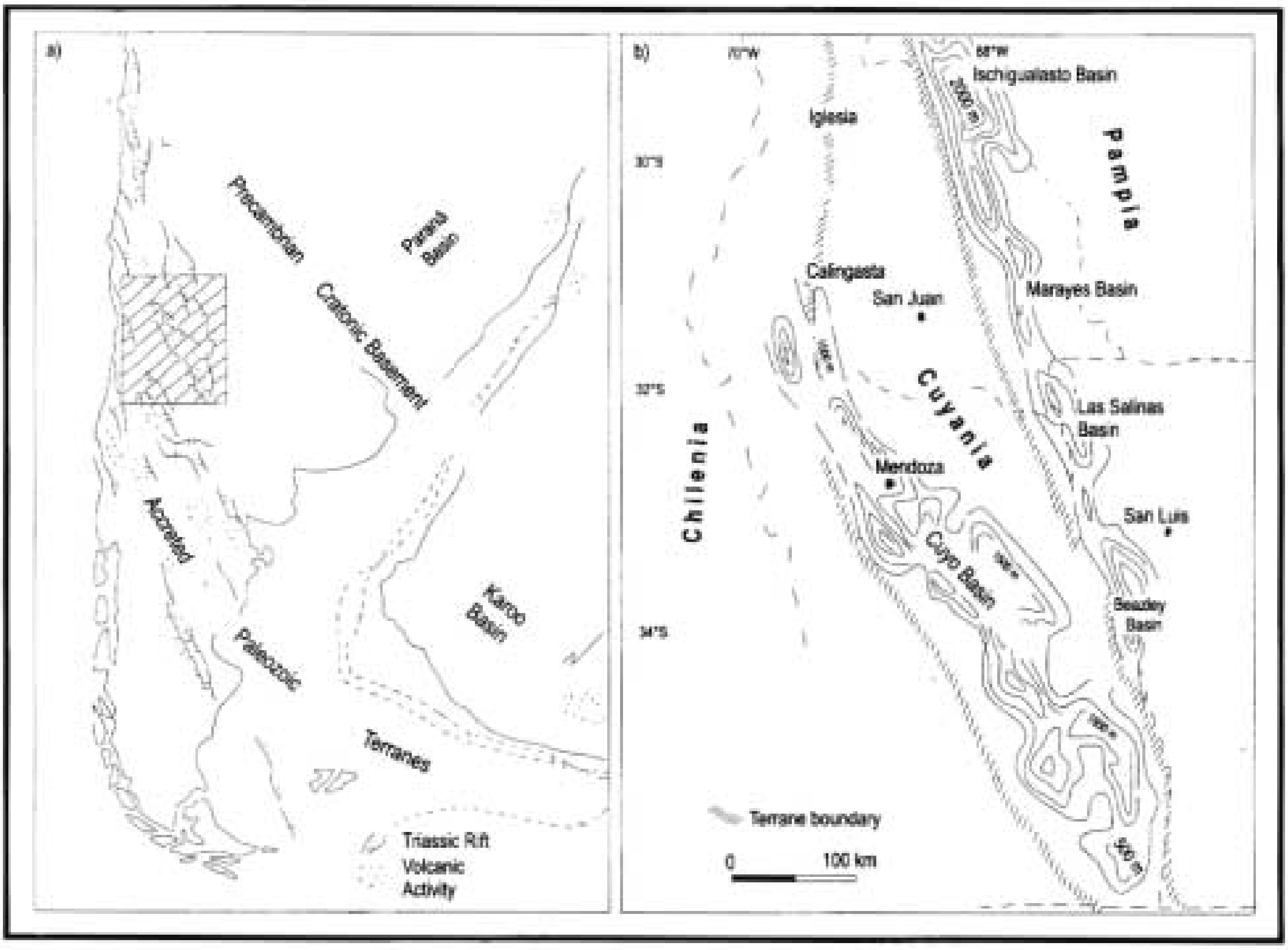


FIGURE 14 - a) Paleogeographic map of the Triassic deposits in western Gondwana with the location of the main rift systems bounding the Precambrian craton. b) Isopach maps of the Triassic rifts located in the inset of Fig. 14 a (after Uliana and Biddle, 1988; Ramos and Kay, 1991; Alvarez and Ramos, 1999).

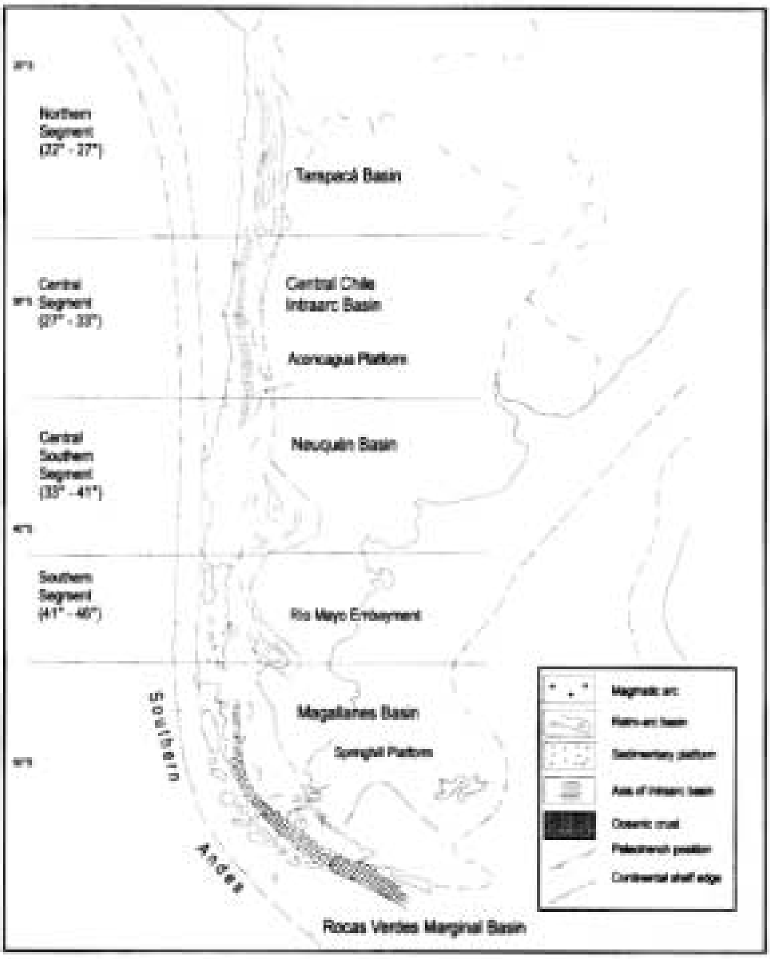


FIGURE 15 - Major segments of the Jurassic-Early Cretaceous arc and retroarc basin systems of the Southern Central Andes (after Mpodozis and Ramos, 1990).

The La Negra Magmatic Arc has been partially eliminated by crustal erosion (Fig. 17) related to subduction processes from Late Mesozoic times (Schweller and Kulm, 1978; Stern, 1991). This magmatic arc was characterized by thousands of metres of tholeiitic basalt and poorly evolved andesite with sporadic marine and continental sediments. The arc developed under an extensional regime that prevailed all along the Pacific margin up to the Early Cretaceous. The sedimentary fill was essentially formed since the Sinemurian, coeval with the onset of the magmatic arc. Total Jurassic-Early Cretaceous thickness exceeded 1500 m in the El Profeta Depocenter (Prinz *et al.*, 1994).

The magmatic arc was bounded to the E by the Atacama Fault, along which there occurred the first strike-slip displacement during Early Cretaceous times (Arabaz, 1971). Naranjo *et al.* (1984) reported mylonitic zones with Early Cretaceous ages and Scheuber *et al.* (1994) demonstrated the link between left-lateral strike-slip displacement and oblique subduction of the Aluk (or Phoenix) Plate from Jurassic times. The obliqueness of the convergence vector exceeded 45°, producing a transtensional regime during Jurassic-Early Cretaceous times.

Jurassic and Early Cretaceous plutons were emplaced in this extensional tectonic regime. Plutons emplaced on ramps within a hinterland-propagating extensional duplex were fed by dykes that transferred magmas from the lower crust and alternated with phases of volcanism (Grocott *et al.*, 1994).

The sea retreated from the retro-arc basin from N to S, remaining in the southern part of the basin up to the Neocomian. A major reorganization took place in Middle Cretaceous times when magmatic activity ceased and the retro-arc region was deformed. As a consequence of this deformation the Protocordillera of Domeyko was uplifted (Mpodozis and Ramos, 1990).

Sedimentological analyses and seismic lines across the Purilactis-Atacama Basin on the eastern slope of the Cordillera de Domeyko show an extensional regime up to Early Cretaceous times (Flint *et al.*, 1993). During the Cretaceous compression, a foreland basin developed according to Muñoz *et al.* (1997), although it should be noted that the stratigraphy of the basin may be much more complex, based on the new radiometric dating (Mpodozis *et al.*, 1999).

Along this segment in the Argentina slope of the Andean Cordillera the rifts of the Salta Group Basin were developed (Salfity, 1982). The Salta Rift comprises a complex suite of extensional troughs of Early Cretaceous-early Tertiary age (Fig. 18), that in the western part were inverted during the Andean deformation in the late Tertiary, being involved in the complex fold and thrust belt (Salfity *et al.*, 1993; Comínguez and Ramos, 1995; Cristallini *et al.*, 1997).

This rift basin is characterized by syn-rift deposits of the Pirgua Subgroup during Early Cretaceous to Campanian times. The post-rift accumulation records the transgression and the final fill of the Balbuena and Santa Bárbara subgroups during Maastrichtian and middle Eocene times (Salfity and Marquillas, 1994). Volcanic and plutonic rocks during the extensional faulting in Late Jurassic and Cretaceous times, as well as alkaline basalt, during reactivation of the rifts in Maastrichtian-Paleocene times, characterize the magmatism associated with the rift (Galliski and Viramonte, 1988).

The evolution of these volcanic rocks records migration of the volcanic centres from the NE-SW trending fault margins of the different basins in the Early Cretaceous, to the axis of maximum subsidence in the Alemanía and the Tres Cruces basins by the end of the Cretaceous. The next volcanic cycles migrated to the easternmost basin, and were restricted to the edge of the Salto-Jujeña High and the axis of the Lomas de Olmedo Basin. The general trend within the Pirgua Subgroup is from peralkaline to subalkaline rocks, indicating greater melting in the source associated with increasing extension and larger volumes of the eruptions. The last cycle is again more alkaline, showing minor reactivation of the rifting (Comínguez and Ramos, 1995).

Central segment: 27°S to 33°S

The Jurassic deposits of this segment are mainly exposed along the High Cordillera (Fig. 19). Marine facies are restricted to the Lower and Middle Jurassic and the Upper Jurassic is represented by continental sequences (Segerstrom, 1959, 1968; Reutter, 1974; Jensen and Vicente, 1977; Cornejo *et al.*, 1984; Riccardi *et al.*, 1992). The classical locality of Manflas is within the region, where Burmeister and Giebel (1861) had first described the Jurassic fauna. Ammonites of this area range from Sinemurian to Bajocian (Riccardi *et al.*, 1992).

An inner Early Cretaceous volcanic arc is preserved in the Coastal Cordillera represented by the Bandurrias Formation, a 3000 m thick andesitic flows, breccia, sandstone and conglomerate (Naranjo, 1981; Pincheira and Thiele, 1982). This magmatic activity is coeval with the development of an outer arc along the international frontier (Ramos and Aguirre-Urreta, 1992). The inner arc between 27°S and 29°30'S contains a thick pile of andesitic rocks interbedded with marine deposits (Moscoso, 1984).

To the S of 29°30'S in the La Serena region there occur two units: the Arqueros Formation and the Quebrada Marquesa Formation. The Arqueros Formation consists of 2000 to 5000 m of interbedded volcanic rocks and marine limestone of Neocomian age. The Quebrada Marquesa Formation consists of conglomerate, sandstone and shale interfingered with andesite. Marine deposits comprise the lower members and continental plant-bearing deposits occur at the top (Moscoso, 1984; Riccardi, 1988). Although traditionally the Quebrada Marquesa Formation was considered to overlie the Arqueros Formation, recent ammonite studies by Mourgues (1999) have shown that both units are partially contemporaneous.

The last two units are exposed farther S between 31°S and 32°S along the Coastal Cordillera in the Illapel region. At the same latitude in the High Cordillera of Ovalle (31°S), and its continuation into Río Tascadero and Paso de las Ojotas (32°S), Rivano and Sepúlveda (1991) have recognized two similar units. The Pelambres Formation, a volcanic sequence interbedded with marine deposits, and the Río Tascadero, a marine sequence with ammonite Neocomian fauna. This unit at 31°15'S passes into Argentina, where it is known as the Calderón Limestone and represents the northernmost outcrop of the Mendoza Group (Neocomian).

Along the Pacific margin Jurassic and Cretaceous plutons are respectively included in the Mincha and Illapel superunits by Rivano *et al.* (1985). There are two belts of typical calc-alkaline rocks, which show a 20 km migration from the Pacific coast at 173 Ma (Mincha Superunit) to the vicinity of Illapel at 104 Ma (Illapel Superunit) (Rivano and Sepúlveda, 1991). These two belts represent the roots of the inner magmatic arc, whereas the Early Cretaceous volcanic rocks correspond to supracrustal sequences coeval with the Illapel Superunit.

Between 32°S and 33°S, the geological setting is similar. Here the Lo Prado and Las Chilcas formations are equivalent to Arqueros and Quebrada Marquesa formations, according to Rivano and Sepúlveda (1991). The eastern belt goes into the Argentina slope of the High Cordillera in the La Ramada and Aconcagua regions. In this area there are exposed complete marine sequences of Early to Middle Jurassic age underlying the Early Cretaceous rocks (Alvarez, 1996). These highly fossiliferous beds include such classical localities as Paso del Espinacito, which has been known since the pioneering work of Stelzner (1873), and the paleontological studies of Göttsche (1878). The Titho-Neocomian Mendoza Group is represented by black shale and shallow marine limestone in a retro-arc setting that defines the Aconcagua carbonate platform (Ramos *et al.*, 1996a).

The Jurassic and Early Cretaceous evolution of this segment is controlled by an extensional regime (Mpodozis and Ramos, 1990). This generalized extension has been interpreted as the result of continental spreading responsible for a thick pile of volcanic rocks bearing an important burial metamorphism (Aguirre *et al.*, 1989). This tectonic scenario has also been explained as an intra-arc setting by Charrier (1984) and Ramos (1985), implying that the limestone interbedded with the Las Chilcas Formation is coeval with the fossiliferous carbonate of the Lo Prado Formation in the W, as well as with the Mendoza Group to the E (Ramos and Aguirre-Urreta, 1992) (Fig. 20). Although the scarce radiometric data relating to the volcanic rocks usually show younger ages than Early Cretaceous, the recent study of Aguirre *et al.* (1999) demonstrated a gap of 25 Ma between the volcanic activity and burial metamorphism using precise Ar/Ar dating on different mineral phases.

By the end of the Early Cretaceous, a major plate reorganization took place, ending with the extensional regime and consequently with the marine intra-arc and retro-arc basins (Mpodozis and Ramos, 1990). The new stress field was responsible for the marine regression, and the development of a continental retro-arc basin. The volcanic front migrated to the E and a series of volcanic and volcaniclastic deposits of widespread distribution were deposited on the western slope of the Andean Cordillera (Rivano and Sepúlveda, 1991). The unconformity between Early and Late Cretaceous volcanic rocks can also be detected by mineralogical breaks in the burial metamorphism (Aguirre *et al.*, 1978).

Central southern segment: 33°S to 41°S

This segment includes the Neuquén Basin, a Mesozoic embayment in the foreland area where complete sequences of Jurassic and Cretaceous rocks are widely exposed (Bracaccini, 1970; Gulisano *et al.*, 1984; Legarreta and Uliana, 1996).

Although N of Valparaíso (33°S) some minor Jurassic outcrops are preserved along the Coastal Cordillera, most of the Jurassic marine deposits of this segment lie along the Principal Cordillera (Riccardi, 1983). It is one of the best known areas, due to more than a century of biostratigraphic and sedimentological studies, where Groeber (1946, 1947a, b) defined his framework of sedimentary cycles (Fig. 21).

Based on Groeber's framework, Legarreta and Uliana (1996) recognized a series of cycles with different tectonic settings. The basin development began as the result of crustal extension and incipient marine encroachment during Late Triassic times through the Hettangian. Recent ammonite studies showed the presence of Late Triassic marine deposits on the eastern slope of the Principal Cordillera (Riccardi and Iglesia Llanos, 1999). A stepwise oceanic flooding and large-scale drowning of the successor rift system took place between Sinemurian and Toarcian times. Under more subdued tectonic drive, construction of a foreland clastic embankment occurred between the Aalenian-Bajocian. The Bathonian coincides with forced regressions and basin shrinkage, whereas during Callovian to Oxfordian times an eustatic rise resulted in a basin-wide episode of carbonate deposition. A Messinian-style desiccation event in the Kimmeridgian with evaporite-eolian deposition was followed by renewed marine flooding, widespread anoxia, and outbuilding of a new carbonate terrace in Tithonian times (Legarreta and Uliana, 1996). A widespread series of marine carbonate beds and shale units was interrupted by two major regressions during the Valanginian and Early Hauterivian. The final Pacific marine withdrawal took place during Barremian-Aptian times.

Foreland deposits, mainly red beds and conglomerate, characterized thick molasse sequences during the Late Cretaceous. The first Atlantic transgression occurred in Maastrichtian-Danian times reaching the foothills of the Principal Cordillera.

The main magmatic arc in this segment was relatively stable during the Mesozoic, and the volcanic front did not migrate eastward as in the above- analyzed segment farther N (Fig. 21). This setting has been interpreted as a combination of the lack of continental erosion along the continental margin and stable Benioff geometry (Ramos, 1988b).

Southern segment: 41°S to 46°S

This segment encompasses the northern Patagonian Cordillera and it consists of thick piles of Mesozoic rocks along the Main Andean Cordillera. The Jurassic basin is mainly developed along the extra-Andean region and is closely linked with the foothills of the Patagonian Cordillera (see Ramos and Aguirre-Urreta, this volume).

The Main Andes at these latitudes consists of the Lago La Plata Formation, a thick sequence of andesitic to dacitic volcanic rocks. This unit partially covers isolated patches of Late Paleozoic sedimentary rocks and Early Jurassic black shale beds (González Bonorino, 1973). Radiometric control in these volcanic rocks indicates a Middle to Late Jurassic age (Haller *et al.*, 1981). Farther S, in the Chilean slope, these volcanic rocks comprise the Ibañez Formation (Heim, 1940). This sequence grades eastwards to more acid and slightly older volcanic rocks widespread in the North Patagonian Massif (Rapela and Pankhurst, 1993; Alric *et al.*, 1996).

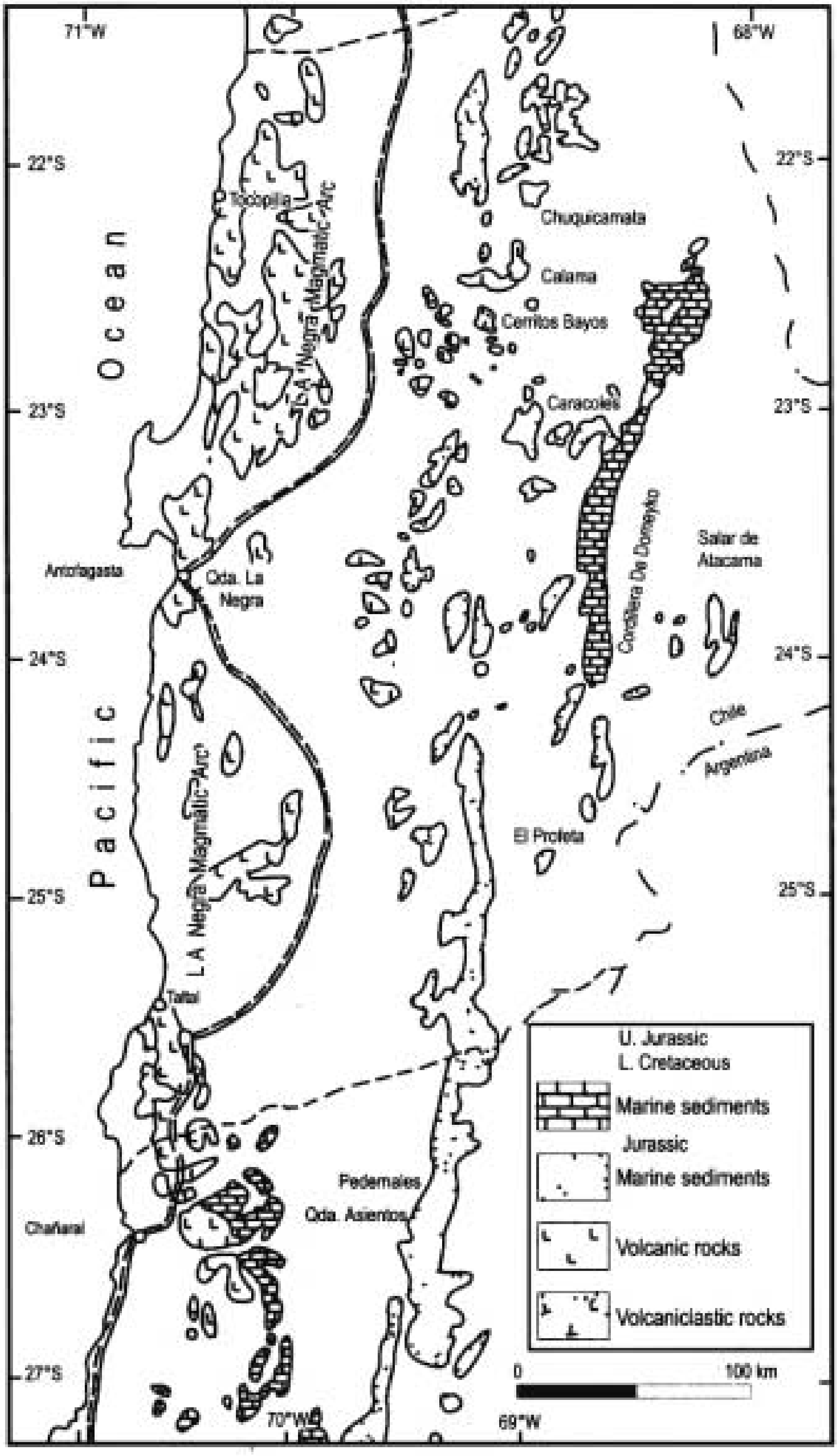
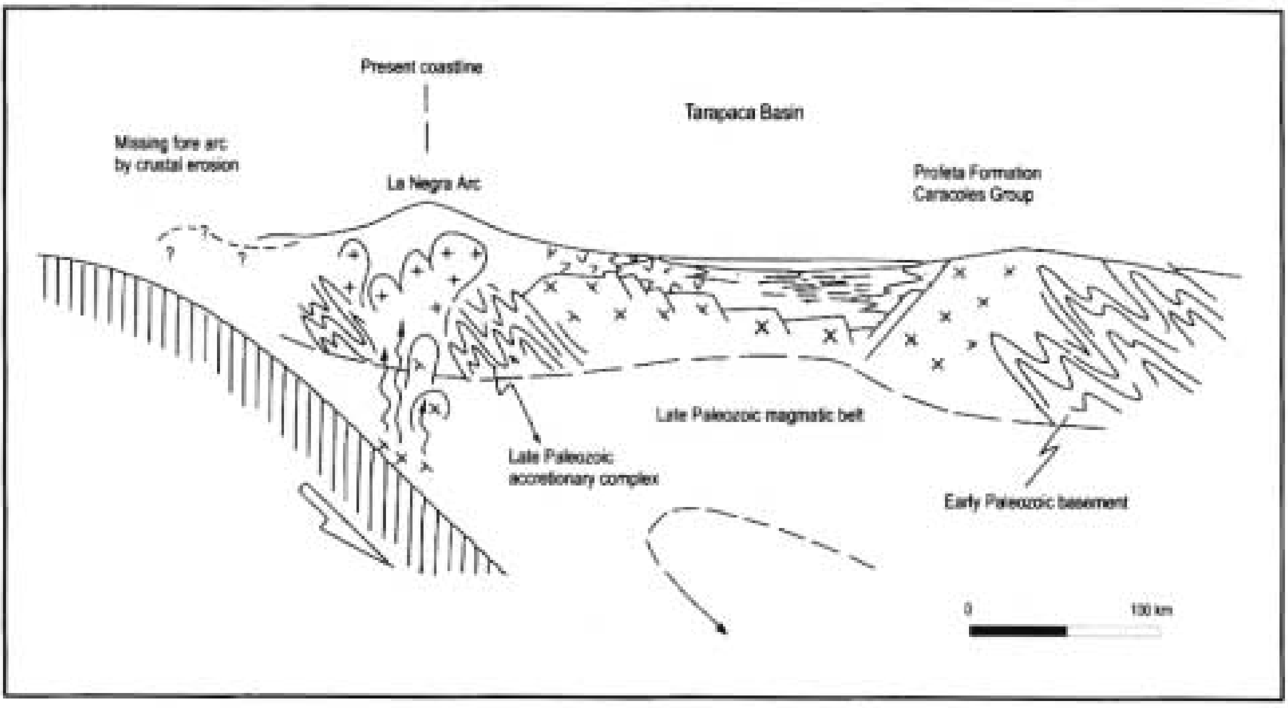


FIGURE 16 - Generalized geologic map of northern Chile with the La Negra Magmatic Arc and the associated Tarapacá Basin (after Riccardi, 1983; Prinz et al., 1994).



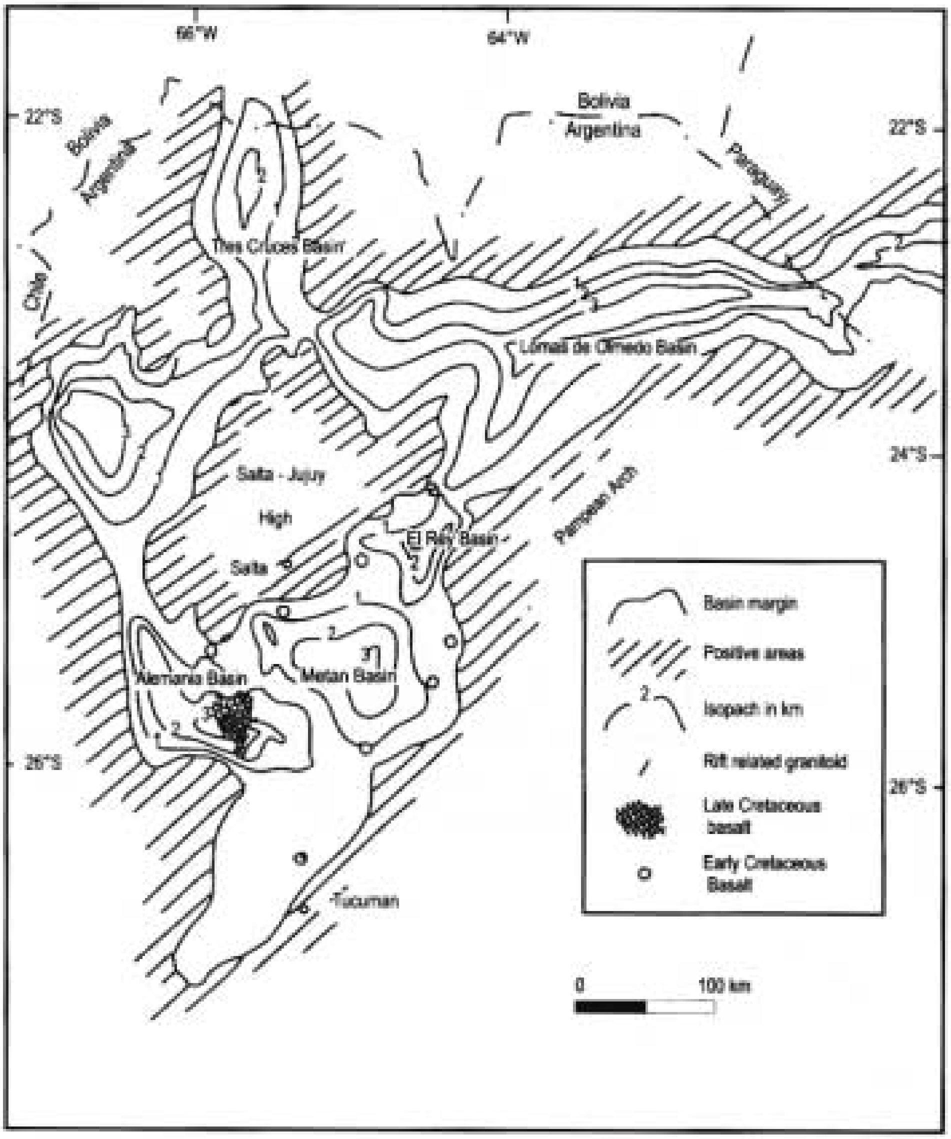


FIGURE 17 - Tectonic setting during Jurassic-Early Cretaceous times in northern Chile (modified after Mpodozis and Ramos, 1990).

FIGURE 18 - Paleogeography of the synrift deposits of the Salta Group Basin with location of the main magmatic activity (after Salfity and Marquillas, 1994).

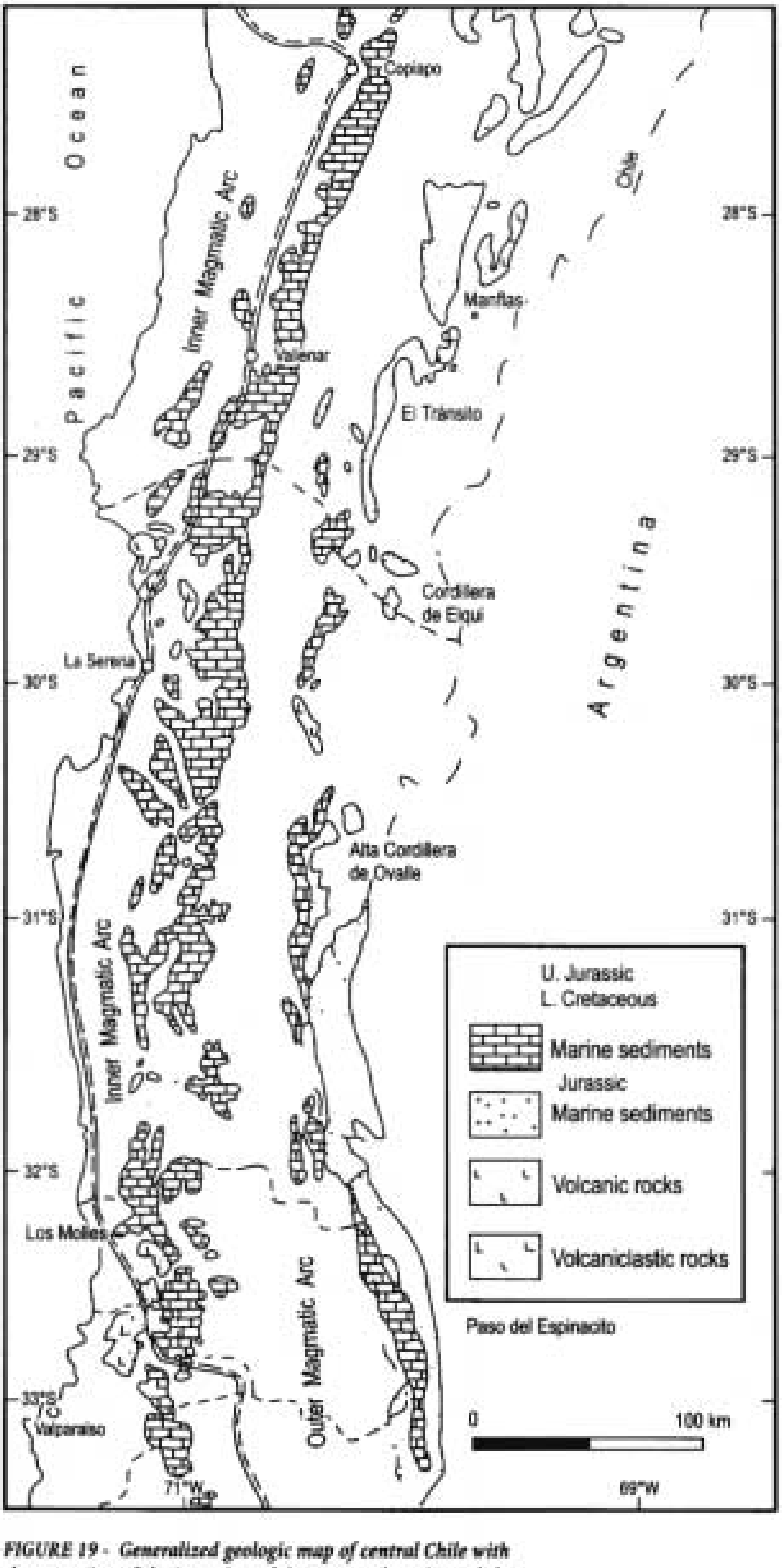


FIGURE 19 - Generalized geologic map of central Chile with the extension of the Jurassic and Cretaceous deposits and the inner an outer magmatic arc (after Riccardi, 1983; Ramos and Aguirre-Urreta, 1992).

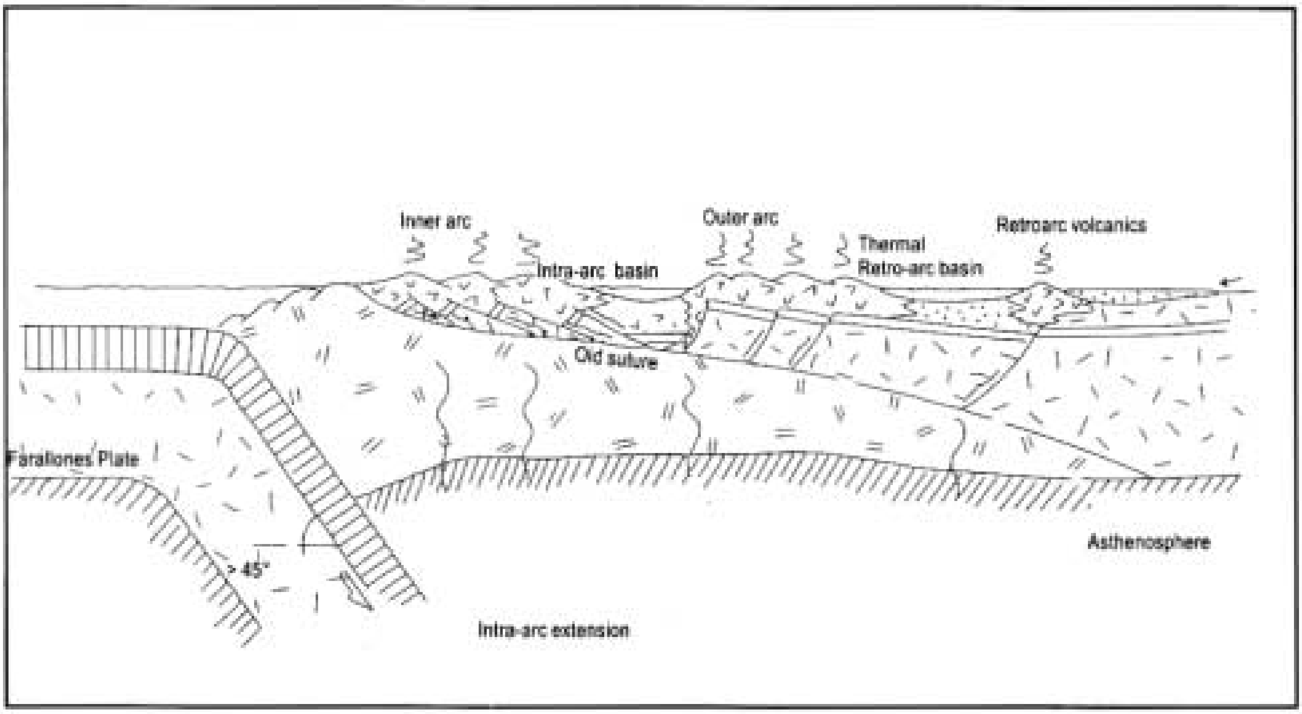
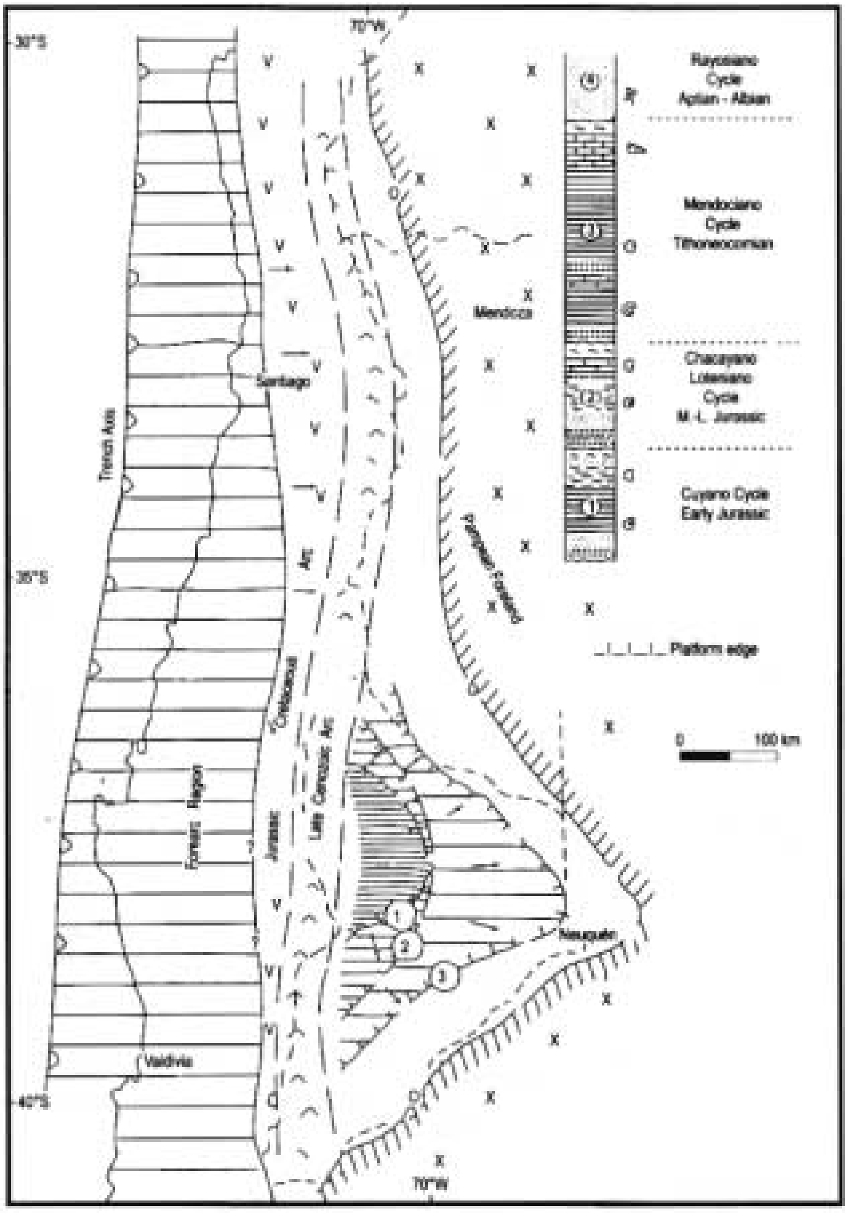


FIGURE 20 - The intra-arc basin at 32-33°S during the Neocomian times (modified after Ramos, 1989).

FIGURE 21 - Different cycles of the Neuquén Embayment showing the stationary position of the magmatic arc during the Mesozoic (after Ramos, 1988b).



A short-lived marine transgression invaded the Patagonian Cordillera from the SW, reaching the latitude of Esquel (43°S) in Neocomian times. Farther S and W, thick sequences of Titho-Neocomian marine deposits are preserved in the Lagos La Plata and Fontana area on the Argentina side, as well as in Palena and Futaleufú on the Chilean slope (Ramos, 1981; Fuenzalida, 1966; De la Cruz *et al.*, 1996).

A second volcanic episode, almost as important as the Jurassic one, took place during Early Cretaceous times. These volcanic rocks known as the Divisadero Group (Heim, 1940) unconformably overlie in the southern part the sedimentary sequence. Dacite, andesitic flows and pyroclastic rocks of late Neocomian to Aptian-Albian age represent them. Two different arcs have been identified in the southern region (45°S) where an inner arc, present intermittently, and consisting of rocks of andesitic to basandesitic composition interfingers with pillow lava and marine sedimentary deposits (Ramos and Palma, 1983). The outer magmatic arc is mainly dacitic to rhyolitic in composition with a dominant pyroclastic facies and interbeds of continental and marine deposits. The outer arc is not present farther S (45°30'S) where the Rio Mayo Embayment occurs (Aguirre-Urreta and Ramos, 1981). This retro-arc embayment is characterized by WNW trending troughs in which thick sequences of marine black shale beds and sandstone units were deposited in an extensional regime. Farther W, on the Chilean side, thick sequences of black shale contain an ammonite fauna of Hauterivian age (Skarmeta, 1978).

The main characteristic of this segment is the widely developed Patagonian Batholith emplaced in multiple episodes. One of the main phases of emplacement took place between 125 and 110 Ma at these latitudes, whereas a second and equally important phase occurred between 90 and 80 Ma (González Díaz, 1982). This batholith consists of typical calc-alkaline granitoid plutons, ranging in composition from tonalite to granite, with a distinctive chemical composition characteristic of subduction related suites (Bartholomew and Tarney, 1984).

The Cenozoic Andean evolution

There are significant differences along strike in the Andean evolution. Three main segments will be described partially matching the Mesozoic segments (Fig. 22). The major Cenozoic segments are controlled by the Benioff geometry of the subduction zone, as first recognized by Barazangi and Isacks (1976) and Jordan *et al.* (1983). This geometry was formed during Neogene times, although there is evidence of previous segmentation prior to the Miocene (Kay *et al.*, 1999).

Northern segment: 22°S to 27°S

This segment encompasses the Western Cordillera, southern Altiplano-Puna high plateau, the Eastern Cordillera and the Subandean and Santa Bárbara systems of Argentina and Chile. The tectonic setting significantly changed from the Late Cretaceous-Paleogene to the Miocene at these latitudes.

Paleogene evolution

The Late Cretaceous-Paleogene evolution of this region was greatly controlled by the Domeyko strike-slip fault system (Mpodozis and Ramos, 1990; Mpodozis *et al.*, 1993; Cornejo *et al.*, 1994). A series of sedimentary troughs formed by extensional listric faults produced half-graben systems where red beds interfinger with basalt flows (Arévalo *et al.*, 1994). These systems were linked to master strike-slip faults, as for example the Hornitos Basin in the Precordillera of Copiapó. During Paleocene to early Eocene times these areas were the focus of an important explosive volcanism, ranging in age between 63 and 55 Ma, where nests of small calderas (5 to 7 km) and stratovolcanoes were emplaced. Occasionally large calderas over 40 km in diameter, such as the Carrizalillo Megacaldera and resurgent associated plutons were formed by the coalescence of older volcanic edifices. This extensional setting extends from Copiapó to the Atacama region and is associated with precious metal mineralization (Cornejo *et al.*, 1994).

The magmatic arc gap during late Eocene-early Oligocene times, was the result of an increase in oblique convergence and low rate of normal subduction (Pardo Casas and Molnar, 1987). The magmatic activity is concentrated along strike-slip systems, as the Domeyko Fault System (Tomlinson and Blanco, 1997), and many giant porphyry copper districts were emplaced at that time (Sillitoe, 1992; Zentili and Maksaev, 1995). Later supergene oxidation and enrichment processes, active since early Oligocene times, and controlled by tectonic uplift pulses favoured mineralization (Sillitoe and McKee, 1996).

Some large basins such as the Arizaro Basin in the Argentina Puna, began to subside, and thick continental and volcaniclastic fill occurred during late Eocene-early Oligocene times (Salfity et al., 1996). These basins were the result of the tectonic compression or transpression mainly developed on the Chilean slope of the Andes, and denote the end of the extension of the Salta Group rift system in northwestern Argentina and adjacent regions of the Chilean border. An angular unconformity separates the extensional rift sequences from the continental deposits of the foreland basins.

Some late Eocene-early Oligocene volcanic rocks are exposed in the western margin of the Argentina Puna between 24°S and 25°S (Zappetini *et al.*, 1997), associated with Paleogene porphyry copper systems such as Taca-Taca.

Neogene evolution

The spatial and temporal distribution and the chemistry of late Oligocene to Recent volcanic rocks of this segment can be broadly explained by changes in the dip of the subducting Nazca Plate and the thickness of the lithosphere mantle and crust beneath the Puna Plateau (Coira *et al.*, 1994). Temporal changes in lithosphere and crustal thickness were tracked by using REE elements as guides to pressure-sensitive residual minerals and source melting percentages (Kay *et al.*, 1999). Thin lithosphere was tracked by large dacitic ignimbrite eruptions and the distribution of mafic lava. The data suggest that the southern Puna has been in an intermediate position between a steepening slab to the N and a shallowing slab to the S throughout the late Tertiary, a view that is in agreement with the previous work of Isacks (1988). Most Puna volcanic rocks are associated with stratovolcanoes and dome complexes, caldera complexes, or monogenetic cones (Coira *et al.*, 1993).

The main volcanic chains in the Puna are stratovolcanoes composed of thick sequences of andesitic to dacitic lava and pyroclastic flows cut by dacitic to rhyodacitic domes. Hot avalanches and small ignimbritic deposits are frequently related to dome structures. Among the Puna stratovolcanoes are the highest (Ojos del Salado, 27°07'S, 68°33'W; 6887 m) and second highest (Llullaillaco, 24°43'S, 68°32'W; 6723 m) active volcanic centers on Earth. The eruption of these and other Puna volcanoes at high elevations has an important effect on associated volcanic deposits as the eruptive columns from both small and large eruptions (typically 5000 to 15 000 to 17 000 m) are influenced by stratosphere winds prevailing from W to E that can have speeds exceeding 150 km/h. This situation produces coarse-grained, proximal pyroclastic fall deposits near the centre, sparse deposits at intermediate distances (Puna region), and very fine to fine-grained distal deposits E of the Puna (Eastern Cordillera, Subandean Belt, Chaco Plain). Deposits from a recent eruption of the Lascar Volcano (23°22'S, 68°34'W) show this pattern (Coira *et al.*, 1994).

Among the deposits of large stratovolcanoes are catastrophic debris avalanches caused by partial collapse of the volcanic edifice (Francis *et al.*, 1985; Francis and Ramírez, 1985). These avalanches are associated with viscous dacitic lava and steep cones, and can be triggered by seismic activity (Francis and Wells, 1988). Several of these avalanche deposits cover more than 100 km^2 and have volumes > 10 km^3. The collapse of the cones of large volcanoes could be considered as a normal stage in their evolution (Coira *et al.*, 1993).

Voluminous ignimbrite sheets that form large volcanic plateaux constitute the dominant late Miocene to early Pleistocene volcanic deposits in the Puna. These ignimbrite flows cover over 500 000 km^2, forming one of the largest young ignimbrite provinces on Earth. Most of these sheets were erupted from huge calderas associated with complex, long-lived volcanic centres occurring parallel to the modern main arc or in the transverse chains that cross the plateau (Fig. 23). The scale of many of these ignimbrite sheets is such that their vents were not recognized until the use of satellite (Landsat and MSS) imagery became widespread (Baker and Francis, 1978; Sparks *et al.*, 1985; Ort *et al.*, 1989; Gardeweg and Ramírez, 1987).

Large ignimbrite sheets of this segment are considered to have erupted from large homogeneous magma chambers. High-volume Puna ignimbrite sheets are characterized by their homogeneity, their bulk dacitic composition, and their phenocryst-rich nature (up to 40 - 50% crystals). There is no evidence for a crystal-poor, volatile-rich cap to the magma chamber. Low volatile contents (see Ort, 1992) and interaction with stratosphere winds lead to thick ignimbrite sheets and rare Plinian deposits (Coira *et al.*, 1994).

Puna ignimbrite sequences from smaller centres are more variable in composition and texture, and are considered to be associated with smaller, zoned magma chambers (Hawkesworth *et al.*, 1982; de Silva, 1991).

Young monogenetic cones that produced Strombolian-type eruptions of mafic andesite are concentrated in the southern Puna (Viramonte *et al.*, 1984; Knox *et al.*, 1989; Fielding, 1989). Volcanic features include lava flows, small spatter cones, bomb fields, and proximal ash fall deposits. In a very few cases (*i.e.*, Salar del Hombre Muerto) interaction with lakes caused base surge deposits. These small volume centres are generally associated with extensional or strike-slip NNW and N trending faults.

Magmatic distribution through time and space shows that the Neogene volcanism started at about 26 Ma, and migrated from 17 to 12 Ma to the foreland, developing a flat-subduction zone in northern Puna (22°S-24°S). Major ignimbrite centres migrated to the W from 12 to 3 Ma, followed by the development of the large ignimbrite center of Cerro Galán in late Pliocene as well as mafic volcanic flows with intraplate and back-arc calc-alkaline chemistry (Coira *et al.*, 1994). This magmatic pattern has been interpreted by Kay *et al.* (1999) as evidence of shallowing and steepening of the subduction zone form the early Miocene to the present as shown in Figure 23.

These volcanic processes were coeval with the development of thrusting and consequent tectonic load and flexural subsidence that controlled thick foreland basins in the Eastern Cordillera and the Subandean areas. These were partially cannibalized during Pliocene times to form the present Cenozoic basins at the foothills of the Subandean System (Salfity *et al.*, 1996; Ramos, 1999b).

The present structure of this segment is characterized by a thin-skinned fold and thrust belt in the Subandean ranges, and a thick-skinned belt in the Eastern Cordillera between 22°S and 24°S. Farther S, the Santa Bárbara System, produced by tectonic inversion of Cretaceous normal faults, characterizes the eastern foothills of the Andes between 24°S and 27°S (Kley *et al.*, 1999; Cristallini *et al.*, 1997).

Central segment: 27°S to 33°S

This central segment of the Andes is the highest non-collisional orogenic belt of the world. It reaches elevations near 7 km (the highest mountains of the Western Hemisphere). These mountains are in an area of no present volcanic activity, and therefore the late Cenozoic shortening is directly related to the present uplift and convergence rates.

Several provinces have been defined across the Andes at these latitudes. The Coastal Cordillera with Mesozoic deposits and plutonic and volcanic rocks; the Principal Cordillera with the Mesozoic marine deposits and Cenozoic volcanic rocks, uplifted during Cenozoic times; the Frontal Cordillera, Late Paleozoic rocks uplifted during late Cenozoic times; the Precordillera, Early Paleozoic rocks uplifted during the latest Cenozoic; and the Sierras Pampeanas, Precambrian-Early Paleozoic basement blocks uplifted during the Andean Orogeny (Ramos, 1988b).

Paleogene evolution

The early Tertiary evolution is somewhat obliterated by the intense Neogene deformation. Some relicts of the Eocene uplift and shortening have been preserved in northern Valle del Cura (29°S-30°S), on the Argentina side of the Andes. There, the syn-orogenic deposits began with continental red beds intercalated with ash-fall tuff that range in age from middle to late Eocene (45 to 34 Ma; Limarino *et al.*, 1999b).

These tuff beds indicate active volcanism on the Chilean slope, where the volcanic complexes of Cerro Blanco and Altos del Yaretal (29°30'S-30°S) of middle late Eocene age (45 - 44 Ma) have been identified. They consist of pyroclastic flows and lava of andesitic to rhyodacitic composition (Emparan and Pineda, 1999).

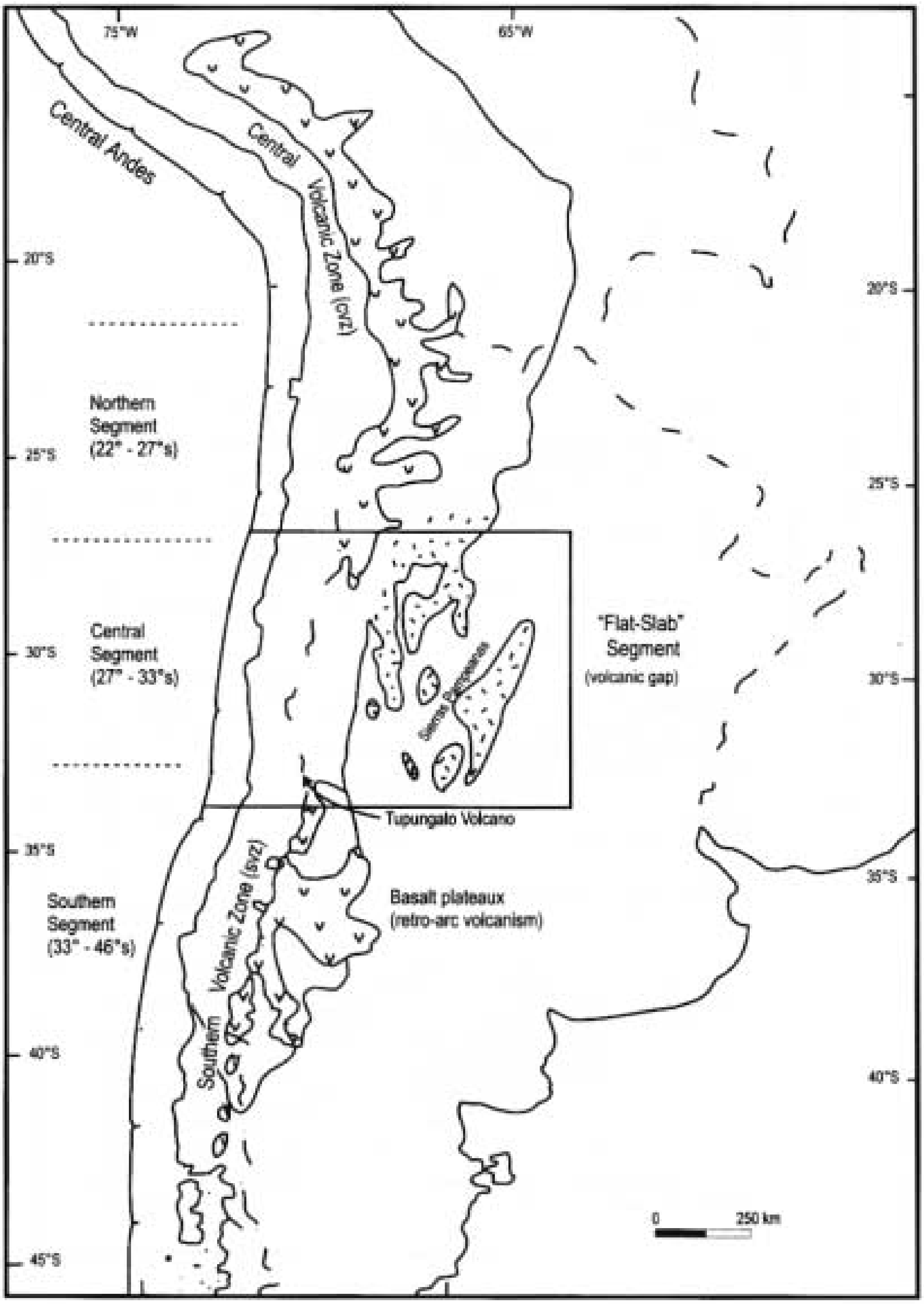


FIGURE 22 - Major segments of the Southern Central Andes related to the Nazca Plate segmentation (based on Jordan et al., 1983).

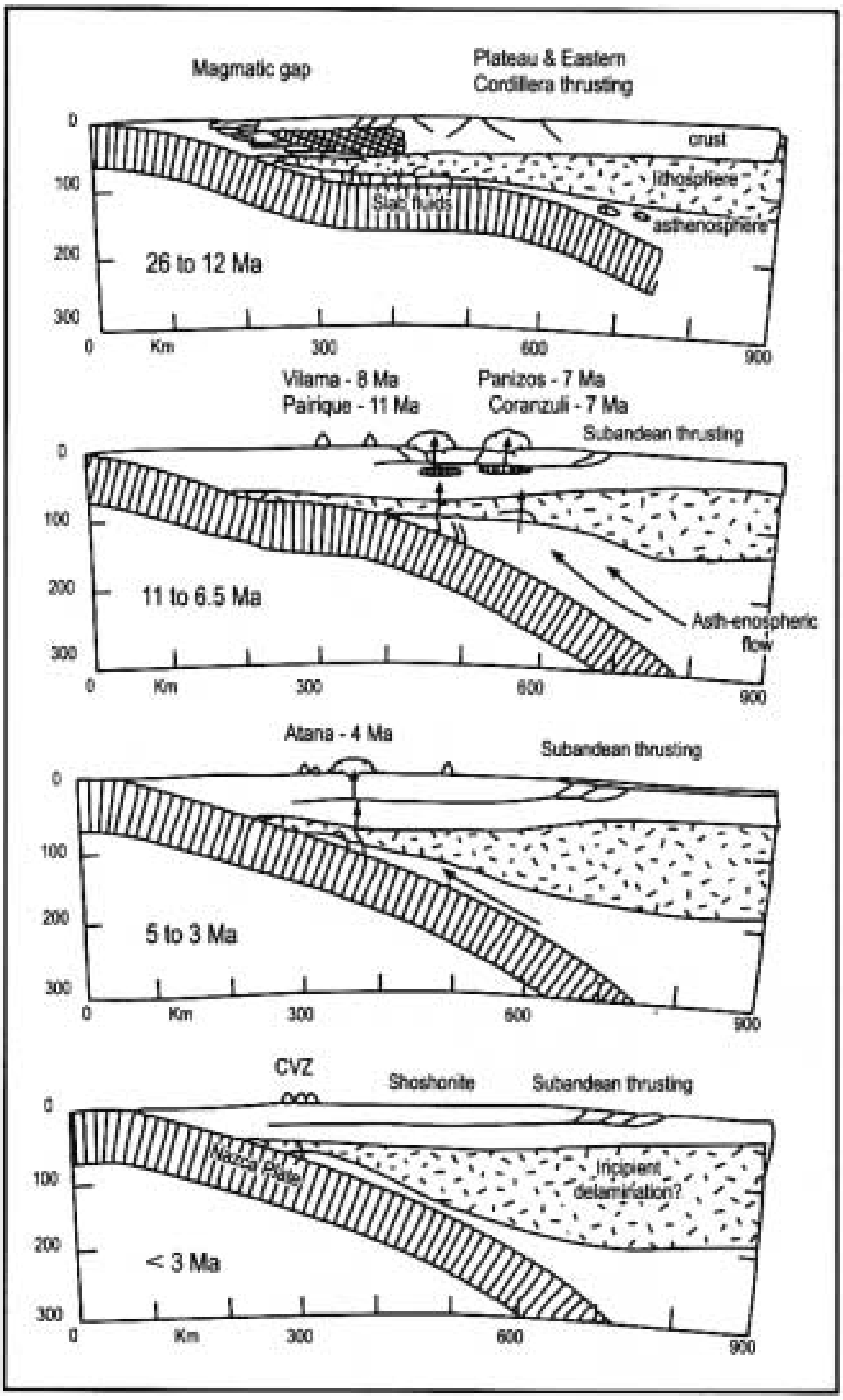
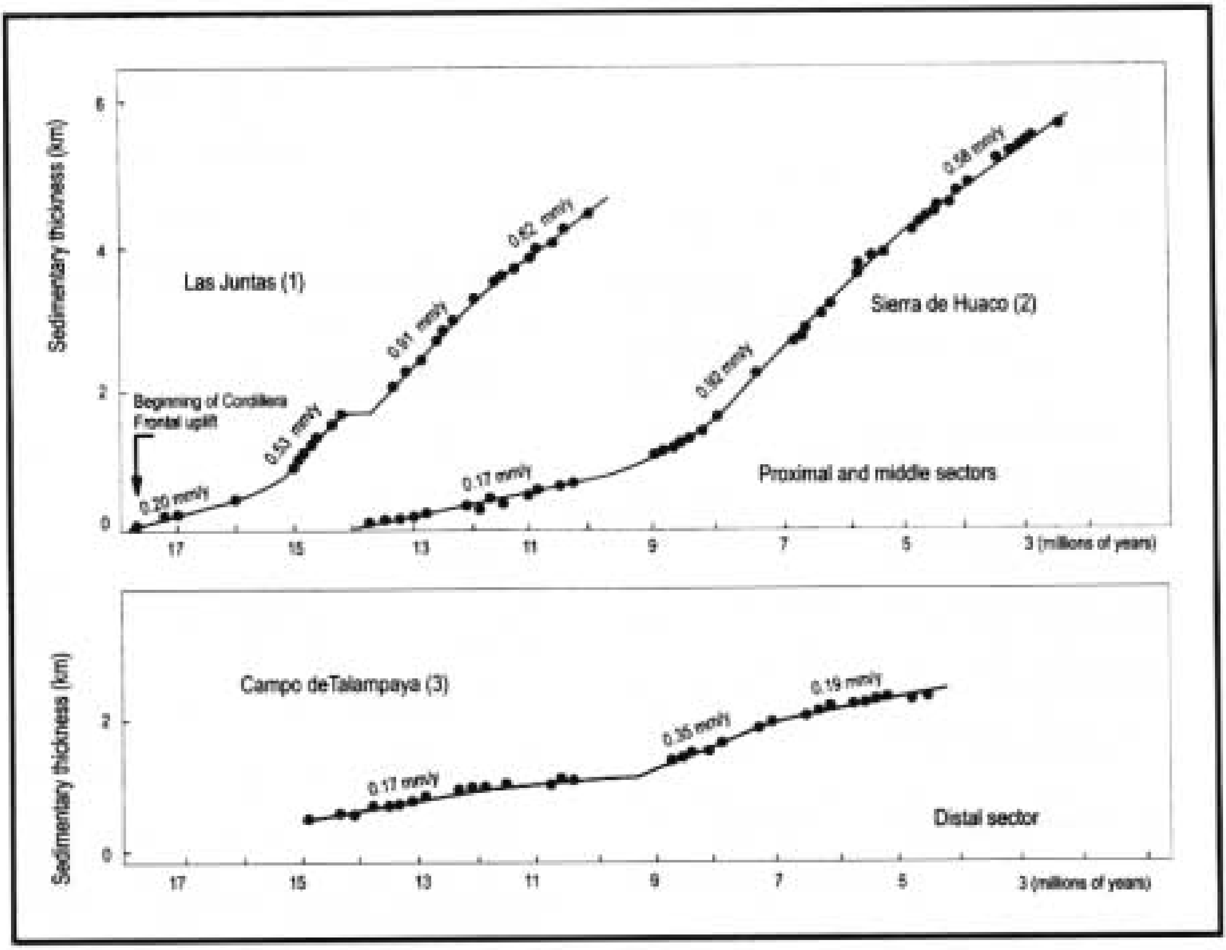


FIGURE 23 - Diagrammatic cross section of the Puna near 23°S showing the processes associated with shallowing and steepening of the subduction zone (after Kay et al., 1999).

FIGURE 24 - Subsidence-sedimentation rates from different stratigraphic sections from foreland basins located about 30°S (after Reynolds, 1987 (1); Johnson et al., 1986 (2); Malizia et al., 1995a and b (3). Proximal and distal sectors refer to the orogenic front.



There is also some evidence of Paleogene uplift in the Sierras Pampeanas as determined by fission track analyses and some Eocene molasse deposits (Ramos, 1999b).

Neogene evolution

The Neogene volcanic rocks permit one to address one of the first order features of the modern Andean subduction history in a segment where a shallow-dipping subducting plate corresponds with a gap in arc volcanism (Barazangi and Isacks, 1976; Fig. 22). The lack of volcanism in this region is a latest Miocene to Recent feature. The distribution and geochemistry of Neogene volcanic rocks across this segment of the Andes (Kay *et al.* 1987a, b; Ramos *et al.*, 1991) along with the evolution of sedimentological and structural features (Jordan *et al.* 1983, 1988; Reynolds, 1987) suggest that the angle of subduction was steeper in the mid-Tertiary and that this has become shallower over the last 18 Ma.

The andesitic lava and rhyolitic pyroclastic flows of the Farellones Formation on the Chilean slope record the activity of the magmatic arc between 20 and 15 Ma (Rivano *et al.*, 1990). The Aconcagua Volcanic Complex (15 to 8 Ma) indicates the migration of the magmatic arc to the Argentina side, that continues farther E.

The shallowing of the subduction zone during the Neogene has been related to a thickening crust. Isotope data combined with trace element data from several regions of the shallow subduction zone permit one to infer differences in crustal thickening patterns in the development of the modern seismic zone (Kay *et al.*, 1991).

Along the Cordillera Principal the thickening is detected prior to 16 Ma, and the end of the subduction-related magmatism occurred at 6 Ma. The Precordillera of San Juan shows magmatic activity between 18 and 6 Ma, whereas in Sierras Pampeanas, volcanism ended at 4.9 Ma in Sierra de Pocho (31°S, Kay and Gordillo, 1994), and at 1.9 Ma in the Sierra del Morro (33°S, Ramos *et al.*, 1991).

The Cenozoic sedimentary history also records the eastward migration of the orogenic front. Thick sequences of continental deposits, mainly coarse-grained conglomerate, unconformably overlie the Mesozoic rocks in the Principal Cordillera. The angular unconformity is clearly seen E of Cerro Aconcagua and W of Cerro Penitentes. These conglomeratic deposits are interpreted as alluvial-fan sediments, interfingering with the volcanic rocks of the Farellones and Aconcagua volcanic complexes (20 to 8 Ma). A minimum age of 8.6 Ma was obtained in the continental deposits based on K/Ar dating of pyroclastic rocks, interbedded in the uppermost section (Ramos *et al.*, 1996 a).

The Tertiary deposits farther to the E of the High Cordillera are represented by distal fluvial facies partially synchronous with the previously described conglomerate beds and volcanic rocks of the Farellones Formation. The Tertiary extra-Andean sequences at these latitudes (30°S-33° S) contain several tuff layers, which attest to the cordilleran volcanic activity at that time (Irigoyen *et al.*, 1995).

Complete magnetostratigraphic studies in the Precordillera of San Juan demonstrate the sequence of thrusting, as well as the out of sequence uplift of the Frontal Cordillera at these latitudes (Jordan *et al.*, 1997). These magnetostratigraphic data permit one to interpret the sedimentation rates and the history of subsidence of the Neogene foreland basins (Fig. 24).

The eastern border of the Frontal Cordillera was uplifted during the late Miocene which in the latest Pliocene gave rise to the present configuration of Frontal Cordillera and Precordillera (Polanski, 1964; Yrigoyen, 1979). The folding and thrusting of these cordilleras produced the subsequent deposition of coarse alluvial fan deposits during the Plio-Pleistocene that form the present bajadas in the Andean foothills (Fig. 25).

Therefore, the Neogene sedimentary facies show a migration of the coarse alluvial fan facies from the inner area of the High Andes between 20 to 8 Ma; to the Uspallata Valley and Cacheuta between 10 to 5 Ma, and to the outer foothills of Precordillera between 2 Ma and the present active front. Even the Plio-Pleistocene fanglomerate cropping out W of the City of Mendoza (Cerro La Gloria) and other younger alluvial fans have been deformed by neotectonic activity, such as the one described by Polanski (1962) in the extra-Andean area of the southern Mendoza region.

The seismic sections of the plains situated to the E of the Precordillera clearly show that the present orogenic front is composed of a set of imbricated overthrusts (Bettini, 1981; Bettini and Turic, 1981). This structural style was corroborated by drilling, and it has somewhat similar characteristics to the previous fronts.

The thrust front on the eastern side of the Precordillera is still active. Intense compressive deformation as seen in the Sierra de Las Peñas, and as inferred from earthquake focal mechanisms and displacements on the alluvial fans, is continuing today (Ramos *et al.*, 1996b).

The Andean structure of the Central Andes is the result of a combination of several tectonic mechanisms. There is a striking coincidence between the increase of plate motion rates, the cessation of magmatism, and the compressive deformation at the orogenic fronts.

Most of the Oligocene was quiescent with low rates of convergence (Pilger, 1981, 1984) coincident with the lack of volcanic activity in the Cordillera. This quiescence ended at 26 Ma when the volcanic activity of the Abanico Formation started (Munizaga and Vicente, 1982). Two interrelated tectonic features demonstrate that change in the geometry of the Benioff Zone followed:

- Eastward migration of the subduction-related magmatic focus from a position about 180 km from the trench at approximately 26 Ma to a point about 500 km from the trench at 6 Ma (Leveratto, 1976; Kay *et al.*, 1987a), and up to the Central Sierras Pampeanas in latest Pliocene-Quaternary times (700 km from the trench, Ramos *et al.*, 1991).

- Geochemical characteristics indicate between 18 Ma and the present a thickening of the continental crust, related to the tectonic stacking of the Andean Cordillera (Kay *et al.*, 1991).

- Eastward shifting of the orogenic deformation during the last 25 Ma: 275 km from the trench at 20 to 8 Ma; 325 km at 10 to 5 Ma, and 350 to 365 km at 2 Ma. This implies an average shifting rate of 2.5 mm/year of the orogenic front over the last 25 million years, although the shifting was probably sporadic.

The causes of the change in the geometry of the Benioff Zone and segmentation are probably complex and multifaceted. The break-up of the Farellones Plate into the Cocos and Nazca plates, which occurred at *c.* 26 - 25 Ma, seems to mark the beginning of a period of higher convergence rates (Handschumacher, 1976). This age coincides with the initiation of Abanico Magmatism and is a milestone in the geodynamic evolution of the area (Ramos, 1988b). The collision of an aseismic ridge, the Juan Fernández Ridge, as proposed by Pilger (1984), is still the more suitable explanation for the shallowing of the subduction zone geometry.

Southern segment (33°S - 46°S)

This segment has a normal subduction geometry and active volcanism. The main geological units are the Coastal, Principal and Frontal Cordilleras, and to the E of the foothills a slightly uplifted range, known as the San Rafael Block (Fig. 25). The first three cordilleras have similar characteristics to the above-described segment, whereas the San Rafael Block consists of a pre-Miocene peneplain of Early and Late Paleozoic rocks that is being uplifted by latest Cenozoic tectonics.

The existence of an active volcanic arc S of Cerro Tupungato (Fig. 26) is related to the normal subduction angle (*c.* 30°) in the southern segment. This correlates with the northern boundary of the unbroken foreland and the truncation of Sierras Pampeanas S of 33°30'S (for discussion, see Isacks *et al.*, 1982; Jordan *et al.*, 1983; Ramos *et al.*, 1991). Several authors have proposed different mechanisms and processes for this first order segmentation (Pilger, 1984; Isacks, 1988).

However, a series of morphostructural changes are not correlated with the geometry of the Benioff-Wadati Zone. The southern end of the Frontal Cordillera, as well as the change from thin to thick-skinned thrusting in the Principal Cordillera are not controlled by the first order segmentation of the Nazca Plate (Ramos *et al.*, 1996 b).

Paleogene evolution

The Early Tertiary rocks of this region can be subdivided into three distinct segments, northern (33°S -*c.* 37°S), central (*c.* 37°S-40°S), and southern (40°S-46°S).

The studies of Charrier *et al.* (1996) in the northern segment in the Upper Tinguiririca Valley (35°S) of Central Chile, have demonstrated the existence of thick Paleogene sequences. Based on mammal fossils and new radiometric dating, older volcaniclastic sequences assigned to the Early Cretaceous are now identified as latest Eocene-early Oligocene (37.5 - 31.5 Ma). The Coya Machalí Formation is interpreted as the Paleogene magmatic arc on the Chilean slope coeval with some syn-orogenic deposits recognized on the eastern slope of the Andes (Gorroño *et al.*, 1979; Yrigoyen, 1993). Retro-arc magmatism was recorded N of 37°S as a series of andesitic lava flows, pyroclastic flows and dykes (Ramos and Barbieri, 1989; Bettini *et al.*, 1979).

The southern segment has a Paleogene gap in the arc magmatism in the western slope of the cordillera (López Escobar and Vergara, 1998). The Eocene-Oligocene magmatic arc is mainly developed in the Neuquén Cordillera (37°S-39°S), where andesitic and dacitic domes flows and pyroclastic rocks are exposed in the foothills of the

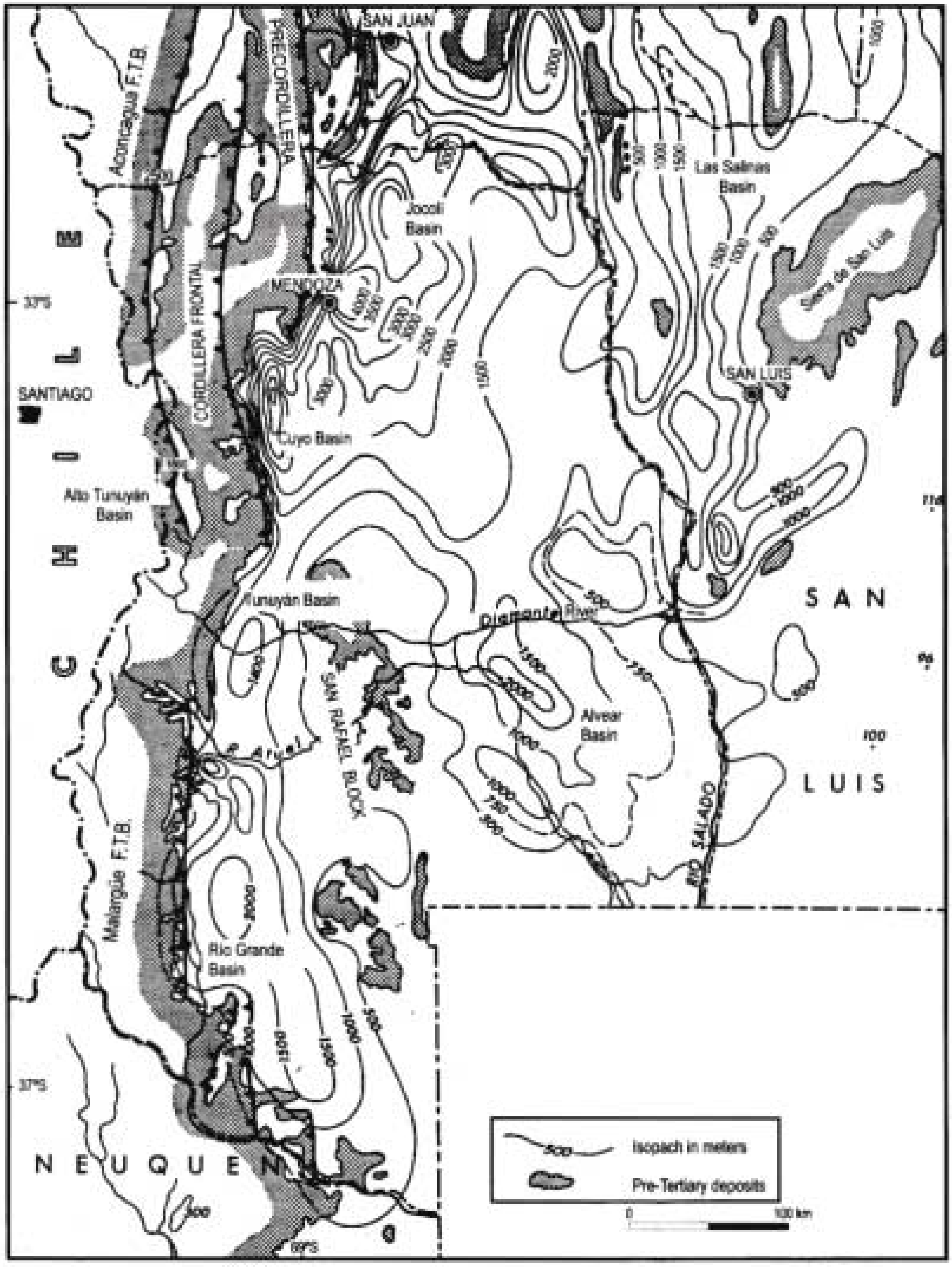


FIGURE 25 - Neogene foreland basins between 33° and 37°S, with location of the San Rafael Block in the foothills (after Yrigoyen, 1993).

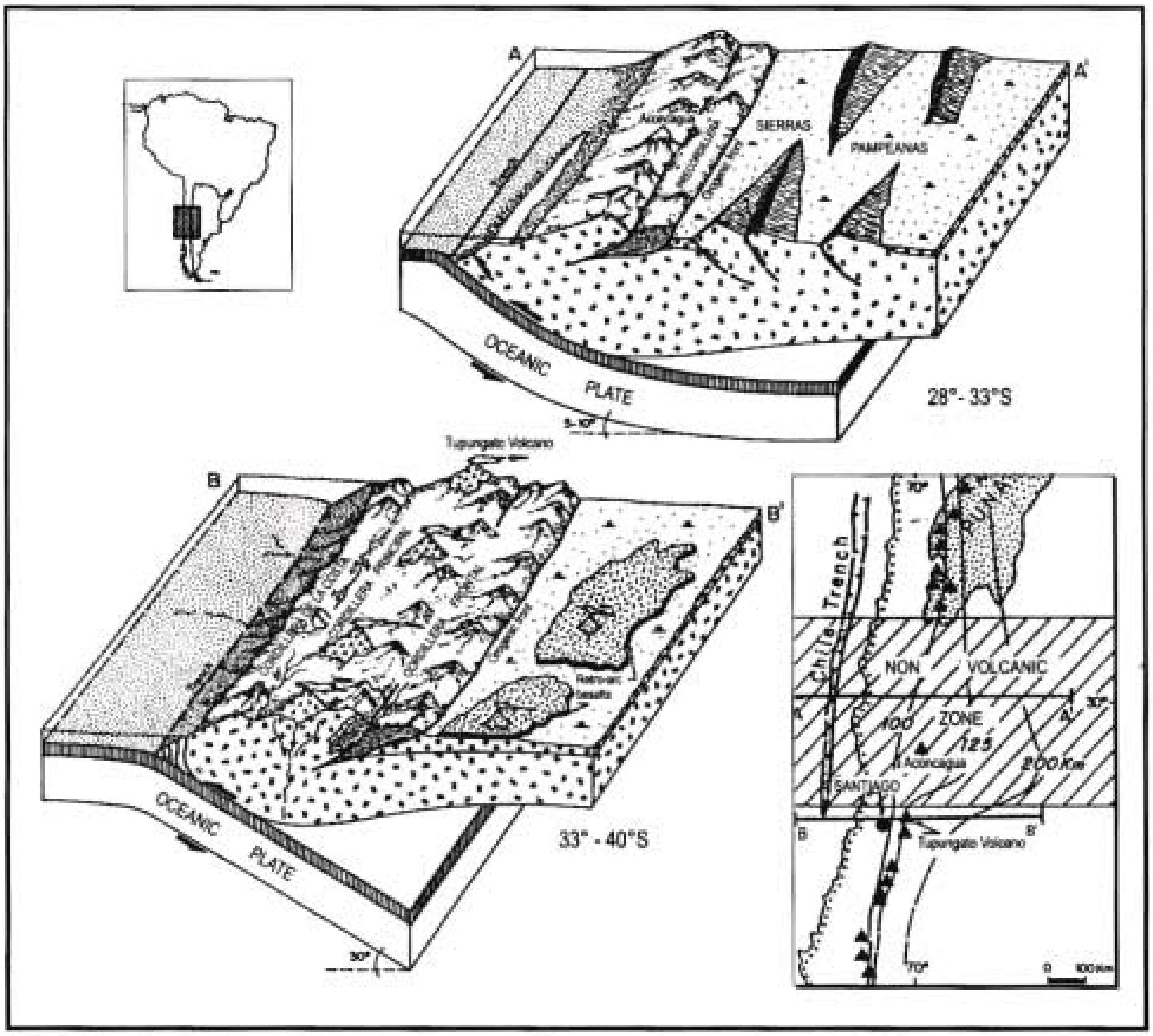


FIGURE 26 - Schematic sections of the different geological units in the subhorizontal and the normal subduction segments. Note that S of Volcán Tupungato there is active magmatic arc and retro-arc basaltic magmatism (after Jordan et al., 1983 and Ramos and Nullo, 1993).

Principal Cordillera (Llambías and Rapela, 1989; Ramos, 1998).

Paleogene volcanism in the southern segment is widespread, with a thick andesitic sequence developed along the main arc in the western belt, and with rocks having a more acid composition in the retro-arc belt (Dalla Salda *et al.*, 1981; Rapela *et al.*, 1988). Between these two belts there developed the Ñirihuau Basin, accounting for an unique Eocene to Oligocene marine transgression derived from the Pacific Ocean (Ramos, 1989; Cazau *et al.*, 1989). Tectonic inversion of the basin was heavily controlled by the oblique component of the subduction at these latitudes (Dalla Salda and Francese, 1987; Diraison *et al.*, 1998).

To the S of 43°30'S there is a new volcanic gap related to a collision of a seismic ridge along the trench, and the volcanism is only related to alkaline plateau basalt (Ramos and Kay, 1992).

The distribution of the magmatic and sedimentary rocks along this segment shows a distinctive control of the geometry of the subduction zone that has changed from Paleogene to the present.

Neogene evolution

The Neogene evolution is highlighted by the important Miocene compression associated with an increase of the subduction rates (Pardo Casas and Molnar, 1987; Somoza, 1998). This Miocene compression was important N of 36°S; slight between 36°S and 40°S; and almost non-existent S of this latitude. The boundary between the Principal and Patagonian Cordilleras at about 38°S-39°S, divide an area of active Neogene compression from a region where the present structure was mainly produced during Late Cretaceous times.

The late Cenozoic magmatic arc has also two distinctive segments with petrological, isotopic, and chemical differences. Rocks of andesitic to dacitic composition predominate between 33°S and 37°S, whereas basaltic rocks are dominant between 37°S and 46°S (López Escobar, 1984; Hickey *et al.*, 1984). These differences, probably controlled by the basement where the arc is emplaced, also match some present changes in the subduction geometry. The subduction zone dips at 30° beneath the magmatic arc S of 33°S, whereas between 35°S and 36°S changes to almost 40°S (López *et al.*, 1997). This steepening of the subduction geometry was identified by Muñoz and Stern (1988) based on the distribution of the Late Tertiary and Quaternary volcanic rocks.

The Neogene foreland basins are well developed N of 36°S (Fig. 25), where thick syn-orogenic deposits are found. The distribution of these sedimentary basins displays different patterns and tectonic styles (Giambiagi, 1999; Pérez *et al.*, 1997), although most of them show a progressive migration to the foreland, beginning in the early Miocene, and still in progress. However, to the S of 36°S latitude the Cenozoic foreland basins show a minor reactivation of the Mesozoic structures during this time.

Farther S at the latitude of Bariloche (41°S-42°S) the Ñirihuau Basin presents a conspicuous early Miocene tectonic inversion, that uplift the previous transtensional structures produced during late Eocene-Oligocene times (Cazau *et al.*, 1989).

Concluding remarks

The Southern Central Andes between 22°S and 46°S denote conspicuous changes along strike in width, elevation, and structural grain. These changes reflected in the Andean topography can be correlated with the geology and paleogeography of the different segments. The main control is related to present subduction parameters and the characteristics of the continental lithosphere, such as subduction vector, age of oceanic crust, thermal regime, absolute plate motions, sublithosphere mantle flows, rheological characteristics of the continental plate, former structural anisotropies, and previous geological history.

The collision of seismic and aseismic oceanic ridges exerted another important control on the extension and evolution of the magmatism. Aseismic segments increase the coupling between the continental and the subducting plate, producing the Highest Andes, almost 7 km high in the Aconcagua region.

The basement anisotropy is another important component, since the rheological properties of the different terranes and their sutures influence the present structure and the reactivation during Andean compression. The present structure of Sierras Pampeanas is a clear example of the control of the previous terrane boundaries in the Mesozoic extension and the late Cenozoic mountain block uplifts.

Therefore, this section of the Andes through its variability and complex history constitutes a key region to test through the geological history the relationships between active subduction and mountain building.

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REFERENCES

Aceñolaza, F.G. (1992). El Sistema Ordovícico de Latinoamérica. In: *Paleozoico inferior de Ibero-América*, eds. Gutiérrez Marco, J.C., Saavedra, J., Rábano, I., pp. 85-118. Universidad de Extremadura, Mérida.

Aguirre, L. (1992). Metamorphic pattern of the Cretaceous Celica Formation, SW Ecuador, and its geodynamic implications. *Tectonophysics*, **205**, 223-237.

Aguirre, L. and Offler, R. (1985). Burial metamorphism in the western Peruvian trough: its relation to Andean magmatism and tectonics. In: *Magmatism at the Plate edge: the Peruvian Andes*, eds. Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D., pp. 59-71. Blackie Halstead Press, London.

Aguirre, L., Levi, B. and Nyström, J.O. (1989). The link between metamorphism, volcanism and geotectonic setting during the evolution of the Andes. In: *Evolution of metamorphic belts*, eds. Daly, J.S., Cliff, R.A., Yardley, B.W.D. *Geological Society of London Special Publication*, **43**, 223-232.

Alemán, A. and Marksteiner, R.M. (1997). Petroleum Systems and Structural Styles in the Santiago Fold and Thrust Belt: A salt Related Orogenic Belt. *VI° Simposio Bolivariano, Bogotá, Petroleum Exploration in the Subandean Basins*, **2**, 171-186.

Aleman, A. and Ramos, V.A. (2000). The Northern Andes. In: Tectonic evolution of South America, eds. Milani, E.J. and Thomaz Filho, A. *31° International Geological Congress* (Rio de Janeiro), this volume.

Allmendinger, R., Jordan, T., Palma, M. and Ramos, V.A. (1982). Perfil estructural en la Puna Catamarqueña (25-27°S), Argentina. *V°Congreso Latinoamericano de Geología (Buenos Aires), Actas*, **1**, 499-518.

Allmendinger, R.W., Figueroa, D., Snyder, D., Beer, J., Mpodozis, C. and Isacks, B.L. (1990). Foreland shortening and crustal balancing in the Andes at 30° S latitude. *Tectonics*, 9(4), 789-809.

Allmendinger, R.W., Jordan, T.E., Kay, S.M. and Isacks, B.L. (1997). The evolution of the Altiplano-Puna Plateau of the Central Andes. *Annual Reviews of Earth and Planetary Sciences*, **25**, 139-174.

Alvarez-Marrón, J., McClay, K.R., Harambour, S., Rojas, L. and Skarmeta, J. (1993). Geometry and evolution of the frontal part of the Magallanes foreland thrust and fold belt (Vicuña area), Tierra del Fuego, southern Chile. *American Association of Petroleum Geologists Bulletin*, 77(11), 1904-1921.

Aspden, J.A. and Litherland, M. (1992). The geology and Mesozoic collisional history of the Cordillera Real, Ecuador. *Tectonophysics*, **205**, 187-204.

Aspden, J.A. and McCourt, W.J. (1986). Mesozoic oceanic terrane in the Central Andes of Colombia. *Geology*, **14**, 415-418.

Astini, R.A. and Thomas, W.A. (1999). Origin and evolution of the Precordillera terrane of western Argentina: A drifted Laurentian orphan. In: Laurentia-Gondwana connections before Pangea, eds. Ramos, V.A. and Keppie, D. *Geological Society of America Special Paper*, **336**, 1-20.

Atherton, M.P. and Sanderson, L.M. (1987). The Cordillera Blanca Batholith: a study of granite intrusions and the relation of crustal thickening to peraluminosity. *Geologische Rundschau*, **76**, 213-232.

Atherton, M.P. and Webb, S. (1989). Volcanic facies, structure and geochemistry of the marginal basin rocks of Central Perú. *Journal of South American Earth Sciences*, **2**, 241-261.

Atherton, M.P., Pitcher, W.S. and Warden, V. (1983). The Mesozoic Marginal basin of Central Perú. *Nature*, **305**, 303-305.

Auboin, J.A., Borrello, A.V., Cecione, G., Charrier, R., Chotin, P., Frutos, J., Thiele, R. and Vicente, J.C. (1973). Esquisse paleogeographique et structurale des Andes Meridionales. *Revue* *de Géographie Physique et de Geologie Dynamique*, **XV**(1-2), 11-71.

Avila Salinas, W. (1991). Petrologic and Cenozoic evolution of the Cenozoic volcanism in the Bolivian Western Andes. In: Andean Magmatism and its Tectonic Setting, eds. Harmon, R.S. and Rapela, C.W. *Geological Society of America Special Paper*, **265**, 245-257.

Baby, P., Heráil, G., Salinas, R. and Sempere, T. (1992). Geometric and cinematic evolution of passive roof duplexes deduced from cross section balancing: example from the foreland thrust system of the southern Bolivian Subandean zone. *Tectonics*, **11**(3), 523-536.

Baby, P., Moretti, I., Guillier, B., Limachi, R. Méndez, E. Oller, J. and Specht, M. (1995). Petroleum system of the northern and central Bolivian Subandean zone. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 445-458.

Bahlburg, H. (1990). The Ordovician basin in the Puna of NW Argentina and N Chile: geodynamic evolution from back-arc to foreland basin. *Geotektonische Forschungen*, **75**, 1-107.

Bahlburg, H. and Hervé, F. (1997). Geodynamic evolution and tectonostratigraphic terranes of Northwestern Argentina and Northern Chile. *Geological Society of America Bulletin*, **109**(7), 869-884.

Barazangi, M. and Isacks, B. (1976). Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America. *Geology*, **4**, 686-692.

Barrero, D. (1979). Geology of the Western Cordillera, west of Buga and Roldanillo, Colombia. *Publicaciones Geológicas Especiales del Ingeominas*, **4**, 1-75.

Bastías, H.E., Uliarte, E., Paredes, J.D., Sánches, A., Bastías, J.A., Ruzycki, L. and Perucca, P. (1990). Neotectónica de la provincia de San Juan. In: Geología y Recursos Naturales de la provincia de San Juan, ed. Bordonaro, O. *XI° Congreso Geológico Argentino, San Juan, Relatorio*, 228-244.

Beck Jr., M.E. (1998). On the mechanism of crustal block rotations in the Central Andes. *Tectonophysics*, **299**, 75-92.

Beck, S.L., Zandt, G., Myers, S.C., Wallace, T.C., Silver, P.G. and Drake, L. (1996). Crustal-thickness variations in the central Andes. *Geology*, **24**(5), 407-410.

Beer, J.A., Allmendinger, R.W., Figueroa, D.E. and Jordan, T. (1990). Seismic stratigraphy of a Neogene Piggyback basin, Argentina. *American Association of Petroleum Geologist Bulletin*, 74(8), 1183-1202.

Bellizzia, A. and Pimentel, N. (1994). Terreno Mérida: un cinturón alóctono Herciniano en la Cordillera de Los Andes de Venezuela. *V Simposio Bolivariano Exploración Petrolera en las Cuencas Subandinas, Memoria*, 271-299.

Benavides-Cáceres, V. (1999). Orogenic evoluion of the Peruvian Andes: the Andean cycle. In: Geology and Mineral Deposits of Central Andes, ed. Skinner, B. *Society of Economic Geology Special Publication*, 7, 61-108.

Biddle, K.T., Uliana, M.A., Mitchum, R.M., Fitzgerald, M.G. and Wright, R.C. (1986). The stratigraphic and structural evolution of the central eastern Magallanes basin, southern South America. In: Foreland Basins, eds. Allen, P. and Homewood, P. *International Association of Sedimentologists Special Publication*, **8**, 41-61.

Boltovskoy, E. (1991). Ihering's hypothesis in the light of foraminiferological data. *Lethaia*, **24**, 191-198.

Bordonaro, O. (1992). El Cámbrico de Sudamérica. In: *Paleozoico inferior de Ibero-América*, eds. Gutiérrez Marco, J.C., Saavedra, J. and Rábano, I., pp. 69-84. Universidad de Extremadura, Mérida.

Bosch, M. and Rodríguez, I. (1992). North Venezuelan collisional crustal block: the boundary between the Caribbean and Southamerican plates. *Journal of South American Earth Sciences*, 6(3), 133-144.

Bourgois, J., Toussaint, J.F., Gonzalez, H., Orrego, A., Azema, J., Calle, B., Desmet, A., Murcia, L.A., Pablo, A.P., Parra, E. and Tournon, J. (1985). Les ophiolites des Andes de Colombia: évolution structurale et signification géodynamique. In: *Géodynamique des Caraibes*, ed. Mascle, A., pp. 475-493. Technip, Paris.

Bourgois, J., Toussaint, J. F., Gonzalez, H., Azema, J., Calle, B., Desmet, A., Murcia, L. A., Acevedo, A.P., Parra, E. and Tournon, J. (1987). Geological history of the Cretaceous ophiolitic complexes of northwestern South America (Colombian Andes). *Tectonophysics*, **143**, 307-327.

Brito Neves, B.B. and Cordani, U.G. (1991). Tectonic evolution of South America during the Late Proterozoic. *Precambrian Research*, **53**, 23-40.

Brito Neves, B.B., Campos Neto, M.C. and Fuck, R.A. (1999). From Rodinia to Western Gondwana: an approach to the Brasiliano-Pan African Cycle and orogenic collage. *Episodes*, **22**(3), 155-166.

Burke, K. (1988). Tectonic evolution of the Caribbean. *Annual Reviews Earth Planetary Sciences*, **16**, 201-230.

Cabassi, I., Cabassi, R. and Introcaso, A. (1999). Los Andes peruanos y argentino-chilenos: un estudio cortical comparativo. *XIV° Congreso Geológico Argentino, Salta, Actas*, **I**, 295-297.

Cahill, T. and Isacks, B.L. (1992). Seismicity and the shape of the subducted Nazca plate. *Journal of Geophysical Research*, **97**, 17503-17529.

Campbell, C.J. (1968). The Santa Marta wrench fault of Colombia and its regional setting. *4th Caribbean Geology Conference, Transactions*, 247-261.

Cande, S.C. and Leslie, R.B. (1986). Late Cenozoic Tectonic of the Southern Chile Trench. *Journal of Geophysical Research*, **91**(B1), 471-496.

Cembrano, J., Hervé, F., and Lavenu, A. (1996). The Liquiñe Ofqui fault zone: a long-lived intra-arc fault system in southern Chile. *Tectonophysics*, **259**, 55-66.

Cembrano, J., Schermer, E., Lavenu, A. and Sanhueza, A. (2000). Contrasting nature of deformation along an intra-arc shear zone, the Liquiñe-Ofqui fault zone, southern Chilean Andes. *Tectonophysics*, **319**(2), 129-149.

Charrier, R. (1979). El Triásico de Chile y regiones adyacentes de Argentina: una reconstrucción paleogeográfica y paleoclimática. *Comunicaciones*, Santiago de Chile, **26**, 1-37.

Charrier, R. (1984). Areas subsidentes en el borde occidental de la cuenca de tras-arco jurásico-cretácica, Cordillera Principal Chilena entre 34° y 34°30'S. *IX° Congreso Geológico Argentino (S.C. Bariloche), Actas*, **2**, 107-124.

Chmielowski, J., Zandt, G. and Haberland, C. (1999). The Central ANdean Altiplano-Puna magma body. *Geophysical Research Letters*, **26**(6), 783-786.

Coira, B.L., Davidson, J.D., Mpodozis, C. and Ramos, V.A. (1982). Tectonic and magmatic evolution of the Andes of Northern Argentina and Chile. *Earth Science Review*, **18**(3-4), 303-332.

Coira, B.L., Mahlburg Kay, S. and Viramonte, J. (1993). Upper Cenozoic magmatic evolution of the Argentine Puna: A model for changing subduction geometry. *International Geology Review*, **35**(8), 677-720.

Coira, B.L., Mahlburg Kay, S., Pérez, B., Woll, B., Hanning, M. and Flores, P. (1999). Magmatic sources and tectonic setting of Gondwana margin Ordovician magmas, northern Puna of Argentina and Chile. In: Laurentia Gondwana Connections before Pangea, eds. Ramos, V.A. and Keppie, D. *Geological Society of America Special Paper*, **336**, 145-170.

Colletta, B., Hebrard, F., Letouzey, J., Werner, P. and Rudkiewicz, J-L. (1990). Tectonic style and crustal structure of the Eastern Cordillera (Colombia) from a balanced cross-section. In: *Petroleum and Tectonics in Mobile Belts*, ed. Letouzey, J., pp. 81-

100. Editions Technip.

Colletta, B., Roure, F., de Toni, B., Loureiro, D., Passalacqua, H. and Gou, Y. (1997). Tectonic inheritance, crustal architecture and contrasting structural styles in the Venezuela Andes. *Tectonics*, 16(5), 777-794.

Comínguez, A. and Ramos, V.A. (1995). Geometry and seismic expression of the Cretaceous Salta rift system, northwestern Argentina. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, 62, 325-340.

Cooper, M.A., Addison, F.T., Alvarez, R., Coral, M., Graham, R.H., Hayward, A.B., Howe, S., Martínez, J., Naar, J., Peñas, R., Pulham, A.J. and Taborda, A. (1995). Basin development and tectonic history of the Llanos basin, Eastern Cordillera, and Middle Magdalena valley, Colombia. *American Association of Petroleum Geologists Bulletin*, 79(10), 1421-1443.

Cornejo, P.C., Mpodozis, C., Kay, S.M. and Tomlinson, A.J. (1994). Volcanismo bimodal potásico en regimen extensional del Cretácico superior-Eoceno en la región de El Salvador (26°-27°S), Chile. *7° Congreso Geológico Chileno (Concepción), Actas*, II, 1306-1310.

Cortés, J.M. (1993). El frente de corrimiento de la Cordillera Frontal y el extremo sur del valle de Uspallata, Mendoza. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas*, III, 168-178.

Corvalán, J. (1981). Paleotectonic Map of the Circumpacific region Southeast Quadrant. *Circumpacific Council for Energy and Mineral Resources, A.A.P.G.*, scale 1:10.000.000.

Cosma, L., Lapierre, H., Jaillard, E., Laubacher, G., Bosch, D., Desmet, A., Mamberti, M. and Gabriele, P. (1998). Petrographie and geochimie de la Cordillere Occidentale du Nord de l'Equateur (0°30'): implications tectoniques. *Bulletin Societé Geologique de France*, 169, 739-751.

Costa, C.H. (1992). *Neotectónica del sur de la Sierra de San Luis.* Ph.D. Thesis, Universidad Nacional de San Luis, 390 p. (unpublished)

Cunningham, W.D. (1993). Strike-slip faults in the southernmost Andes and the development of the Patagonian orocline. *Tectonics*, 12(1), 169-186.

Cristallini, E.O. and Ramos, V.A. (2000). Thick-skinned and thin-skinned thrusting in La Ramada Fold and Thrust belt: crustal evolution of the High Andes of San Juan, Argentina (32°SL). *Tectonophysics*, 317, 205-235.

Cristallini, E., Comínguez, A.H. and Ramos, V.A. (1997). Deep structure of the Metan - Guachipas region: Tectonic inversion in northwestern Argentina. *Journal of South American Earth Sciences*, 10(5-6), 403-421.

Dalla Salda, L. and Francese, J. (1987). Las megaestructuras del Macizo y la Cordillera Norpatagónica argentina y la génesis de las cuencas volcano-sedimentarias terciarias. *Revista Geológica de Chile*, 31, 3-13.

Daly, M. (1989). Correlations between Nazca/Farallon plate kinematics and forearc evolution in Ecuador. *Tectonics*, 8(4), 769-790.

Dalziel, I.W.D. (1981). Back-arc extension in the southern Andes: a review and critical reapraisal. *Philosophycal Transactions Royal Society of London*, A 300, 319-335.

Dalziel, I.W.D. (1986). Collision and Cordilleran orogenesis: an Andean perspective. In: Collision Tectonics, eds. Coward, M.P. and Riers, A.C. *Geological Society Special Publication*, 19, 389-404.

Dalziel, I.W.D. (1991). Pacific margins of Laurentia and East-Antarctica-Australia as a conjugate rift pair: evidence and implications for an Eocambrian supercontinent. *Geology*, 19(6), 598-601.

Dalziel, I.W.D. (1994). Precambrian Scotland as a Laurentia-Gondwana link: the origin and significance of cratonic promontories. *Geology*, 22(7), 589-592.

Dalziel, I.W.D. (1986). Collision and Cordilleran orogenesis: an Andean perspective. In: Collision Tectonics, eds. Coward, M.P. and Riers, A.C. *Geological Society Special Publication*, 19, 389-404.

Dalziel, I.W.D. and Brown, R.L. (1989). Tectonic denudation of the Darwin metamorphic core complex in the Andes of Tierra del Fuego, southernmost Chile: implications for Cordilleran orogenesis. *Geology*, 17(8), 699-703.

Dalziel, I.W.D. and Forsythe, R.D. (1985). Andean evolution and the terrane concept. Tectonostratigraphic Terranes of the Circum-Pacific Region. *Earth Sciences Series*, 1, 565-581.

Dalziel, I.W.D, de Wit, M.F. and Palmer, K.F. (1974). Fossil marginal basin in the southern Andes. *Nature*, 250(5464), 291-294.

Davidson, J.P., Harmon, R.S. and Wörner, G. (1993). The source of Central Andean magmas: some considerations. In: Andean Magmatism and its tectonic setting, eds. Harmon, R.S. and Rapela, C.W. *Geological Society of America Special Paper*, 265, 233-243.

Dengo, G. (1983). Mid-America: Tectonic setting for the Pacific margin from southern Mexico to northern Colombia. *Centro de Estudios Geológicos de América Central*, Guatemala, 1-56.

Dengo, C.A. and Covey, M.C. (1993). Structure of the Eastern Cordillera of Colombia: Implications for traps and regional tectonics. *American Association of Petroleum Geologists Bulletin*, 77, 1315-1337.

De Silva, S.L. (1989). Altiplano-Puna volcanic complex of the Central Andes. *Geology*, 17, 1102-1106.

Dewey, J.F. and Bird, J. (1970). Mountain belts and the new global tectonics. *Journal of Geophysical Research*, 75, 2625-2647.

Dewey, J.F. and Lamb, S.H. (1992). Active tectonics of the Andes. *Tectonophysics*, 205, 79-95.

Díaz Martínez, E. (1996). Síntesis estratigráfica y geodinámica del Carbonífero de Bolivia. *XII° Congreso Geológico de Bolivia (Tarija), Memorias*, 1, 355-367.

Dickinson, W. (1974). Plate tectonics and sedimentation. In: Tectonics and sedimentation, ed. Dickinson, W.R. *Society of Economic Paleontologists and Mineralogists Special Publication*, 22, 1-27.

Diraison, M., Cobbold, P.R., Rossello, E.A. and Amos, A.J. (1998). Neogene dextral transpression due to oblique convergence across the Andes of northwestern Patagonia, Argentina. *Journal of South America Earth Sciences*, 11(6), 519-532.

Dorbath, L., Dorbath, C., Jiménez, E. and Rivera, L. (1991). Seismicity and tectonic deformation in the Eastern Cordillera and the sub-Andean zone of Central Perú. *Journal of South American Earth Sciences*, 4(1-2), 13-24.

Duque-Caro, H. (1990). The Choco block in the northwestern corner of South America: structural, tectonostratigraphic, and paleogeographic implications. *Journal of South American Earth Sciences*, 3(1), 71-84.

Du Toit, A.L. (1937). *Our wandering continents. An hypothesis of continental drifting.* Oliver & Boyd, London, 366 p.

Ego, F., Sébrier, M., Lavenu, A., Yepes, H. and Egues, A. (1996). Quaternary stress in the Northern Andes and the restraining bend model for the Ecuadorian Andes. *Tectonophysics*, 259(1-3), 101-116.

Engebretson, D.C., Cox, A. and Gordon, G.R. (1985). Relative plate motions between oceanic and continental plates in the Pacific basin. *Geological Society of America Special Paper*, 206, 1-64.

Erlich, R.N., Macsotay, O., Nederbragt, A.J. and Lorente, M.A. (1999). Palaeoecology, Palaeogeography and depositional environments of Upper Cretaceous rocks of western Venezuela. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 153, 203-238.

Etayo-Serna, F. (1994). *Estudios geológicos del valle superior del Magdalena.* Universidad Nacional de Colombia, 1-266.

Etayo-Serna, F. and Barrero, D. (1983). Mapa de terrenos geológicos de Colombia. *Instituto Nacional de Investigaciones Geológico-Mineras Publicaciones Especiales*, **14**, 235 p.

Farrar, E., Clark, A.H., Kontak, D.J. and Archibal, D.A. (1988). Zongo-San Gabán zone:Eocene foreland boundary of the Central Andean orogen, northwest Bolivia and southeast Peru. *Geology*, **16**(1), 55-58.

Feininger, T. (1982). Glaucophane schists in the Andes of Jambaló, Colombia. *Canadian Mineralogist*, **20**, 41-47.

Feininger, T. (1987). Allochthonous terranes in the Andes of Ecuador and northwestern Peru. *Canadian Journal of Earth Sciences*, **24**, 266-278.

Fernández, A.E. and Jordan, T.E. (1996). Analysis of controls on foreland basin stratigraphy using seismic and outcrop data: application to the Bermejo Basin, Central Andes, Argentina. *XIII° Congreso Geológico Argentino y III° Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas*, **1**, 373-384.

Fernández Seveso, F., Pérez, M.A., Brisson, I.E. and Alvárez, L.A. (1993). Sequence stratigraphy and tectonic analysis of the Paganzo basin, Western Argentina. *XII° International Congress on the Carboniferous-Permian Systems (Buenos Aires), Comptes Rendus*, **2**, 223-260.

Forero-Suarez, A. (1990). The basement of the Eastern Cordillera, Colombia: an allochthonous terrane in northwestern South America. *Journal of South American Earth Sciences*, **3**, 141-151.

Froidevaux, C. and Isacks, B. (1984). The mechanical state of the lithosphere in the Altiplano Puna segment of the Andes. *Earth and Science Planetary Letters*, **71**, 305-314.

Galeazzi, J.S. (1996). Cuenca de Malvinas. In: Geología y Recursos Naturales de la Plataforma Continental Argentina, eds. Ramos, V.A. and Turic, M.A. *XIII° Congreso Geológico Argentino y III° Congreso de Exploración de Hidrocarburos, Mendoza, Relatorio*, **15**, 273-309.

Gansser, A. (1973). Facts and theories on the Andes. *Journal of the Geological Society of London*, **129**, 93-131.

Ghiglione, M., Olivero, E.B. and Ramos, V.A. (1999). Estratos de crecimiento en el frente orogénico palógeno de la Isla Grande de Tierra del Fuego. *II° Congreso Latinoamericano de Sedimentología y VIII° Reunión Argentina de Sedimentología* (resúmenes), Mar del Plata, 81-82.

Giacosa, R. and Heredia, N. (1999). La cuenca de antepaís terciaria asociada a la faja plegada y corrida de los Andes Patagónicos entre los 41° y 42°S, SO de Argentina. *Acta Geológica Hispana*, **32**(1-2), 103-112.

Godoy, E., Yañez, G. and Vera, E. (1999). Inversion of an Oligocene volcano-tectonic basin and uplifting of its superimposed Miocene magmatic arc in the Chilean Central Andes: first seismic and gravity evidences. *Tectonophysics*, **306**(2), 217-336.

Gorring, M.L, Kay, S.M., Zeitler, P.K., Ramos, V.A., Rubiolo, D., Fernández, M.I. and Panza, J.L. (1997). Neogene Patagonian plateau lavas: continental magmas associated with ridge collision at the Chile Triple Junction. *Tectonics*, **16**(1), 1-17.

Grier, M.E. and Dalmeyer, R.D. (1990). Age of the Payogastilla Group: implications for foreland basin development, NW Argentina. *Journal of South American Earth Sciences*, **3**(4), 269-278.

Grocott, J., Brown, M., Dallmeyer, R.D., Taylor, G.K. and Treloar, P.J. (1994). Mechanisms of continental growth in extensional arcs: an example from the Andean plate-boundary zone. *Geology*, **22**(5), 391-394.

Groeber, P. (1938). *Mineralogía y Geología*. Espasa-Calpe Argentina, Buenos Aires, 492 p.

Gutscher, M.A. and Malavieille, J. (1999). Style of upper plate deformation linked to interplate coupling over flat slab segments: evidence from the Andean Margin. In: McClay, K. (ed.) *Thrust Tectonics 99 (London), Abstracts*, 214-217.

Gutscher, M.A., Olivet, J.L., Aslanian, D., Eissen, J.P. and Maury, R. (1999a). The "lost Inca Plateau": cause of flat subduction beneath Peru? *Earth and Planetary Science Letters*, **171**, 335-341.

Gutscher, R., Malavieille, J., Lallemend, S. and Collot, J.Y. (1999b). Tectonic segmentation of the North Andean margin: impact of the Carnegie Ridge Collision. *Earth and Planetary Science Letters*, **168**, 255-270.

Hall, M.L. and Beate, B. (1991). El volcanismo plio-cuaternario en los Andes de Ecuador. *Boletín del Colegio de Geógrafos del Ecuador*, **4**, 5-17.

Hall, M.L., Calle, (1982). Geochronological control for the main tectonic-magmatic events of Ecuador. *Earth Science Reviews*, **18**, 215-239.

Hall, M.L. and Wood, C.A. (1985). Volcano-tectonic segmentation of the Northern Andes. *Geology*, **13**(2), 203-207.

Haq, B.U., Hardenbol, J. and Vail, P.R. (1987). Chronology of fluctuating sea level since the Triassic. *Science*, **235**, 1156-1166.

Harmon, R.S., Barreiro, B., Moorbath, S., Hoefs, J., Francis, P.W., Thorpe, R.S., Deruelle, B., McHugh, J. and Viglino, J.A. (1984). Regional O-, Sr-, and Pb-isotope relationships in late Cenozoic calc-alkaline lavas of the Andean Cordillera. *Journal of the Geological Society*, **141**(5), 803-822.

Harrington, H.J. and Leanza, A.F. (1957). Ordovician trilobites of Argentina. *University of Kansas, Department of Geology, Special Publication*, **1**, 1-276.

Hatcher, R.D., Thomas, W.A., Geiser, P.A., Snoke, A.W., Mosher, S. and Wiltschko, D.V. (1989). Alleghanian orogen. In: The Appalachian-Ouachita orogen in the United States, eds. Hatcher, R.D., Thomas, W.A. and Viele, G.W. *The Geology of North America*, **F-2**, 233-318.

Hawkesworth, C. and Clarke, C. (1994). Partial melting in the Lower crust: new constraints on crustal contamination processes in the Central Andes. In: *Tectonics of the Southern Central Andes, Structure and Evolution of an Active Continental Margin*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 93-101. Berlin, Springer Verlag.

Helwig, J. (1972). Stratigraphy, sedimentation, paleogeography and paleoclimates of Carboniferous ("Gondwana") and Permian of Bolivia. *American Association of Petroleum Geologists Bulletin*, **56**, 108-133.

Henderson, W.G. (1979). Cretaceous to Eocene volcanic arc activity in the Andes of northern Ecuador. *Journal of the Geological Society London*, **136**, 367-378.

Hérail, G., Soler, P., Bonhomme, M. and Lizeca, J.L. (1993). Evolution géodynamique du contact Altiplano-Cordillière Orientale au Nord d'Oruro (Bolivie): implications sur le déroulement de l'orogénése andine. *Compte Rendus Academie Sciences, Paris*, **317**, 512-522.

Hervé, F. (1988). Late Paleozoic subduction and accretion in Southern Chile. Episodes, **11**(3), 183-188.

Hervé, F. (1994). The Southern Andes between 39° and 44°S Latitude: the geological signature of a transpressive tectonic regime related to a magmatic arc. In: *Tectonics of the Southern Central Andes, Structure and Evolution of an Active Continental Margin*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 249-261. Springer Verlag, Berlin.

Hervé, F., Nelson, E., Kawashita, K. and Suárez, M. (1981). New isotopic ages and the timing of orogenic events in the Cordillera Darwin, southernmost Chilean Andes. *Earth Planetary Science Letters*, **55**, 257-265.

Hildreth, W. and Moorbath, S. (1988). Crustal contributions to arc magmatism in the Andes of Central Chile. *Contributions to Mineralogy and Petrology*, **98**, 455-489.

Hoffman, P.F. (1991). Did the breakout of Laurentia turn Gondwanaland inside-out? *Science*, **252**, 1409-1412.

Hoorn, C. (1993). Marine incursions and the influence of Andean tectonics on the depositional history of northwestern Amazonia:

results of a palynological study. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **105**, 267-309.

Hoorn, C., Guerrero, J., Sarmiento, G.A. and Lorente, M.A. (1995). Andean tectonic as cause of changing drainage patterns in Miocene northern South America. *Geology*, **23**(3), 237-240.

Hungerbuehler, D., Steinmann, M., Winkler, W. and Seward, D. (1996). Neogene fission-track stratigraphy of the southern Ecuadorian basins: Implications for regional tectonic history. *3rd International Symposium of Andean Geodynamics, Saint-Malo*, ORSTOM, 387-390.

Introcaso, A., Pacino, M.C. and Fraga, H. (1992). Gravity, isostasy and Andean crustal shortening between latitudes 30 and 35° S. *Tectonophysics*, **205**(1-3), 31-48.

Jaillard, E., Ordoñez, M., Benítez, S., Berrones, G., Jiménez, N., Montenegro, G. and Zambrano, I. (1995). Basin development in an accretionary, oceanic floored forearc setting: southern Coastal Ecuador during Late Cretaceous to Late Eocene times. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez S.R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 615-631.

Jaillard, E., Laubacher, G., Bengtson, P., Dhondt, A.V. and Bulot, L.G. (1999). Stratigraphy and evolution of the Cretaceous forearc Celica. Lancones basin of southern Ecuador. *Journal of South American Earth Sciences*, **12**(1), 51-68.

Jaillard, E., Hérail, G., Díaz Martínez, E. and Baby, P. (2000). Tectonic evolution of the Andes of Ecuador, Peru, Bolivia, and northernmost Chile (1°N - 23°S). In: *Tectonic evolution of South America*, eds. Milani, E.J. and Thomaz Filho, A. 31° International Geological Congress (Rio de Janeiro), this volume.

James, D.E. and Sacks, I.S. (1999). Cenozoic formation of the Central Andes: a geophysical perspective. In: Geology and Mineral Deposits of Central Andes, ed. Skinner, B. *Society of Economic Geology Special Publication*, **7**, 1-25.

Jordan, T.E. (1995). Retroarc foreland and related basins. In: *Tectonics of Sedimentary basins*, eds. Spera, C. and Ingersoll, R.V., pp. 331-362. Blackwell Scientific, Cambridge, Mass.

Jordan, T.E., Isacks, B., Ramos, V.A. and Allmendinger, R.W. (1983). Mountain building in the Central Andes. *Episodes*, **1983**(3), 20-26.

Jordan, T.E., Flemings, P.B. and Beer, J.A. (1988). Dating thrust-fault activity by use of foreland-basin strata. In: *New Perspectives in Basin Analysis*, eds. Kleinspehn, K.L. and Paola, C., pp. 307-330. Springer-Verlag, New York.

Jordan, T., Reynolds, J.H. and Erikson, J.P. (1997). Variability in age of initial shortening and uplift in the Central Andes, 16-33°30'S. In: *Tectonic uplift and Climate change*, ed. Ruddiman, W., pp. 41-61. Plenum Press, New York.

Kay, S.M. and Abbruzzi, J.M. (1996). Magmatic evidence for Neogene lithospheric evolution of the Central Andean "flat-slab" between 30° and 32°S. *Tectonophysics*, **259**, 15-28.

Kay, S.M., Coira, B. and Viramonte, J. (1994). Young mafic backarc volcanic rocks as indicators of continental lithospheric delamination beneath the Argentina Puna Plateau, Central Andes. *Journal of Geophysical Research*, **99**, 24323-24339.

Kay, S.M., Mpodozis, C. and Coira, B. (1999). Neogene magmatism, tectonism, and mineral deposits of the Central Andes (22°S to 33°S). In: Geology and Mineral Deposits of Central Andes, ed. Skinner, B. *Society of Economic Geology, Special Publication*, **7**, 27-59.

Kay, S.M., Ramos, V.A. and Márquez, M. (1993b). Evidence in Cerro Pampa volcanic rocks for slab-melting prior to ridge-trench collision in southern South America. *Journal of Geology*, **101**, 703-714.

Kay, S.M., Ramos, V.A., Mpodozis, C. and Sruoga, P. (1989). Late Paleozoic to Jurassic silicic magmatism at the Gondwanaland margin: analogy to the Middle Proterozoic in North America? *Geology*, **17**(4), 324-328.

Kay, S.M., Mpodozis, C., Ramos, V.A. and Munizaga, F. (1991). Magma source variations for mid to late Tertiary volcanic rocks erupted over a shallowing subduction zone and through a thickening crust in the Main Andean Cordillera (28-33°S). In: Andean Magmatism and its Tectonic Setting, eds. Harmon, R.S. and Rapela, C. *Geological Society of America Special Paper*, **265**, 113-137.

Kay, S.M., Ardolino, A., Franchi, M. and Ramos, V. (1993a). Origen de la Meseta de Somún Curá: distribución y geoquímica de sus rocas volcánicas máficas. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos (Mendoza), Actas*, **4**, 236-248.

Keidel, J. (1916). La geología de las Sierras de la Provincia de Buenos Aires y sus relaciones con las montañas de Sudáfrica y Los Andes. *Ministerio de Agricultura de La Nación, Sección Geología, Mineralogía y Minería Anales*, **XI**(3), 1-78.

Keidel, J. (1917). Über das Alter, die Verbreitung und die gegenseitigen Beziehungen der verschiedenen tektonischen strukturen in den argentinischen Gebirgen. *XII° Session du Congrés Géologique International (Prague) Compte Rendus*, 671-687.

Keidel, J. (1921). Sobre la distribución de los depósitos glaciares del Pérmico conocidos en la Argentina y su significación para la estratigrafía de la serie del Gondwana y la paleogeografía del Hemisferio Austral. *Academia Nacional de Ciencias Boletín*, **25**, 239-368.

Kellogg, J.N. and Bonini, W.E. (1982). Subduction of the Caribbean plate and basement uplifts in the overriding South American plate. *Tectonics*, **1**, 251-276.

Kellogg, J.N. and Bonini, W.E. (1985). Reply to C. Schubert. *Tectonics*, **4**(7), 785-790.

Kendrick, E.C., Bevis, M., Smalley Jr., R.F., Cifuentes, O. and Galban, F. (1999). Current rates of convergence acrpss the Central Andes: estimates fron continuous GPS observations. *Geophysical Research Letters*, **26**(5), 541-544.

Keppie, J.D. and Ortega-Gutiérrez, F. (1999). Middle America Precambrian basement: a missing piece of the reconstructed 1-Ga orogen. In: Laurentia Gondwana Connections before Pangea, eds. Ramos, V.A. and Keppie, D. *Geological Society of America Special Paper*, **336**, 199-210.

Keppie, J.D. and Ramos, V.A. (1999). Odyssey of terranes in the Iapetus and Rheic oceans during the Paleozoic. In: (eds.) Laurentia Gondwana Connections before Pangea, eds. Ramos, V.A. and Keppie, D. *Geological Society of America Special Paper*, **336**, 267-276.

Kerr, A.C., Marriner, G.F., Tarney, J., Nivia, A., Saunders, A.D., Thirlwall, M.F. and Sinton, C. (1997a). Cretaceous basaltic terranes in western Colombia: elemental, chronological and Sr Nd isotopic constraints on petrogenesis. *Journal of Petrology*, **38**, 677-702.

Kerr, A.C., Tarney, J., Marriner, G.F., Nivia, A., Klaver, G. and Saunders, A.D. (1997b). The geochemistry and tectonic setting of Late Cretaceous Caribbean and Colombian Volcanism. *Journal of South American Earth Sciences*, **307**(1-2), 235-254.

Klepeis, K.A. (1994). Relationship between uplift of the metamorphic core of the southernmost Andes and shortening in the Magallanes foreland fold and thrust belt, Tiera del Fuego, Chile. *Tectonics*, **13**(4), 882-904.

Klepeis, K.A. and Austin, J.A. (1997). Contrasting styles of superposed deformation in the southernmost Andes. *Tectonics*, **16**(5), 755-776.

Kley, J., Monaldi, C.R. and Salfity, J.A. (1999). Along-strike segmentation of the Andean foreland: causes and consequences, *Tectonophysics*, **301**, 75-94.

Kohn, M.J., Spear, F.S., Harrison, T.M. and Dalziel, I.W.D. (1995). $^{40}Ar/^{39}Ar$ geochronology P-T-t paths from the Cordillera de

Darwin metamorphic complex. *Journal of Metamorphic Petrology*, **13**, 251-270.

Kokogián, D.A., Fernández Seveso, F. and Mosquera, A. (1993). Las secuencias sedimentarias triásicas. In: Geología y Recursos Naturales de Mendoza, ed. Ramos, V.A. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos, Relatorio*, **1**(7), 65-78.

Kontak, D.J., Clark, A.H., Farrar, E. and Strong, D.F. (1985). The rift associated Permo-Triassic magmatism of the Eastern Cordillera: a precirsor of the Andean orogeny. In: *Magmatism at the Plate edge: the Peruvian Andes*, eds. Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D., pp. 36-44. Blackie Halstead Press, London.

Kozlowski, E., Manceda, R. and Ramos, V.A. (1993). Estructura. In: Geología y Recursos Naturales de Mendoza, ed. Ramos, V.A. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos, Mendoza, Relatorio*, 235-256.

Lapierre, H., Dupuis, V., Bosch, D., Polvé, M., Maury, R.C., Hernández, J., Monié P., Yéghicheyan, D., Jaillard, E., Tardy, M., Mercier de Lépinay, B., MambertiM., Desnet, A., Keller, F. and Sénebier, F. (1999). Multiple plume events in the genesis of the peri-Caribbean Cretaceous Oceanic Plateau Province. *Journal of Geophysical Research* (in press).

Lavenu, A., Cembrano, J., Arancibia, G., Deruelle, B., López Escobar, L. and Moreno, H. (1997). Neotectónica transpresiva dextral y volcanismo: Falla Iquiñe-Ofqui, Sur de Chile. *8° Congreso Geológico de Chile (Antofagasta), Actas*, **1**, 129-133.

Lebrat, M., Mégard, F., Juteau, T. and Calle, J. (1985). Pre-orogenic volcanic assemblages and structure in the Western Cordillera of Ecuador between 1°40'S and 2°20'S. *Geologische Rundschau*, **74**, 343-351.

Legarreta, L. and Uliana, M.A. (1996). The Jurassic succession in west-central Argentina: stratal patterns, sequences and paleogeographic evolution. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **120**, 303-330.

Limarino, C.O., Fauqué, L. and Net, L. (1999). Análisis tectosedimentario de las cuencas Paganzo y Río Blanco, Noroeste Argentino. Primer Simposio Argentino del Paleozoico Superior. *Ameghiniana*, **36**(4-Suplemento), Resúmenes, 33R.

Litherland, M., Aspden, J.A. and Jemielita, R.A. (1994). The metamorphic belts of Ecuador. *British Geological Survey Overseas Memoir*, **11**, 147 p.

Llambías, E. and Sato, A.M. (1995). El batolito de Colangüil: transición entre orogénesis y anorogénesis. *Asociación Geológica Argentina Revista*, **50**(1-4), 111-131.

López, G., Hatzfeld, D., Madariaga, R., Barrientos, S., Campos, J., Lyon-Caen, H., Zollo, A., Giannacone, G. and Kausel, E. (1997). Microsismicidad en la zona centro-sur de Chile. *7° Congreso Geológico de Chile (Antofagasta) Actas*, **3**, 1771-1774.

López Escobar, L., Cembrano, J. and Moreno, H. (1995). Geochemistry and tectonics of Chilean Southern Andes basaltic Quaternary volcanism (37°-46°S). *Revista Geológica de Chile*, **22**, 219-234.

Lugo, J. and Mann, P. (1995). Jurassic-Eocene tectonic evolution of Maracaibo basin, Venezuela. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez S., R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 699-725.

Macellari, C.E. (1988). Cretaceous paleogeography and depositional cycles of western South America. *Journal of South American Earth Sciences*, **1**(4), 373-418.

Manceda, R. and Figueroa, D. (1995). Inversion of the Mesozoic Neuquén rift in the Malargüe fold-thrust belt, Mendoza, Argentina. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 369-382.

Marechal, P. (1983). *Les Temoins de Chaîne Hercynienne dans le noyau ancien des Andes de Mérida (Vènèzuéla)*. Ph.D. Thesis, Université de Bretagne Occidentale, Brest, 176 p.

Marshall, L.G., Sempere, T. and Gayet, M. (1993). The Petaca (Late Oligocene-Early Miocene) and Yecua (Late Miocene) formations of the subandean-Chaco basin, Bolivia, and their tectonic significance. *Documents Laboratoire Géologique Lyon*, **125**, 291-301.

Martínez, M.P., Gímenez, M.E., Introcaso, A. and Robles, J.A. (1997). Excesos de espesores corticales y acortamientos andinos en tres secciones ubicadas en 36°, 37° y 39° de latitud sur. *8° Congreso Geológico Chileno (Antofagasta), Actas*, **1**, 101-105.

Mathalone, J.M.P. and Montoya, M. (1995). Petroleum geology of the Subandean basins of Peru. In Petroleum Basins of South America, eds. Tankard, A.J., Suárez, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 423-444.

McCourt, W.J. and Feininger, T. (1984). High-pressure metamorphic rocks in the Central Cordillera of Colombia. *British Geological Survey Report Series*, **16**(1), 28-35.

McCourt, W.J., Aspden, J.A. and Brook, M. (1984). New geological and geochronological data for the Colombian Andes: Continental growth by multiple accretion. *Journal of the Geological Society of London*, **141**, 831-845.

Mégard, F. (1987). Cordilleran Andes and Marginal Andes: a review of Andean geology north of the Arica elbow (18°S). In: Circum-Pacific orogenic belts and evolution of the Pacific Ocean basin, eds. Monger, J.M.H. and Francheteau, J. *International Lithosphere Program Contribution, Geodynamic Series*, **18**, 71-96.

Mégard, F., Dalmayrac, B., Laubacher, G., Marocco, R., Martínez, C., Paredes, J. and Tomasi, P. (1971). Lachaine hercynienne au Perou et en Bolivie, premiers resultats. *Cahiers Orstom, Serie Geologie*, **III**(1), 5-44.

Méndez Fajury, R.A. (1989). Catálogo de volcanes activos de Colombia. *Boletín Geológico del Ingeominas*, **30**(3), 1-75.

Meschede, M. and Frisch, W. (1998). A plate tectonic model for the Mesozoic and early Cenozoic history of the Caribbean plate. *Tectonophysics*, **296**(3-4), 269-292.

Millward, D., Marriner, G.F. and Saunders, A.D. (1984). Cretaceous tholeiitic volcanic rocks from the Western Cordillera of Colombia. *Journal of the Geological Society of London*, **141**, 847-860.

Mojica, J. and Dorado, J. (1987). El Jurásico anterior a los movimientos intermálmicos en los Andes Colombianos. In: *Bioestratigrafía de los sistemas regionales del Jurásico y Cretácico de América del Sur*, ed. Volkheimer, W. Comité Sudamericano del Jurásico y el Cretácico, Mendoza, 49-110.

Mojica, J., Villaroel, C., Cuerda, A. and Alfaro, M. (1988). La fauna de graptolitos de la Formación Higado (Llanvirniano-?Landeiliano), Serranía de Las Minas, Valle Superior del Magdalena, Colombia. *V° Congreso Geológico Chileno (Antofagasta), Actas*, II, C189-C202.

Mojica, J., Kammer, A. and Ujeta, G. (1996). El Jurásico del sector noroccidental de Suramérica y guía de la excursión al Valle Superior del Magdalena (Nov. 1-4/95), regiones de Payandé y Prado, Departamento del Tolima, Colombia. *Geología Colombiana*, **21**, 3-40.

Mosquera, A. (1999). Evolución temporal de la deformación en las Sierras Subandinas Orientales. *IV° Congreso de Exploración y Desarrollo de Hidrocarburos (Mar del Plata), Actas*, **II**, 573-568.

Moya, M.C. (1988). *Estratigrafía del Tremadociano en el tramo austral de la Cordillera Oriental Argentina*. Tesis Doctoral, Universidad Nacional de Salta, 223 p. (unpublished).

Moya, M.C., Malanca, S., Hongn, F.D. and Bahlburg, H. (1993). El Tremadoc temprano en la Puna Occidental Argentina. *XII°*

Congreso Geológico Argentino y II° Congreso Exploración de Hidrocarburos (Mendoza), Actas, **2**, 20-30.

Mpodozis, C. and Allmendinger, R.W. (1993). Extensional tectonics, Cretaceous Andes, northern Chile (27°S). *Geological Society of America Bulletin*, **105**, 1462-1477.

Mpodozis, C. and Cornejo, P. (1997). El rift triásico-sinemuriano de Sierra Exploradora, Cordillera de Domeyko (25-26°S): asociación de facies y reconstrucción tectónica. *VIII° Congreso Geológico Chileno (Antofagasta), Actas*, I, 550-553.

Mpodozis, C. and Forsythe, R. (1983). Stratigraphy and geochemistry of accreted fragments of the ancestral Pacific floor in Southern South America. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **41**, 103-124.

Mpodozis, C. and Kay, S.M. (1992). Late Paleozoic to Triassic evolution of the Gondwana margin: evidence from Chilean Frontal Cordilleran Batholiths (28°-31°S). *Geological Society of America Bulletin*, **104**, 999-1014.

Mpodozis, C. and Ramos, V.A. (1990). The Andes of Chile and Argentina. In: Geology of the Andes and its relation to Hydrocarbon and Mineral Resources. eds. Ericksen, G.E., Cañas Pinochet, M.T. and Reinemud, J.A. *Circumpacific Council for Energy and Mineral Resources, Earth Sciences Series*, 11, 59-90.

Mpodozis, C., Tomlinson, A.J. and Cornejo, P.C. (1994). Acerca del control estructural de intrusivos eocenos y pórfidos cupríferos en la región de Potrerillos-El Salvador. *7° Congreso Geológico Chileno (Concepción), Actas*, II, 1596-1600.

Mujica, H.J. and Zorzín, V. (1996). Extensión oriental de la tectónica andina en la llanura salteña, Provincia de Salta. *XIII° Congreso Geológico Argentino y III° Congreso Exploración de Hidrocarburos, (Buenos Aires), Actas*, I, 385-398.

Muñoz, J. and Stern, C. (1988). The Quaternary volcanic belt of the Southern continental margin of South America: transverse structural and petrochemical variations across the segment between 38°S and 39°S. *Journal of South American Earth Sciences*, 1(2), 147-162.

Muñoz, N., Charrier, R. and Reutter, K.-J. (1997). Evolución de la cuenca Salar de Atacama: inversión tectónica y relleno de una cuenca de antepaís de retroarco. *VIII° Congreso Geológico Chileno (Antofagasta), Actas*, I, 195-199.

Naranjo, J.A., Hervé, F., Prieto, X. and Munizaga, F. (1984). Actividad cretácica de la falla Atacama al este de Chañaral: milonización y plutonismo. *Comunicaciones*, Santiago de Chile, **34**, 37-45.

Nelson, E.P. (1982). Post-tectonic uplift of the Cordillera de Darwin orogenic core complex: evidence for fission track geochronology and closing temperature-time relationship. *Journal of the Geological Society of London*, **130**, 755-762.

Norabuena, E., Leffler-Griffin, L., Mao, A., Dixon, T., Stein, S., Sacks, I.S., Ocola, L. and Ellis, M. (1998). Space geodetic observations of Nazca-South America convergence across the Andes. *Science*, **279**, 358-362.

Olivero, E.B. (1998). Mesozoic-Paleogene geology of the marginal-Austral basin of Tierra del Fuego. *Third Annual Conference, IGCP Project 381, Aouth Atlantic Mesozoic Correlations, Field Trip Guide*, 1-42.

Olivero, E.B. and Malumián, N. (1999). Eocene stratigraphy of southeastern Tierra del Fuego island, Argentina. *American Association of Petroleum Geologists Bulletin*, **83**(2), 295-313.

Omarini, R.H., Sureda, R.J., Götze, H.J., Seilacher, A. and Plüger, F. (1999). Puncoviscana fold belt in Northwestern Argentina: testimony of Late Proterozoic Rodinia fragmentation and pre-Gondwana collisional episodes. *International Journal of Earth Sciences*, **88**, 76-97.

Orrego, L.A., Cepeda, H.V. and Rodríguez, S.G.I. (1980). Esquistos glaucofánicos en el área de Jambaló, Cauca, Colombia. *Geología Norandina*, **1**, 5-10.

Palma, M.A., Parica, P. and Ramos, V.A. (1987). El Granito Archibarca: su edad y significado tectónico. *Asociación Geológica Argentina Revista*, **41**(3-4, 1986), 414-418.

Pankhurst, R., Rapela, C., Saavedra, J., Baldo, E., Dahlquist, J., Pascua, I. and Fanning, C.M. (1998). The Famatinian magmatic arc in the Central Sierras Pampeanas: an Early to mid-Ordovician continental arc on the Gondwana margin. In: The Proto-Andean margin of Gondwana, eds. Pankhurst, R.J. and Rapela, C.W. *Geological Society of London Special Publication*, **142**, 343-367.

Parnaud, F., Gou, Y., Pascual, J.C., Capello, M.A., Truskowski, I. and Passalacqua, H., (1995). Stratigraphic synthesis of western Venezuela. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez S., R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 681-698.

Pardo Casas, F. and Molnar, P. (1987). Relative motion of the Nazca (Farallon) and South American Plates since Late Cretaceous time. *Tectonics*, **6**(3), 233-248.

Pennington, W.D. (1981). Subduction of the Eastern Panama Basin and seismotectonics of Northwest South America. *Journal of Geophysical Research*, **86**, 10753-10770.

Pérez, D.J., Ottone, G. and Ramos, V.A. (1996). La ingresión marina miocena en la provincia de San Juan: sus implicancias paleogeográficas. *XIII° Congreso Geológico Argentino y III° Congreso Exploración de Hidrocarburos (Buenos Aires), Actas*, I, 385-398.

Pilger, R.H. (1984). Cenozoic plate kinematics, subduction and magmatism: South American Andes. *Journal Geological Society of London*, 141, 793-802.

Pindell, J.L. (1985). Alleghenian reconstruction and subsequent evolution of the Gulf of Mexico, Bahamas, and Proto-Caribbean. *Tectonics*, **4**(1), 1-39.

Pindell, J.L. and Barrett, S.F. (1990). Geological evolution of the Caribbean region: a plate-tectonic perspective. In: The Caribbean Region, eds. Dengo, G. and Case, J.E., pp. 405-432. (The Geology of North America, Geological Society of America, H).

Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D. (1985). *Magmatism at a Plate Edge: The Peruvian Andes*. Blackie Halstead Press, London, 289 p.

Quenardelle, S. and Ramos, V.A. (1999). The Ordovician western Sierras Pampeanas magmatic belt: record of Precordillera accretion in Argentina. In: Laurentia Gondwana Connections before Pangea, eds. Ramos, V.A. and Keppie, D. *Geological Society of America Special Paper*, **336**, 63-86.

Ramos, V.A. (1970). Estratigrafía y estructura de la Sierra de Los Colorados, provincia de La Rioja. *Asociación Geológica Argentina Revista*, **XXV**(3), 359-382.

Ramos, V.A. (1982). Las ingresiones pacíficas del Terciario en el Norte de la Patagonia. *III° Congreso Geológico Chileno (Concepción), Actas*, I(A), 262-288.

Ramos, V.A. (1983). Evolución tectónica y metalogénesis de la Cordillera Patagónica. *II° Congreso Nacional Geología Económica (San Juan), Actas*, **1**, 108-124.

Ramos, V.A. (1985). El Mesozoico de la Alta Cordillera de Mendoza: reconstrucción tectónica de sus facies. Argentina. *IV° Congreso Geológico Chileno (Antofagasta), Actas*, I(2), 104-118.

Ramos, V.A. (1986). El diastrofismo oclóyico: un ejemplo de tectónica de colisión durante el Eopaleozoico en el noroeste Argentino. *Revista Instituto Ciencias Geológicas*, Jujuy, **6**, 13-28.

Ramos, V.A. (1988a). Tectonics of the Late Proterozoic-Early Paleozoic: a collisional history of southern South America. *Episodes*, **11**(3), 168-174.

Ramos, V.A. (1988b). The tectonics of the Central Andes: 30° to 33° S latitude. In: Processes in Continental Lithospheric Deformation, eds. Clark, S. and Burchfiel, D. *Geological Society of*

America Special Paper, **218**, 31-54.

Ramos, V.A. (1989a). The birth of Southern South America. *American Scientist*, 77(5), 444-450.

Ramos, V.A. (1989b). Foothills structure in Northern Magallanes Basin, Argentina. *American Association Petroleum Geologists Bulletin*, 73(7), 887-903.

Ramos, V.A. (1999). Plate tectonic setting of the Andean Cordillera. *Episodes*, **22**(3), 183-190.

Ramos, V.A. and Aguirre-Urreta, M.B. (1994). Cretaceous evolution of the Magallanes basin. In: *Cretaceous Tectonics of the Andes*, Earth Evolution Series, ed. Salfity, J.A., pp. 315-345. Fried. Vieweg & Sohn, Braunschweig/Wiesbaden.

Ramos, V.A. and Aguirre-Urreta, M.B. (2000). Patagonia. In: Tectonic evolution of South America, eds. Milani, E.J. and Thomaz Filho, A. *31° International Geological Congress* (Rio de Janeiro), this volume.

Ramos V.A. and Alonso, R.N. (1995). El mar paranense en la provincia de Jujuy. *Revista Instituto de Geología y Minería*, Jujuy, **10**, 73-82.

Ramos, V.A. and Barbieri, M. (1989). El volcanismo Cenozoico de Huantraico: edad y relacionesisotopicas iniciales, provincia del Neuquen. *Asociación Geológica Argentina Revista*, **XLIII**(2), 210-223.

Ramos, V.A. and Basei, M. (1997a). The basement of Chilenia: an exotic continental terrane to Gondwana during the Early Paleozoic. *Symposium on Terrane Dynamics'97 (Christchurch)*, 140-143.

Ramos, V.A. and Basei, M. (1997b). Gondwanan, Perigondwanan, and exotic terranes of southern South America. *South American Symposium on Isotope Geology (São Paulo), Abstracts*, 250-252.

Ramos, V.A. and Kay, S.M. (1991). Triassic rifting and associated basalts in the Cuyo basin, central Argentina. In: Andean Magmatism and its Tectonic Setting, eds. Harmon, R.S. and Rapela, C.W. *Geological Society of America Special Paper*, **265**, 79-91.

Ramos, V.A. and Kay, S.M. (1992). The Southern Patagonian plateau basalts: retroarc testimony of a ridge collision, Argentina. *Tectonophysics*, **205**, 261-282.

Ramos, V.A., Munizaga, F. and Kay, S.M. (1991). El magmatismo cenozoico a los 33°S de latitud: geocronología y relaciones tectónicas. *VI° Congreso Geológico Chileno (Viña del Mar), Actas*, **1**, 892-896.

Ramos, V.A., Cegarra, M. and Cristallini, E. (1996a). Cenozoic tectonics of the High Andes of west-central Argentina, (30°-36°S latitude). *Tectonophysics*, **259**, 185-200.

Ramos, V.A., Niemeyer, H., Skarmeta, J. and Muñoz, J. (1982). The magmatic evolution of the Austral Patagonian Andes. *Earth Science Reviews*, **18**(3-4), 411-443.

Ramos, V.A., Jordan, T.E., Allmendinger, R.W., Mpodozis, C., Kay, S., Cortés, J.M., Palma, M.A. 1986. Paleozoic terranes of the Central Argentine-Chilean Andes. *Tectonics*, 5(6), 855-880.

Ramos, V.A., Aguirre Urreta, M.B., Alvarez, P.P., Cegarra, M., Cristallini, E.O., Kay, S.M., Lo Forte, G.L., Pereyra, F. and Pérez, D. (1996b). *Geología de la Región del Aconcagua, Provincias de San Juan y Mendoza*. Dirección Nacional del Servicio Geológico Anales, **24**, 1-510.

Rapalini, A.E. and Vilas, J.F. (1991). Preliminary paleomagnetic data from the Sierra Grande Formation: Tectonic consequences of the first mid-Paleozoic paleopoles from Patagonia. *Journal of South American Earth Sciences*, **4**, 25-41.

Rapalini, A.E. and Vizán, H. (1993). Evidence of intrapangea movements in Gondwanaland. *XII° International Congress on Carboniferous - Permian systems (Buenos Aires), Comptes Rendus*, **1**, 405-434.

Rapalini, A.E., Astini, R.A. and Conti, C.M. (1999). Paleomagnetic constraints on the evolution of Paleozoic suspect terranes from South America. In: Laurentia Gondwana Connections before Pangea, eds. Ramos, V.A. and Keppie, D. *Geological Society of America Special Paper*, **336**, 171-182.

Rapela, C.W., Pankhurst, R.J., Dahlquist, J. and Fanning, C.M. (1999). U-Pb Shrimp ages of Famatinian granites: new constraints on the timing, origin and tectonic setting of I- and S-type magmas in an ensialic arc. *II° South America Symposium on Isotope Geology (Córdoba), Actas*, 264-267.

Räsänen, M.E., Linna, A.M., Santos, J.C.R. and Negri, F. (1995). Late Miocene tidal deposits in the Amazonian foreland basin. *Science*, **269**, 386-390.

Restrepo, J.J. and Toussaint, J.F. (1973). Obducción cretácea en el occidente Colombiano. *Publicaciones Especiales de Geología, Universidad Nacional, Facultad de Minas*, Medellin, **3**, 1-26.

Restrepo, J.J. and Toussaint, J.-F. (1988). Terranes and continental accretion in the Colombian Andes. *Episodes*, **11**, 189-193.

Restrepo-Pace, P.A. (1992). Petrotectonic characterization of the Central Andean Terrane, Colombia. *Journal of South American Earth Sciences*, **5**, 97-116.

Restrepo-Pace, P.A. (1999). Fold and thrust belt along the western flankof the Eastern Cordillera of Colobia: style, Kinematics and timing constraints derived from seismic data and detailed surface mapping. In: *Thrust Tectonics*, ed. McClay, K., 112-114.

Restrepo-Pace, P.A., Ruiz, J., Gehrels, G. and Cosca, M. (1997). Geochronology and Nd isotopic data of Grenville-age rocks in the Colombian Andes: new constraints for Late Proterozoic-Early Paleozoic paleocontinental reconstructions of the Americas. *Earth and Planetary Science Letters*, **150**, 427-441.

Reynaud, C., Jaillard, E., Lapiere, H., Mamberti, M. and Mascle, G. (1999). Oceanic plateau and island arcs of southwestern Ecuador: their place in the geodynamic evolution of northwestern South America. *Tectonophysics*, **307**, 235-254.

Robin, C., Hall, M., Jimenez, M., Monzier, M. and Escobar, P. (1997). Mojanda volcanic examples (Ecuador): development of two adjacent contemporaneous volcanoes with contrasting eruptive styles and magmatic suites. *Journal of South American Earth Sciences*, **10**(5-6), 345-359.

Robison, R. and Pantoja-Alor, J. (1968). Tremadocian trilobites from the Nochixtlán region, Oaxaca, Mexico. *Journal of Paleontology*, **42**(3), 767-800.

Rossello, E. and Mozetic, M.E. (1999). Caracterización estructural y significado geotectónico de los depocentros cretácicos continentales del centro-oeste argentino. *5° Simposio sobre o Cretáceo do Brasil (Rio Claro), Boletim*, 107-113.

Ruiz, T., Tosdal, R.M., Restrepo, P.A. and Murillo-Muñetón, G. (1999). Pb Isotope evidence for Colombia-southern Mexico connections in the Proterozoic. In: Laurentia Gondwana Connections before Pangea, eds. Ramos, V.A. and Keppie, D. *Geological Society of America Special Paper*, **336**, 183-198.

Russo, R.M. and Silver, P.G. (1996). Cordillera formation, mantle dynamics, and the Wilson cycle. *Geology*, **24**(6), 511-514.

Rutland, R.W.R. (1971). Andean orogeny and sea floor spreading. *Nature*, **233**, 252-255.

Sadowski, G.R. and Bettencourt, J.S. (1994). Amazon and Laurentia, former neighbours. *38°Congresso Brasileiro de Geologia (Camboriú), Resumos*, **110**.

Salfity, J. (1982). Evolución paleogeográfica del Grupo Salta (Cretácico-Eogénico), Argentina. *V° Congreso Latinoamericano de Geología (Buenos Aires), Actas*, **1**, 11-26.

Salfity, J.A., ed. (1994). *Cretaceous Tectonics of the Andes*. Earth Evolution Series. Fried. Vieweg & Sohn, Braunschweig/Wiesbaden, 345 p.

Salfity, J.A. and Marquillas, R.A. (1994). Tectonic and sedimentary evolution of the Cretaceous-Eocene Salta Group basin, Argentina.

In: *Cretaceous Tectonics of the Andes*, Earth Evolution Series, ed. Salfity, J.A., pp. 266-315. Fried. Vieweg & Sohn, Braunschweig/ Wiesbaden.

Sandeman, H.A., Clark, A.H. and Farrar, E. (1995). An integrated tectono-magmatic model for the evolution of the Southern Peruvian Andes (13°-20°S) since 55 Ma. *International Geology Review*, 37, 1039-1073.

Sanjines-Saucedo, G. (1982). Estratigrafía del Carbónico, Triásico y Cretácico Boliviano en el borde oriental de las Sierras Subandinas Bolivianas. *5° Congreso Latinoamericano de Geología (Buenos Aires), Actas*, I, 301-318.

Scheuber, E., Bogdanic, T., Jensen, A. and Reutter, K.-J. (1994). Tectonic development of the North Chilean Andes in relation to plate convergence and magmatism since the Jurassic. In: *Tectonics of the Southern Central Andes, Structure and evolution of an active continental margin*, eds. Reutter, K.-J., Scheuber, E. and Wigger, P.J., pp. 121-139. Springer-Verlag.

Schmitz, M. (1994). A balanced model of the southern Central Andes. *Tectonics*, 13(2), 484-492.

Schubert, C. (1982). Neotectonics of Boconó fault, Western Venezuela. *Tectonophysics*, **85**, 205-220.

Schubert, C. and Vivas, L. (1993). *El Cuaternario de la Cordillera de Mérida, Andes Venezolanos*. Universidad de Los Andes, Fundación Polar, Mérida, 345 p.

Sébrier, M. and Soler, P. (1991). Tectonics and magmatism in the Peruvian Andes from Late Oligocene time to the present. In: *Andean Magmatism and its Tectonic Setting*, eds. Harmon, R.S. and Rapela, C. *Geological Society of America Special Paper*, **265**, 259-276.

Sébrier, M., Mercier, J.L., Mégrad, F., Laubacher, G. and Carey-Gaihardis, E. (1985). Quaternary normal and reverse faulting and the state of stress in the central Andes of south Peru. *Tectonics*, 7, 739-780.

Sébrier, M., Mercier, J.L., Macharé, J., Bonnot, D., Cabrera, J. and Blanc, J.L. (1988). The state of stress in an overriding plate situated above a flat slab: The Andes of central Peru. *Tectonics*, 7, 895-928.

Sempere, T. (1995). Phanerozoic evolution of Bolivia and adjacent regions. In: Petroleum Basins of South America, eds. Tankard, A.J., Suárez, R. and Welsink, H.J. *American Association of Petroleum Geologists Memoir*, **62**, 207-230.

Sempere, T., Hérail, G., Oller, J. and Bonhomme, M. (1990). Late Oligocene-early Miocene major tectonic crisis and related basins in Bolivia. *Geology*, **18**(10), 946-949.

Smalley, R.F. and Isacks, B.L. (1990). Seismotectonics of thin and thick-skinned deformation in the Andean foreland from local network data: evidence for a seismogenic lower crust. *Journal of Geophysical Research*, **95**, 12487-12498.

Smalley, R., Pujol, J., Regnier, M., Ming Chiu, J., Chatelain, J. L., Isacks, B., Araujo, M. and Puebla, N. (1993). Basement seismicity beneath the Andean Precordillera thin-skinned thrust belt and implications for crustal and lithospheric behaviour. *Tectonics*, **12**(1), 63-76.

Soler, P. and Bonhomme, M. (1990). Relations of magmatic activity to plate dynamics in Central Peru from Late Cretaceous to present. In: Plutonism from Antarctica to Alaska, eds. Kay, S. and Rapela, C.W. *Geological Society of America Special Paper*, **241**, 173-191.

Somoza, R. (1995). *Paleomagnetismo de rocas cretácicas de la Patagonia y la curva de deriva polar aparente de América del Sur: implicancias geocinemáticas y tectónicas*. PhD Thesis, Universidad de Buenos Aires, 238 p. (unpublished)

Steinmann, M., Hungerbühler, Seward, D. and Winkler, W. (1999). Neogene tectonic evolution and exhumation of the southern Ecuadorian Andes: a combined stratigraphiy and fission-track approach. *Tectonophysics*, **307**, 255-276.

Stephan, J.F. (1982). *Evolution geodynamique du domaine caraibe Andes et chaines caraibe sur la transversele de Barquisimeto, Venézuéla*. Ph.D. Thesis, Université Pierre et Marie Curie, Paris, 1 and 2, 512 p.

Stern, C.R. (1990). Pliocene to present migration of the volcanic front, Andean Southern Volcanic zone. *Revista Geológica de Chile*, **16**(2), 145-162.

Stern, Ch.R. (1991). Role of subduction erosion in the generation of Andean Magmas. *Geology*, **19**(1), 78-81.

Stern, C. and Kilian, R. (1996). Role of the subducted slab, mantle wedge and continental crust in the generation of adakites from the Andean Austral Volcanic Zone. *Contributions to Mineralogy and Petrology*, **123**, 263-281.

Stern, Ch., Skewes, M.A. and Durán, M. (1976a). Volcanismo orogénico en Chile Austral. *I° Congreso Geológico Chileno (Santiago), Actas*, II (F), 195-212.

Stern, Ch., de Wit, M.J. and Lawrence, J.R. (1976b). Igneous and metamorphic processes associated with the formation of Chilean ophiolites and their implications for ocean floor metamorphism, seismic layering, and magmatism. *Journal of Geophysical Research*, **81**, 4370-4380.

Suárez, G., Molnar, P. and Burchfield, B.C. (1983). Seismicity, fault plane solutions, depth of faulting, and active tectonics of the Andes of Peru, Ecuador and southern Colombia. *Journal of Geophysical Research*, **88**, 10403-10428.

Suárez, M. (1976). Plate tectonic model for Southern Antarctica Peninsula and its relation to Southern Andes. *Geology*, 4, 211-214.

Suárez, M. and Bell, C.M. (1993). Triassic-rift related sedimentary basins in northern Chile (24°-29°S). *Journal of South American Earth Sciences*, 6(3), 109-121.

Suárez Soruco, R. (1999). Bolivia: Late Proterozoic - Early Paleozoic. *IV° International Symposium on Andean Geodynamics, (Goettingen), Proceedings*, 716-718.

Schwartz, D. (1988). Paleoseismicity and neotectonics of the Cordillera Blanca fault zone, Northern Peruvian Andes. *Journal of Geophysical Research*, **93**, 4712-4730.

Taylor, G.K., Grocott, J., Pope, A. and Randall, D.E. (1998). Mesozoic fault systems, deformation and fault block rotation in the Andean forearc: a crustal scale strike-slip duplex in the Coastal Cordillera of northern Chile. *Tectonophysics*, **299**(1-3), 93-109.

Tebbens, S.F., Cande, S., Kovacs, L., Parra, J.C., LaBreque, J.L. and Vergara, H. (1997). The Chile ridge: a tectonic framework. *Journal Geophysical Research*, **102**(B6), 12035-12059.

Thorpe, R.S. (1984). The tectonic setting of active Andean volcanism. In: *Andean Magmatism*, eds. Harmon, R.S. and Barreiro, B.A., pp. 4-8. Cheshire, Shiva Pub. Ltd.

Thorpe, R.S. and Francis, P.W. (1979). Variations in Andean andesite compositions and their petrogenetic significance. *Tectonophysics*, **57**, 53-70.

Tomlinson, A.J. and Blanco, N. (1997). Structural evolution and displacement history of the west fault system, Precordillera, Chile: Part I: Synmineral history. *VIII° Congreso Geológico Chileno (Antofagasta), Actas*, **3**, 1873-1877.

Tomlinson, A.J., Mpodozis, C., Cornejo, P., Fernández, C.F. and Dumitru, T. (1994). El sistema de fallas Sierra Castillo-Agua Amarga: transpresión sinistral eocena en la Precordillera de Potrerillos-El Salvador. *7° Congreso Geológico Chileno (Concepción), Actas*, **II**, 1459-1463.

Toselli, A., Saavedra, J. and Rossi de Toselli, J.N. (1996). Interpretación geotectónica del magmatismo del sistema de Famatina. In: Geología del Sistema de Famatina, eds. Aceñolaza, F.G., Miller, H. and Toselli, A.J. *Münchner Geologische Hefte*, **A 19**, 283-292.

Toussaint, J.-F. (1993). *Evolución Geológica de Colombia*, *Precámbrico-Paleozoico.* Universidad Nacional de Colombia, Medellín, 229 p.

Toussaint, J.-F. and Restrepo, J.J. (1982). Magmatic evolution of the northwestern Andes of Colombia. *Earth Science Reviews*, **18**, 205-213.

Toussaint, J.-F. and Restrepo, J.J. (1994). The Colombian Andes during Cretaceous times. In: *Cretaceous Tectonics of the Andes.* Earth Evolution Sciences, Monograph Series, ed. Salfity, J., pp. 61-100. Friedr. Vieweg & Sohn.

Uliana, M.A. and Biddle, K.T. (1988). Mesozoic-Cenozoic paleogeographic and geodynamic evolution of southern South America. *Revista Brasileira de Geociências*, **18**, 172-190.

Uliana, M.A. and Legarreta, L. (1993). Hydrocarbons habitat in a Triassic-to-Cretaceous sub-Andean setting: Neuquén basin, Argentina. *Journal of Petroleum Geologists*, **16**(4), 397-420.

Uyeda, S. (1983). Comparative Subductology. *Episodes*, **1983**(2), 19-24.

Van der Wiel, A.M. (1991). *Uplift and volcanism of the SE Colombian Andes in relation to Neogene sedimentation in the Upper Magdalena Valley.* Thesis University of Wageningen, Amsterdam, 208 p.

Van Thornout, F., Hertogen, J. and Quevedo, L. (1992). Allochthonous terranes in northwestern Ecuador. *Tectonophysics*, **205**(1-3), 205-221.

Vicente, J.C. (1972). Apercu sur l'organization et l'evolution des Andes argentino-chiliennes centrales au paralléle de Aconcagua. *XXIV° International Geological Congress (Ottawa), Section*, 3, 423-436.

Vicente, J.C. (1990). Early Cretaceous overthrusting in the Western Cordillera of southern Peru. In: Geology of the Andes and its relation to Hydrocarbon and Mineral Resources, eds. Ericksen, G.E., Cañas Pinochet, M.T. and Reinemud, J.A. *Circumpacific Council for Energy and Mineral Resources, Earth Sciences Series*, 11, 91-117.

Vicente, J.C., Sequeiros, F., Valdivia, M.A. and Zavala, J. (1979). El sobre-escurrimiento de Cuicha-Lluta: elemento del accidente mayor andino al noroeste de Arequipa. *Sociedad Geológica del Perú Boletín*, **61**, 67-100.

Villamil, T. (1999). Campanian-Miocene tectonostratigraphy, depocenter evolution and basin development of Colombia and western Venezuela. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **153**, 239-275.

Von Huene, R. and Scholl, D.W. (1991). Observations at convergent margins concerning sediment subduction, subduction erosion and the growth of continental crust. *Review Geophysique*, **29**(3), 279-316.

Von Huene, R., Weinrebe, W. and Heeren, F. (1999). Subduction erosion along the Northern Chile margin. *Journal of Geodynamics*, **27**, 345-358.

Webb, S.D. (1995). Biological implications of the Middle Miocene Amazon seaway. *Science*, **269**, 361-362.

Woodward, H.P. (1958). Alleghany orogeny. *American Association of Petroleum Geologists Bulletin*, **42**, 193.

Wörner, G., Moorbath, S., Horn, S., Entenmann, J., Harmon, R.S., Davidson, J.P. and López Escobar, L. (1994). Large- and fine scale geochemical variations along the Andean Arc of Northern Chile (175°-22°S). In: *Tectonics of the Southern Central Andes, Structure and Evolution of an Active Continental Margin*, eds. Reutter, K.J., Scheuber, E. and Wigger, P.J., pp. 77-91. Berlin, Springer Verlag.

Zambrano, J.J. (1987). Las cuencas sedimentarias de América del Sur durante el Jurásico y el Cretácico: su relación con la actividad tectónica y magmática. In: *Bioestratigrafía de los sistemas regionales del Jurásico y Cretácico de América del Sur*, ed. Volkheimer, W., pp. 1-48. Comité Sudamericano del Jurásico y el Cretácico, Mendoza.

Zapata, T.R. and Allmendinger, R.W. (1996). The thrust front zone of the Precordillera thrust belt, Argentina: a thick-skinned triangle zone. *American Association of Petroleum Geologists Bulletin*, **80**(3), 359-381.

Zapata, T.R., Brissón, I. and Dzelalija, F. (1999). The role of basement in the Andean fold and thrust belt of the Neuquén basin. In: McClay, K., (ed.) *Thrust Tectonics 99 (London), Abstracts*, 122-124.

Zentili, M. and Maksaev, V. (1995). Metallogenetic model for the Late Eocene-Early Oligocene Supergiant porphyry event, Northern Chile. In: Giant Ore Deposits II, ed. Clark, A.H. *Proceedings of the Second Giant Ore Deposits Workshop, Ontario*, 138A-138N.

Zeil, W. (1979). *The Andes: A Geological Review*. Gebrüder Borntaeger, 1-260.

The Southern Andes

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The Andes is the mountain range that formed in response to eastward subduction of Pacific oceanic plates beneath the overriding edge of the South American Plate. Although the present range has acquired its topographic relief since late Cenozoic times, its geological constitution reveals a long history of tectonic activity starting in the Paleozoic or earlier. Several authors have drawn attention to the segmented character of the Andes as a whole (Corvalán, 1989; Mpodozis and Ramos, 1989), arising from both active subduction processes and their earlier geological evolution.

Segments are transitional to one another because of shifting boundaries through time, and are thus difficult to establish precisely. We will refer in this paper to an arbitrary segment of the Andes, extending between latitudes, 44 °S and 56 °S, which encompasses the whole of the Chilean regions of Aysén and Magallanes. We will emphasize the geology of Chile, and refer to the slope of the range provinces of Argentina when appropriate. The considered segment will be divided into the Patagonian Andes, N of the Straits of Magellan, and the Fuegan Andes, farther to the S. However, in its wider acceptation, Patagonia extends to the S of the established limit and includes Tierra del Fuego.

We present below an up-to-date account of the geological evolution of this segment, with emphasis on the main geological units and their tectonic significance. In many areas the geology is not known in much detail, and the region is at present being investigated by different research groups from various perspectives, with a rapid increase in the knowledge of the basic geology. We will refer to previous works for detailed description of geological sections, but will incorporate as much detailed information as possible from our own recent and ongoing research which mainly concerns the Patagonian Andes. The only published geological maps of the whole area are the 1:1 000 000 map of Escobar *et al.* (1980) and the 1:2 000 000 map of Frutos *et al.* (1986).

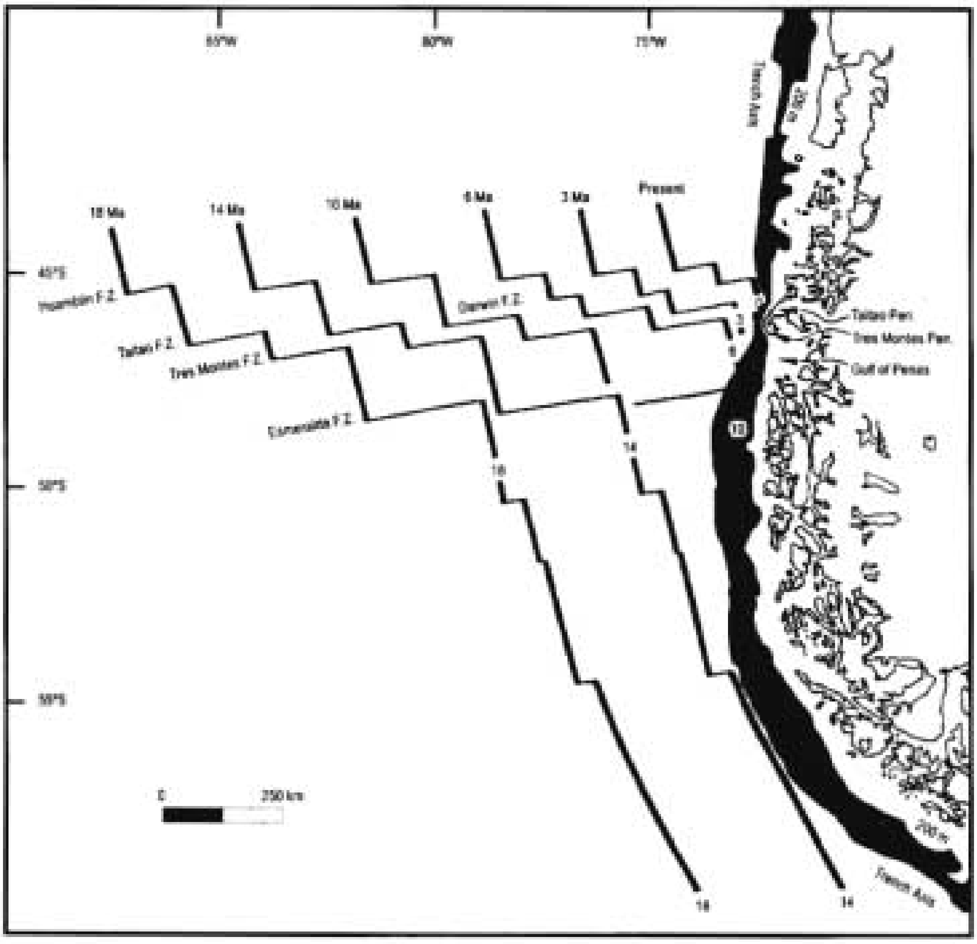


FIGURE 1 - Successive relative positions of the Chile Ridge and the South American margin since 18 Ma (after Cande and Leslie, 1986).

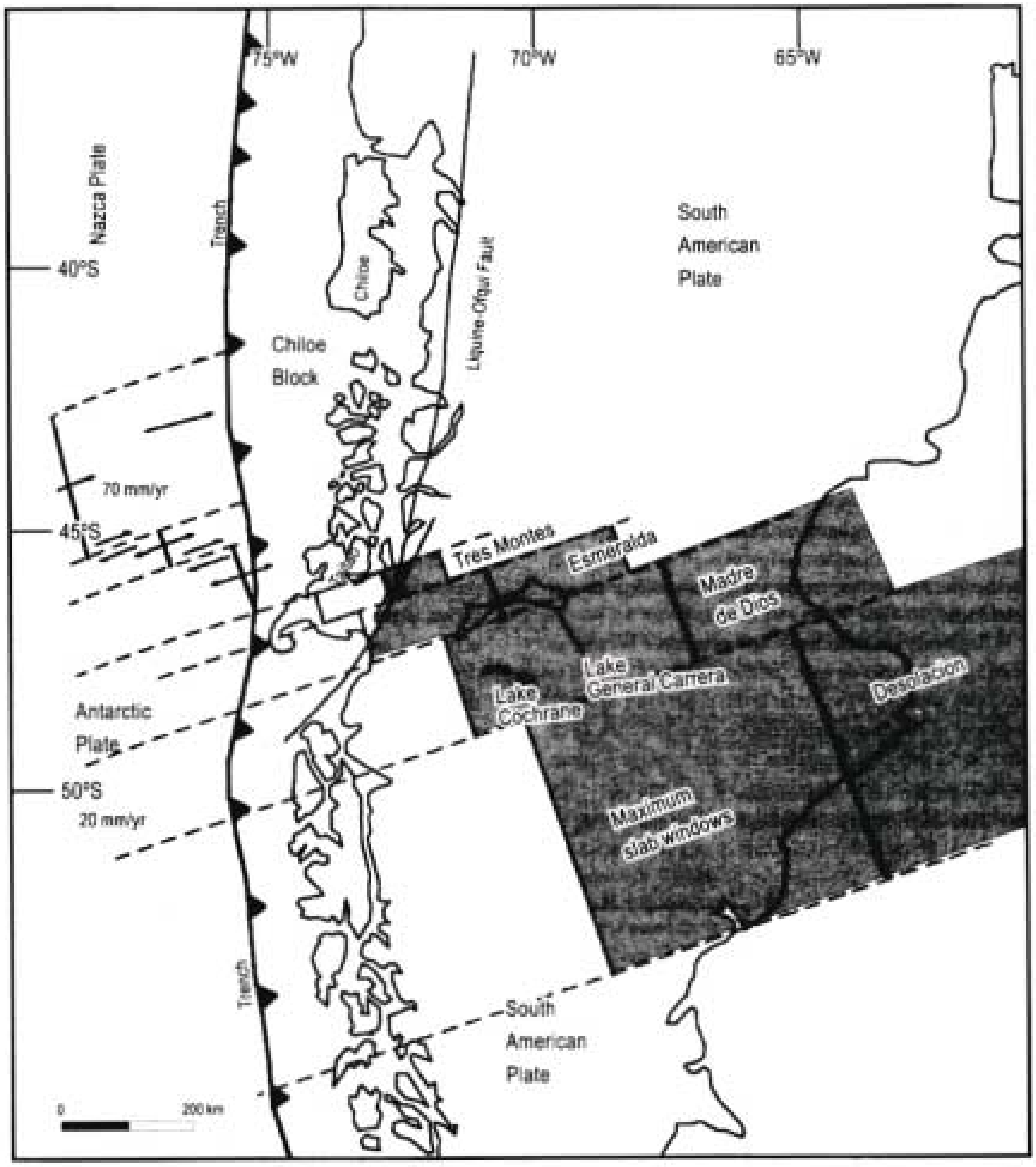


FIGURE 2 - Sketch map showing the supposed present position of subducted ridge segments, named after the fracture zone at its southern end. The maximum extent of slab windows supposes that crustal accretion ceased immediately after the ridge segments entered the trench, assuming a 4.5 cm/year separation of the Nazca and Antarctic plates (after Murdie, 1994).

Morphotectonic units

In terms of the classical threefold morphostructural E-W division of Central Chile, this segment can be divided into a Coastal Range, a Central Valley, and a Main Range that can be traced down to latitude 47 °S. The Coastal Range is represented by the Chonos Archipelago; the Central Valley by the Moraleda Canal; and the Main Range by the Andean Chain, which reaches its highest elevation in the segment at Mount San Valentín (4050 m). A feature of this sector of the Andes is that the watershed lies to the E of the Main Range, whereas in Central Chile, the two coincide. To the E of the main Andean Chain, there lies a vast flat-lying plateau referred to as extra-Andean Patagonia that extends as far as the Atlantic coast.

To the S of the Gulf of Penas, a large embayment at 47 °S, there exists no similar feature to the Central Valley, and the coastal archipelagos merge progressively into the Main Range, which is covered here by the Southern Patagonian Icefield, the third largest in the world. A similar situation exists in the Fuegan Andes, where the Beagle Channel separates the islands of the Pacific coast from the Main Range, here called the Darwin Cordillera.

Cenozoic plate tectonic setting

The Cenozoic subduction history of the segment is relatively well known from the analysis of the adjacent sea floor magnetic anomalies (Cande and Leslie, 1986; Pardo Casas and Molnar, 1987). It shows a consistent right-oblique subduction of the Farallon (Nazca) Plate beneath South America for the last 48 Ma, with a nearly orthogonal convergence between 26 and 20 Ma. The present day obliquity is about 26° for Jarrard (1986) who also estimated the subduction dip at 16 °. The age of the subducting Nazca Plate decreases from *c.* 25 Ma at 38 °S to 0 Ma at the Chile Triple Junction where the Nazca, Antarctica and South America plates (NAS) meet at a point presently situated at 46 °S. This triple junction has migrated northwards during late Cenozoic times, after its origin near the southern tip of South America in the Miocene, 14 Ma ago. A long segment of the ridge was then subducted, and several smaller segments have been subducted in the last 5 Ma near the Gulf of Penas (Fig. 1). A short segment of the Chile Rise is presently being subducted, somewhere N of the Taitao Peninsula.

This scenario results in different subduction regimes N and S of the Chile Triple Junction. To the N, subduction of the Nazca Plate is taking place with a slight obliquity at nearly 8 cm/year, while to the S subduction is very slow, around 2 cm/year. This relationship is in keeping with the differential development of the active volcanic arc; with the more active volcanoes to the N of the Chile Triple Junction. The angle of subduction is 25 to 30 ° to the N of the NAS (Bourgois *et al.*, 1996) and is unknown to the S.

Another consequence of spreading ridge subduction is the possible generation of asthenosphere windows below the South American plate lithosphere. The probable positions of these windows related to successively subducted segments of the ridge are shown in Figure 2. The possible geological effect of these features will be analysed in the section on volcanism to follow.

Although less well defined, Cande and Leslie (1986) present evidence that a triple junction between the Farallon, Aluk and South America plates migrated southwards during the Paleogene, reaching the southern tip of the South American continent *c.* 40 Ma ago. Although its geological effects are less constrained, traces of this tectonic feature are preserved in the geological record of the upper plate.

Seismic activity

Seismic activity appears to be limited in the Patagonian Andes, as compared with more northern segments of the Andes. The available data suggest that the Chile fore arc between 39 °S and 46 °S is mainly undergoing a dextral strike slip motion, and that the volcanic arc is absorbing a small trench parallel component (Cembrano, 1998). The back arc is not undergoing shortening at present.

Cifuentes (1989), in her analysis of the 1960 earthquake, concluded that the rupture zone associated with this event, the largest earthquake ever recorded (9.3 Ms), extended from the hipocentre at 37 °S to the NAS Chile Triple Junction 46 °S, comprising the whole sector here denominated the Northern Patagonian Andes. Murdie (1994) interpreted the microseismicity recorded near the end of the Taitao Peninsula as representing normal faulting (down to the W) of the fore-arc region. Chinn and Isacks (1983) determined a strike-slip focal plane solution for a 6.0 Ms tremor in 1965 at 46 °S beneath the trace of the Liquiñe-Ofqui Fault Zone (LOFZ).

Fuenzalida (1974) analysed the seismicity S of the NAS Chile Triple Junction. Two earthquakes with left lateral strike-slip solutions have hypocentres situated along the NNW trending Magellan Fault Zone, which extends from the Pacific margin into the North Scotia Ridge. He interpreted this fault zone as a connecting structure between the Pacific Trench and the North Scotia Transform Fault, and as a factor in the oroclinal bending and dispersion of the southern end of the Andes.

The Patagonian Andes

The backbone of the Patagonian Andes mainly consists of the subduction-related Patagonian Batholith (PB, Fig. 3) of Mesozoic to Cenozoic age, which runs continuously along the segment, with an average width of over 120 km. To the N of the Gulf of Penas, the Patagonian Batholith occurs below the active magmatic arc, whereas to the S it is situated wholly in a fore arc position relative to the present day volcanic chain.

Most of the rocks exposed to the W of the Patagonian Batholith, represent Late Paleozoic to Early Mesozoic subduction complexes: the Chonos Metamorphic Complex N of Gulf of Penas, and the Denaro, Duque de York, Tarlton and Almagro complexes to the S. The Diego Ramírez Islands, the southernmost exposures of the South American continent, consist of a Late-Jurassic melange with blueschist mineralogy (Davidson *et al.*, 1987; Wilson *et al.*, 1989). The late Cenozoic Taitao Ophiolite and related volcanic and sedimentary sequences are a singular rock association, which occurs at the western tip of the Taitao Peninsula.

To the E of the Patagonian Batholith, there occurs a supposedly Late Devonian to Early Carboniferous low-grade turbidite sequence, with massive limestone bodies towards its northern exposures near Lake General Carrera, and with sub-greenschist to greenschist facies metamorphism. These rocks are the basement upon which lie the Mesozoic and Cenozoic volcanic and sedimentary units. The Mesozoic sequence starts with the Ibañez Formation, an acid volcanic sequence related to vast units of similar age and lithology which cover most of extra-Andean Patagonia, all the way to the Atlantic coast (Pankhurst *et al.*, 1998). This silicic province is, initially at least, the result of extensional processes which preceded the opening of the South Atlantic. Subsequent sedimentary units of the Aysén Basin (equivalent to the Mayo River embayment in Argentina) were deposited in Tithonian to Hauterivian times. These were covered during the Aptian by a second acid volcanic sequence, the Divisadero Formation, which at his time was restricted to the Andean region. Plateau basalts were erupted from Late Cretaceous to the present, with a related Atlantic marine incursion in Oligocene to Miocene times, recorded in the deposits of the Cosmelli Basin in the eastern foothills of the Main Range. Holocene volcanic activity is represented by large stratovolcanoes and by numerous minor eruptive centres, which are mainly situated in valleys that follow late Cenozoic faults and shear zones.

The Patagonian Andes were the site of extensive Pleistocene ice sheets, of which the present day icefields are remnants. Glaciers extended to the Pacific Coast until 20

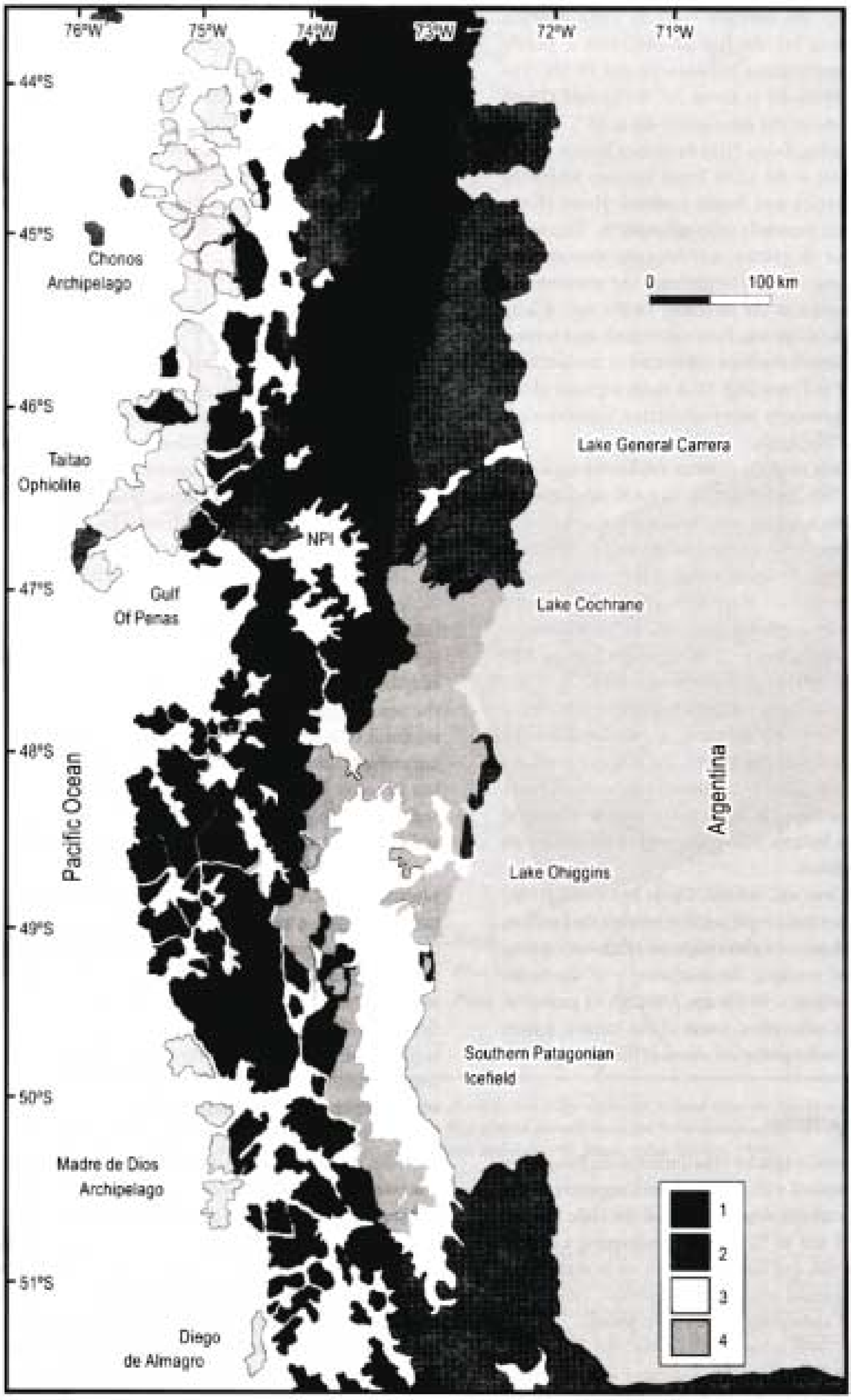


FIGURE 3 - Geological sketch map of the Patagonian segment (44 °S to 51 °S) of the Southern Andes, as referred to in text. 1. Patagonian Batholith, of Mesozoic to Cenozoic age. 2. Mesozoic and Cenozoic stratified units of volcanic and sedimentary rocks. 3. Fore-arc accretionary complexes with indication of the main outcrop localities: Chonos Archipelago, Madre de Dios Archipelago, Diego de Almagro. Late Paleozoic to Early Mesozoic ages. 4. Eastern Andes Metamorphic Complex, of probable late Devonian to Carboniferous depositional age. The patagonian icefields, (NPI = North Patagonian Icefield) in white. The late Cenozoic Taitao (Bahia Barrientos) ophiolite included in 3. Active volcanoes not shown (modified after Frutos et al., 1986).

000 years ago or less throughout the segment, and extended eastwards far into extra-Andean Patagonia, where well developed moraine systems are found.

The coastal subduction complexes

Subduction processes have been active along the continental margin since the Late Paleozoic and have built an accretionary complex, which forms the country rock of the Chonos Archipelago, the Taitao Peninsula, and the Madre de Dios and Diego de Almagro archipelagos. These rocks probably extend towards the present trench, but the Cenozoic components of the accretionary wedge are mainly submarine. We will refer in this section to the exposed old components of the accretionary complex.

The Chonos Metamorphic Complex

The low-grade metamorphic rocks that make up most of the Chonos Archipelago and the Taitao Peninsula, as mapped by Fuenzalida and Etchart (1975), were defined as the Chonos Metamorphic Complex by Hervé *et al.* (1984). Miller (1975, 1979) distinguished three formations in this area separated by structural unconformities and thus of different ages: from older to younger these are the Canal King, Potranca and Canal Perez Sur formations. A fossil assemblage found in the intermediate Potranca Formation, assigned to Late Silurian to Early Devonian by Miller and Sprechmann (1978) was the basis for the ages given to each formation. The concept of the Chonos Metamorphic Complex (CMC) employed here is not a lithostratigraphic one, but views the ensemble of rocks as a tectono-metamorphic unit in the sense of Trouw *et al.* (1998).

The Chonos Metamorphic Complex comprises two belts of different lithology and structure (Fig. 4). An eastern belt in which the rocks preserved their original sedimentary features grades into a western belt in which the primary features of the rocks were lost by increased deformation and metamorphism. The eastern belt is composed of a meta-turbidite sequence with minor metachert and greenstone. These rocks form coherent bodies with intervening bands of broken formation, in a pattern that was mapped by Garrido (1987) and Hormazábal (1991) within restricted areas of the complex. The metamorphic grade is the sub-greenschist facies, as indicated by the illite cristallinity values in pelite, and by the presence of a pumpellyite-actinolite-stilpnomelane assemblage in the Italia Island-Dring Island greenstone rocks, in an area transitional to the western belt (Hervé *et al.*, 1994). The western belt is composed of psammo-pelitic schist, greenschist and metachert. The metamorphic grade is mainly in the greenschist facies, but blue amphibole has been found locally (Fuenzalida and Etchart, 1975). Metamorphic conditions were established at 5.5 kbar and 250 - 300 °C and 8 - 10 kbar and 480 °C for the eastern and western belts, respectively, by Willner *et al.* (in press).

Recently, Fang *et al.* (1998) have identified Late Triassic fauna in the same locality where Miller and Sprechmann (1978) collected the supposedly Late Silurian-Early Devonian brachiopods. This indicates that marine sedimentation was taking place at the Chonos region during the Late Triassic, so at least part, perhaps most, of the accretionary complex can no longer be considered as Paleozoic. The age of the subsequent metamorphism of the Chonos Metamorphic Complex is not known, but a younger limit is given by the Early Cretaceous age of the western part of the North Patagonian Batholith (Pankhurst *et al.*, 1999a). Rb/Sr whole-rock errorchrons for the Chonos Metamorphic Complex (Davidson *et al.*, 1987; Hervé *et al.*, 1988) have often given ages in the range corresponding to Permian to Jurassic, which have traditionally been interpreted as caused by metamorphic resetting. However, in view of the current conclusion that the deposition of these rocks is much younger than was previously considered, it is possible now to see that some of these results, at least, may refer to diagenesis or low-grade metamorphism very soon after sedimentation.

The Madre de Dios accreted terranes

The stratigraphy and structure of the Madre de Dios Archipelago (49 °S to 52 °S), was studied by Forsythe and Mpodozis (1979, 1983) who distinguished: a) The Denaro Complex (DC) of metabasalt and chert, b) The Tarlton Limestone, and c) The Duque de York Complex (DYC), of greywacke, siltstone, shale and conglomerate. These units are intruded by the South Patagonian Batholith (SPB).

The coeval Late Carboniferous - Earliest Permian Tarlton Limestone (Douglass and Nestell, 1976) and Denaro Complex, were deposited, according to Ling *et al.* (1985), in a mid-ocean ridge environment and accreted to the Gondwana margin as allochthonous and exotic terranes before the Early Cretaceous intrusion of the South Patagonian Batholith. In this context, Forsythe and Mpodozis (1983) considered the Duque de York Complex as a continent-derived detrital sequence unconformably deposited over the other stratified units, as they came close to the continent. The three units were then tectonically interleaved into the fore arc of the South American Plate by subduction processes.

The Denaro Complex is composed of metabasalt, often with pillow structures, and chert. The geochemistry of 10 samples of metabasalt of the Denaro Complex indicates an E-MORB affinity; broadly consistent with the tectonic setting suggested above. The samples have well preserved igneous textures, relict igneous clinopyroxene phenocrysts and microcrysts, completely albitized plagioclase, and a metamorphic overprint that includes chlorite, epidote, amphibole, pumpellyite, white mica, stilpnomelane and calcite. This suggests a low-T, probably low-P, static metamorphism, which might have taken place in the mid ocean ridge where the rocks were erupted.

Deposition of the Duque de York Complex is stratigraphically restricted between the Tarlton Limestone (Early Permian) and the Early Cretaceous intrusion of the South Patagonian Batholith. A palaeo-karst surface, tectonically inverted, unconformably separates the Duque de York Complex from the Tarlton Limestone at Seno Soplador. This erosional surface must have developed at or near sea level, or by dissolution at the Carbonate Compensation Depth.

New detrital zircon U/Pb SHRIMP age determinations in two metasandstone units from the Duque de York Complex reveal a prominent Late Early Permian population (*c.* 270 Ma) of zircons of igneous derivation, together with Neoproterozoic to Early Paleozoic components (Hervé *et al.*, 1999a). This occurrence of Late Permian zircons can probably be interpreted as the first flush of detritus from a new igneous complex being developed at the continental margin, which might correspond to the widespread Permian Choyoi magmatic event in south-western South America and the Late Carboniferous Permian granites in extra-Andean Patagonia. This interpretation implies that the continent-derived sediments of the Duque de York Complex, presumably near to the continental margin, covered the oceanic assemblages as early as Late Permian times.

The same units can also be found at the Diego de Almagro Archipelago. However, in this locality, the Denaro Complex has interbedded limestone units some metres thick, which are isoclinally folded in outcrop scale. Also, towards the SW, a blueschist metamorphic complex that includes coarse-grained ultramafic rocks, here referred to as the Almagro Complex, is in tectonic contact with the other units along the mylonites of the Arcabuz Shear Zone. The geology of these remote islands was first described by Cecioni (1956) and the presence of blueschist rocks recorded by Forsythe *et al.* (1981). The latter authors identified a shear zone separating the units mentioned previously from the rocks of the south-western part of Diego de Almagro Island, composed of highly deformed and metamorphosed blueschist, greenschist, micaschist, ultramafic rocks and metachert. This shear zone will be called the Seno Arcabuz Shear Zone (SASZ), and the blueschist-bearing unit will be called here the Diego de Almagro Metamorphic Complex (DAMC).

The Diego de Almagro Metamorphic Complex blueschist rocks are metabasite of E- MORB geochemical affinities. They display strong foliation indicating a complex polyphase metamorphic history. The rocks of the Seno Arcabuz Shear Zone are associated with a late ductile deformational event, which resulted in a strong linear fabric plunging NNW. Metamorphic grade increases substantially from N to S along the western shore of Seno Arcabuz, where the rocks of this shear zone are exposed. The blueschist rocks are only exposed near the southern end of Diego de Almagro Island.

Seven samples from the main geological units of the Diego de Almagro area were analysed by different methods to constrain the timing of the main geological events recorded in the area (Hervé *et al.*, 1999b). SHRIMP U/Pb dating of detrital zircons from a quartz- rich micaschist belonging to the Diego de Almagro Complex, and occurring at the western end of Caleta Diego de Almagro reveals detrital zircons as young as 157 ± 2 Ma, constraining the maximum age of sedimentation of the protolith to Middle Jurassic. In contrast, a sample of a metasandstone from the Duque de York Complex, has detrital zircons no younger than 265 Ma, Late Early Permian. K/Ar analysis of blue amphibole in blueschist collected near the southern end of Diego de Almagro (samples collected by C. Mpodozis) indicate Early Cretaceous metamorphic (cooling?) ages of 122 ± 21 Ma and 117 ± 11 Ma, whereas biotite from a tonalite of the South Patagonian Batholith, 30 km SE of Diego de Almagro, gave an age of 134 ± 3 Ma. $^{40}Ar/^{39}Ar$ ages in a white mica fish from a mylonite of granitic protolith from the Seno Arcabuz Shear Zone are 136 ± 8 to 120 ± 8 Ma, while masses of smaller, probably recrystallized white mica aggregates, are 89 ± 8 Ma. These data indicate conclusively that the Diego de Almagro blueschist evolved during the Mesozoic. The presence of igneous detrital zircons 157 Ma old, suggest that the protolith of the schist was coeval with the widespread Tobífera acid igneous event which was taking place in Patagonia, within the South America Plate to the E, during the Middle to Late Jurassic (Pankhurst *et al.*, 1998). This protolith is distinctly younger than the sedimentation age of the Duque de York Complex which does not record the presence of igneous rocks in the source area younger than Early Permian.

The Early Cretaceous K/Ar ages of the blue amphibole are probably related to the main deformational and metamorphic event that produced the analysed minerals in the subduction zone. This probably occurred at a depth of more than 20 km contemporaneously with intrusion of the South Patagonian Batholith. The Late Cretaceous ages obtained from mylonite formed in the Seno Arcabuz Shear Zone are probably related to the early stages of exhumation of the blueschist-bearing complex, which cooled below the apatite annealing temperature only in the late Miocene.

The Seno Arcabuz Shear Zone separates rock units of very different depositional and metamorphic histories, and is an important tectonic discontinuity, which allowed a differential exhumation path for the blueschist-bearing Diego de Almagro Complex to the W, and the previously accreted units to the E. The rocks of the Diego de Almagro Metamorphic Complex have younger (Early Cretaceous) metamorphic ages than the Middle Jurassic ones of the tectonic melange, also with blueschist mineralogy, described by Wilson *et al.* (1989) from the Diego Ramirez Islands, the southernmost exposures of the South America fore-arc complexes. The 10 Ma apatite fission track age may be related to the collision of a segment of the Chile Rise that took place in this area 13 - 14 Ma ago.

The eastern Paleozoic mud pile

To the E of the Patagonian Batholith, pre-Late Jurassic low-grade metamorphic rocks crop out between Lake General Carrera and the northern reaches of the South Patagonia Icefield. Hervé (1993) has referred to this unit as the Eastern Andes Metamorphic Complex (EAMC). Farther to the S, at Cordillera Sarmiento (51 °S to 52 °S), similar rocks have been termed the Staines Complex (Forsythe and Allen, 1980; Allen, 1982). The correlation between these two units is uncertain, although physically the continuity of exposures of metamorphic rocks occurs between both areas in a narrow fringe W of the South Patagonian Icefield.

Lagally (1975) distinguished two formations (units) in the Eastern Andes Metamorphic Complex: the northern Lake General Carrera unit composed of micaschist, greenschist and marble, more deformed and metamorphosed that the southern Cochrane unit mainly composed of a sub-greenschist to greenschist facies sequence of greywacke, shale and minor conglomerate, with very rare and minor greenschist and marble. The Cochrane unit correlates with

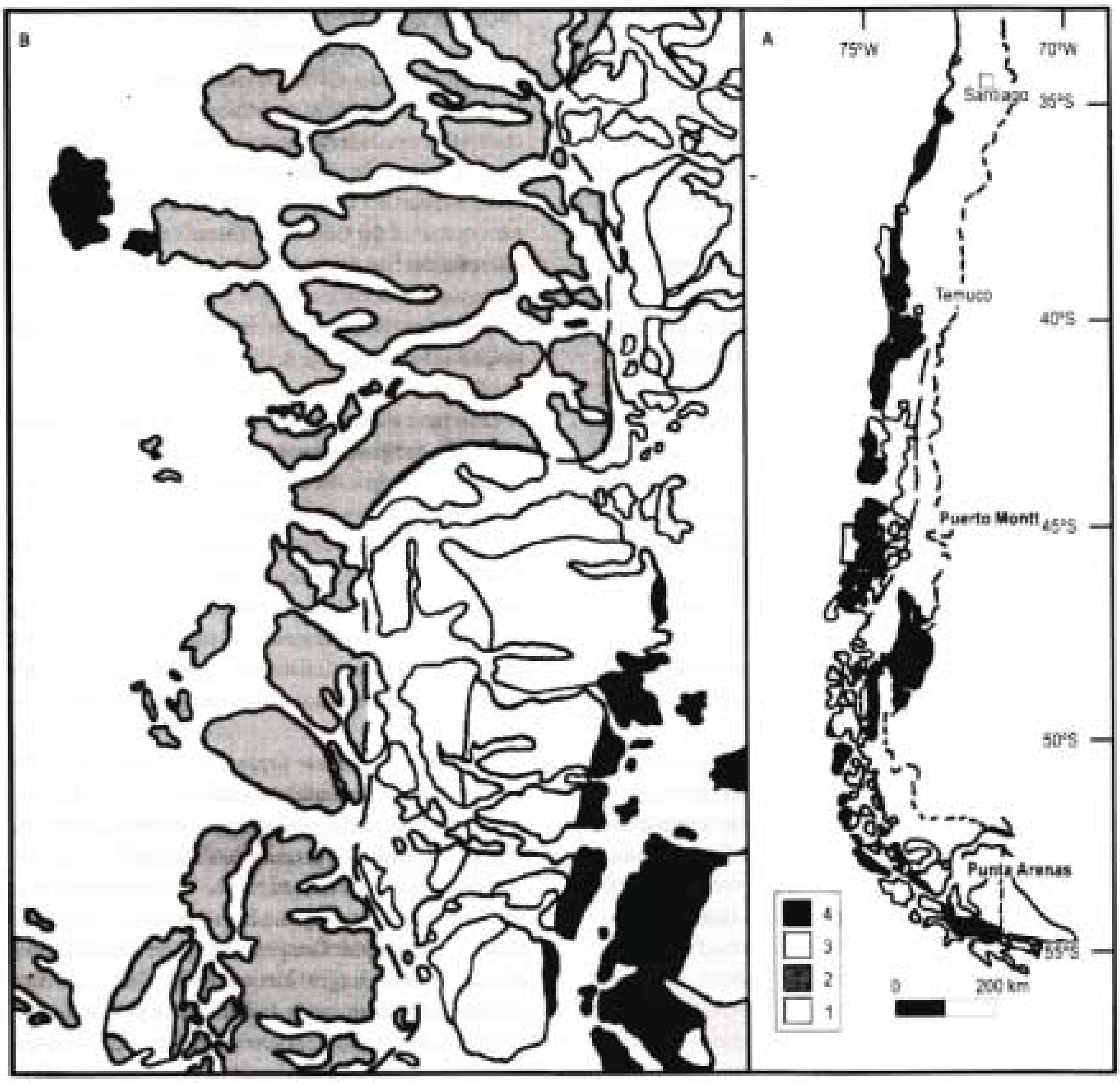


FIGURE 4 - (A) Distribution of the metamorphic basement in Central and Southern Chile (in black), with indication of the location of (B) the central sector of the Chonos Archipelago. 1. Eastern belt of the Chonos Metamorphic Complex (CMC). 2. Western belt of the CMC. 3. North Patagonian Batholith. 4. Traiguen Formation (mid-Cenozoic). For details see text (after Willner et al., in press).

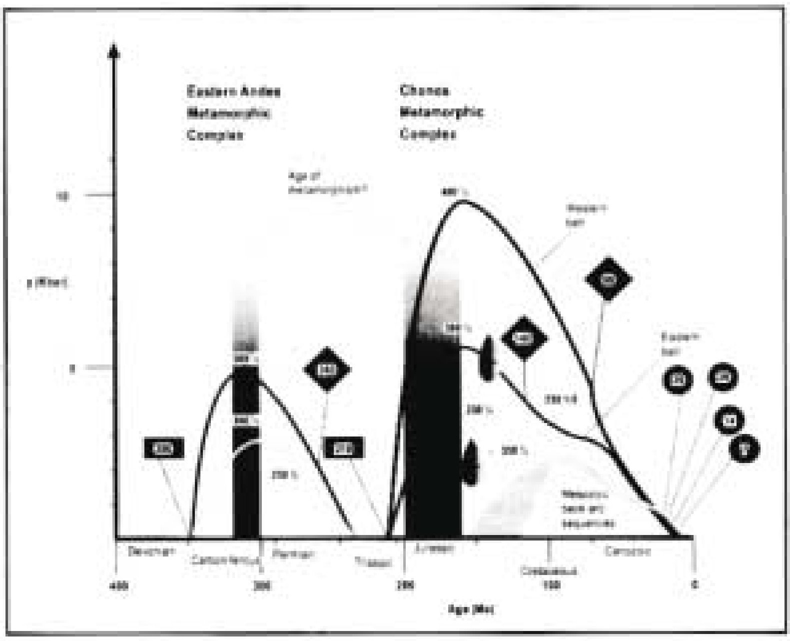


FIGURE 5 - A pressure vs. age diagram to illustrate the differences in timing of deposition and metamorphism between the Eastern Andes Metamorphic Complex in Aysén (47 °S - 48 °S) and the Chonos Metamorphic Complex. Rectangles indicate the younger age of detrital zircon in each of the units. Diamonds, indicate FT ages in zircon (S. Thomson, written communication) and the 230 °C closing temperature indicated by this system. Circles are apatite fission track ages from Thomson et al. (1999b) and Parada et al. (1997). Age of metamorphism (vertical bars) from diverse geochronological and geologic evidence. Lower gray curve at bottom right from Aguirre et al. (1999). Drops with a cross indicate initiation of plutonism in the present day back arc region (Late Jurassic) and in the fore-arc Chonos Archipelago area (Early Cretaceous).

the Río Lacteo (Leanza, 1972) and the Bahia de la Lancha (Riccardi, 1971) formations in Argentina. The latter contains fossil pollen, which has provided chronological indication of the deposition of the unit during the Late Devonian-Early Carboniferous. This is consistent with the Carboniferous age of icnites of tetrapods indicated by Riccardi and Rolleri (1980). However, it is not known if this age can be extended to the rest of the complex. SHRIMP U/Pb ages of detrital zircons are no younger than 354 Ma (Hervé, unpublished data). Unpublished Rb/Sr analysis (F. Hervé and R. Pankhurst) for schist of the Lake General Carrera unit and for metapelite of the Cochrane unit at Valle Chacabuco, yield imprecise ages of 315 Ma, which are interpreted as possible metamorphic ages. An Early Permian fission track age in zircon (Thomson *et al.*, 1999a) indicates Late Paleozoic exhumation of the complex. These possible ages of sedimentation and metamorphism do not contradict the field relationships of the unit. The Late Jurassic acid volcanic rocks of the Ibañez Formation were deposited unconformably over the Eastern Andes Metamorphic Complex, when it had already acquired its main structural and metamorphic characteristics. A summary of the evolution of the Chonos Metamorphic Complex and of the Eastern Andes Metamorphic Complex in Aysén is shown in Figure 5.

The protolith of the Eastern Andes Metamorphic Complex has been considered by several authors as a turbidite sequence (*e.g.*, Pino, 1976; Yoshida, 1981) that was deposited in a passive margin setting (Hervé *et al.*, 1998). The presence of typical alkali basalt at the Florida Peninsula within the Cochrane unit of the Eastern Andes Metamorphic Complex (Hervé *et al.*, in press) shows that eruptions occurred within the plate tectonic environment. The huge limestone masses of the Lake General Carrera unit point to a platform environment, in which a progressive upward change from marl beds to limestone was indicated by Hasegawa *et al.* (1971). The relationships of these limestone masses to those of the Madre de Dios accreted terranes is not known, but a coeval relationship of these otherwise exceptional rocks in the accretionary complexes of southern Chile is not improbable. Lacassie and Godoy (1999) have recorded the presence of structures developed during three deformational events. Basin and dome structures of the stratification being formed during the second deformational event.

Tectonic history of the Gondwana margin

The new evidence presented above, suggests a different tectonic scenario from that proposed in the early 1980's (*e.g.*, Hervé *et al.*, 1981; Forsythe, 1982; Forsythe and Mpodozis, 1983). The oldest rocks in the coastal subduction complexes are the Tarlton Limestone of Early Permian age, and the coeval Denaro Complex (Ling and Forsythe, 1987). Their origin is undoubtedly oceanic, but the E-MORB chemical signature of the Denaro Basalt opens the possibility that they were generated on an oceanic plateau or aseismic ridge rather than at an oceanic ridge. The probable Early Permian age of the Duque de York Complex limits the drifting time of the Tarlton Limestone and of the Denaro Complex to a maximum of 20 or 30 Ma. This probably excludes an exotic origin from the area of the north-western Pacific as was suggested by Ozawa and Kanmera (1984).

Late Triassic sedimentation and accretion (?) was taking place in the Chonos Metamorphic Complex to the N, and in the Diego de Almagro Complex (Late Jurassic) to the S. High-pressure metamorphism appears to have taken place in the Jurassic and Early Cretaceous, and exhumation of the subduction complex started in the Cretaceous and persisted until the Cenozoic. There is no evidence of an active subduction zone in the Paleozoic, given the very low grade non-deformational metamorphism present in the Paleozoic units, as well as the absence of a regional magmatic belt of this age in the region.

Recent studies of the present-day continental margin of Chile have allowed the formulation of a tectonic model of periodic and discontinuous accretion and tectonic erosion along the margin. The subduction of mid-ocean ridges, seamount chains or prominent fracture zones generate tectonic erosion, whereas the ponding of thick sedimentary infill in the trench promotes accretion. These situations will very probably lead to an irregular and asystematic generation and preservation of rock bodies along the margin. The discontinuous and patchy record of ages observed on the fore-arc complex may well be explained in a similar way.

The arrival of linear topographic elevations of the oceanic crust, or of seamounts, to the subduction zone, will favour tectonic erosion, which implies deep transport of the sedimentary wedge of the continental rise and their tectonic mixing with the oceanic material. This environment is probably represented by the high P western belt of the Chonos Metamorphic Complex and by the blueshist rocks of Diego de Almagro Metamorphic Complex. The relationship of these deeply underthrusted units with their backstops, is with units that were variably transported down the subduction zone; quite deeply as the eastern belt of the Chonos Archipelago, or quite shallowly as the Tarlton Limestone-Denaro Complex-Duque de York ensemble.

The Cretaceous and Cenozoic subduction below the continental margin, probably contributed further to the patchy preservation of the Late Paleozoic, Triassic and Jurassic units. It is difficult at present to evaluate the transport of blocks along the margin.

The Mesozoic Aysén Domain: a back-arc system

The Austral Basin of southernmost South America developed on Paleozoic continental crust in a back arc setting between Mid to Late Jurassic and Tertiary times. The basin extended for about 1400 km from at least the region of Futaleufu (43°30'S) in the N to southernmost South America (56 °S) in the S, and 500 km from W to E. The predominantly clastic sedimentary fill reaches a maximum of 7000 m in the S. It has a complex tectonic history related to the interaction of a subducting Pacific Plate and the active continental margin. The Austral Basin can be considered in terms of two domains or sub-basins with different geological histories. In the S the Magallanes Domain and in the N the Aysén Domain (Bell *et al.*, submitted) the boundary between these domains is approximately located at latitude 50 °S.

The general lithostratigraphy of the Mesozoic Aysén Domain

The principal stratified units are represented in the geological map of Niemeyer *et al.* (1984). More recently, new geological mapping in the eastern part of Aysén, between latitudes 43°30'S and 48°S, has produced the stratigraphic column presented herein (De La Cruz *et al.*, 1996; Suárez *et al.*, 1996), that from base to top includes the following units:

-Ibáñez Group, mainly subaerial acid volcanic rocks dated as Upper Jurassic (Niemeyer *et al.*, 1984; Suárez and De La Cruz, 1996; Suárez *et al.*, 1996) to Berriasian (Covacevich *et al.*, 1994). These rocks have a calc-alkaline affinity and represent subduction-related volcanism (Suárez *et al.*, in press).

- Coyhaique Group, representing the marine sedimentary infill of the Aysén Basin and comprising the Toqui, Katterfeld and Apeleg formations (Suárez *et al.*, 1996).

- Divisadero Group (Heim, 1940), subaerial volcanic rocks of latest Lower Cretaceous age.

- Alto Coyhaique Volcanic Association including the Casa de Piedra Volcanic Complex, consisting of rhyolitic and dacitic domes; the Morro Negro Basalt and the Laguna del Toro Volcanic Complex, consisting of basalt, andesite and dacite (Suárez *et al.*,1996).

The geographical extension of the Aysén Domain

The lithostratigraphy of the Aysén Basin has been identified as far N as in the area of Futaleufú (43°30'S; De La Cruz *et al.*, 1996) and as far S as N of Lake Viedma, in Argentina (50 °S). The basic stratigraphy of this basin has been followed from Aysén southwards up to the N of Lake Viedma, and the following correlations have been made:

- The Toqui Formation has been correlated with the Springhill Formation.

- The Katterfeld Formation with the Rio Mayer Formation extending N of Lake Viedma. The Rio Mayer Formation exposed from Lake Argentino to the S has a different facies comprising turbidites.

- The Apeleg Formation is equivalent to the Rio Belgrano Formation exposed to the S.

- The Divisadero Group is represented further S by the Rio Tarde and Kachaike formations, the latter with very distal volcaniclastic facies N of Lake Viedma. To the E, the Divisadero Group is represented by the more distal volcaniclastic succession of the Chubut Group.

The transgression (Toqui Formation): Tithonian-Berriasian

The transgression of the Aysén Basin took place diachronously during the Tithonian (De La Cruz *et al.*, 1996) and Berriasian (Niemeyer *et al.*, 1984; Covacevich *et al.*, 1994). These deposits include a series of different shallow marine sedimentary facies with interbedded pyroclastic beds grouped in the Toqui Formation (Suárez and De La Cruz, 1996; Suárez *et al.*, 1996). It has a restricted exposure in the area of the Toqui Mine and in the area of Palena (De La Cruz *et al.*, 1996). To the W and S of Chile Chico, the transgressive deposits have been included in the Cerro Colorado Formation (Suárez and De La Cruz, 1996). These beds contain rare Lower Cretaceous marine fossils and were formerly included in the "Primer Nivel Marino con Ostrea" assigned to the Upper Cretaceous-Paleocene (Charrier *et al.*, 1979; Niemeyer *et al.*, 1984).

The Toqui Formation, 30 to 215 m thick, in the Toqui Mining District consists of green volcaniclastic sandstone units, locally with fossil shell fragments. Many of these units have been interpreted as turbidites. The Toqui Formation also includes beds of green, well-sorted lapilli tuff less than 5 cm thick, interpreted as air-fall deposits. In additon there are tuffaceous sandstone and grey pelite beds; rare ignimbrite flows, possibly of submarine deposition; debris-flow deposits; and oyster beds, 8 to 15 m thick, mainly at the base, which are the mineralized zones (lead-zinc, interpreted as a skarn) occurring at the mine (Suárez *et al.*,1996). To the E of the Toqui Mine, and adjacent to the border with Argentina, outcrops of this formation at Estero La Horqueta, contain ammonites of Upper Berriasian age (Covacevich *et al.*, 1994). In the area to the S of Laguna Foitzick, S of Coyhaique, the Toqui Formation overlies the Ibáñez Group, and locally a dacitic peperite was emplaced along the contact. At this locality this formation includes calcareous beds (30-40 m thick), locally with dacitic fragments, probably of pyroclastic origin, and transgressive conglomerate at the base. The limestone units contain fragments of corals, echinoderms, bryozoa, oysters and algae.

Although no diagnostic fossils were collected from these outcrops, they include a faunal association comparable to that with Berriasian ammonites occurring at Estero La Horqueta (Suárez *et al.*, 1996). In the area of Palena, this formation has a minimum thickness of 380 m, and comprises a marine succession of fossiliferous volcaniclastic sandstone, tuffite and shale beds, with oyster beds at the base and oncolitic shales at the top. It has intercalated ash and lapilli tuff, some representing ignimbrite deposits (15 m thick) with marine fossils at the base, intercalated between volcanic sandstone with marine fossils. In Cerro Redondo, SSW of Futaleufú (43°21'S), shale beds with flattened perisphinctides, of Upper Jurassic age, crop out. Previously these beds were assigned to the Middle Jurassic (Tres Monjas Member of the Tamango Formation). In Cerro Campana, S of Palena (43°44'S), ammonites of Tithonian age have been found in these rocks that were previously included in the Alto Palena Formation and assigned to the Lower Cretaceous. In the area of Arroyo Culebra, near Palena, Berriasian fossils have been reported. Consequently, in the area of Palena-Futaleufú, the Toqui Formation has Tithonian and Berriasian zones (De La Cruz *et al.*, 1996).

Therefore, the Toqui Formation represents shallow marine deposits, at least occasionally in a restricted environment and occasionally subject to high energy capable of destroying the shells (storms?). Deposition was contemporaneous with explosive volcanism that generated ignimbrites and ash-fall deposits. The Toqui Formation was deposited in shallow marine basins adjacent to coastal volcanoes, probably calderas or even volcanic islands comparable to Krakatau, in an intra-arc setting.

To the S and W of Chile Chico, the earliest marine deposits overlying the Ibáñez Group are represented by the Cerro Colorado Formation (Suárez and De La Cruz, 1996). The Cerro Colorado Formation underlies subaerial tuff and tuffite of the Estero de los Flamencos Tuff of Barremian age, and with an angular and erosional unconformity; the Ligorio Márquez Formation (upper Paleocene-lower Eocene), and Eocene basalt flows with an erosional unconformity. A rhyolitic peperite occurs along part of the basal contact with the Ibáñez Group. *Steinmanella sp.*, collected from these beds, indicates a Lower Cretaceous age for beds (V. Covacevich, 1996, written comm., in Suárez and De La Cruz, 1996). This formation is a minimum of 100 m thick and contains several different facies associations: quartz-rich basal sandstone and conglomerate beds (8 m thick), with fossil flora and coal laminae, locally with abundant trace fossils; quartz-rich sandstone with ostrea (10-20 m thick), overlying the basal sandstone; bioturbated green sandstone with *Steinmanella sp.* (15 m minimum thickness); cross-bedded heterolithic sandstone with mud-chips and silt foresets, locally with abundant trace fossils, fossil wood and some coal laminae. These cross-bedded heterolithic sandstones overlie a 40 to 50 m thick succession of poorly exposed shale beds that, in turn, overlie the basal quartz-sandstone beds. These deposits represent a shallow marine environment and probably some tidal deposits. An abrupt lateral facies change occurs at about 20 km to the W, where the Katterfeld Formation directly overlies the Ibáñez Group.

Black pelite deposition (Katterfeld Formation): Valanginian-Hauterivian-(?)Barremian

A marked facies change represented by marine black shale of the Katterfeld Formation, equivalent to the Rio Mayer Formation in Argentina (Riccardi, 1971), succeeded the transgressive Toqui Formation, implying a deepening of the basin. Marine black shale conformably overlies the Toqui Formation, and locally these directly overlie the Ibáñez Group. The Katterfeld Formation was defined in Argentina. They are predominantly black shale beds, massive, and locally with parallel lamination, and with rare distal turbidite intercalations of fine-grained sandstone beds (up to 40 cm thick). They contain ammonites, belemnites, *Gryphaea*, *Entolium*, *Chlamys*, *Oxitoma*, trigonias, gastropods, fish teeth, and plant fragments. *Favrella sp.* indicates an Hauterivian age, and the occurrence of a thick pelitic succession underlying the *Favrella*-bearing zone suggests a Valanginian age. They represent deposits in a partially protected marine environment with normal salinity, calm, anoxic conditions where sedimentation was mainly by settling with local input of turbidity currents. However, locally, the occurrence of oriented belemnites and overturned gryphaeas suggest occasional strong currents (Suárez *et al.*, 1996).

Shelf sandstones (Apeleg Formation): Hauterivian to Lower Aptian

In Hauterivian-Barremian times the Austral Basin underwent a major reorganization, with the Aysén Basin in the N differing from the Magallanes Basin in the S. In the Aysén Basin the black shales of the Katterfeld and Rio Mayer formations, deposited in an anoxic restricted environment, were conformably overlain by open marine sediments of the Apeleg and Rio Belgrano formations, which were deposited in a tide-dominated, epicontinental seaway (Riccardi, 1971, 1988; Bell and Suárez, 1997). In the Magallanes Basin, deposition of the regionally extensive blanket of dilute turbidites continued unchanged. The Apeleg Formation was defined between 44°30'S and 45 °S in Argentina, and its extension into the Aysén Region was recognised by Suárez and De La Cruz (1996).

The formation at Aysén consists of 1200 m of well-sorted sandstone and mudstone beds with mud-draped ripples and a well-preserved and varied trace fossil assemblage. The fossil assemblage includes *Favrella sp.*, *Favrella americana*, *Pterotrigonia*, *Steinmanella gr. Steinmanella herzogi-holubi*, which indicates a Valanginian-Hauterivian-(?)Barremian age. Shark fossil teeth have also been reported. To the S of Lake General Carrera (46°30'S), Lower Aptian fossils have been found (Suárez and De La Cruz, 1996; Covacevich *et al.*, 1994) near the base of the formation in the area of Cerro Bayo. Hauterivian and Lower Aptian ammonites have been recorded S of Lake General Carrera at about 46°45'S (Suárez and De La Cruz, 1996). Abundant trace fossils characteristic of a high energy marine environment are common, including *Gyrochorte*, *Lockeia*, *Ophiomorpha*, *Asteriacites*, *Chondrites*, *Thalassinoides*, *Arenicolites*, *Skolithos?*, *Planolites?*, *Helminthoides*, *Diplocraterion*, *Rhizocorallium*. The deposits were mainly accumulated as offshore tidal sandbars or sand ridges on a shallow marine shelf (Bell and Suárez, 1997).

Lower Cretaceous volcanism

Tuff cones, locally with preserved diatremes, were emplaced in the Apeleg Formation in the area of Baño Nuevo (Baño Nuevo Volcanic Complex; De La Cruz *et al.*, 1996; Suárez *et al.*, 1996). A diatreme composed of abundant fragments of Apeleg sandstone and basaltic juveniles contains a large fragment of the Apeleg sandstone, with diffuse margins, as well as basaltic fragments, suggesting that the sandstone fragment was incorporated in the volcanic conduit while still humid. This would indicate that the tuff cones were emplaced shortly after the Apeleg shallow sea retreated from the area, and that the tuff cones would be of Hauterivian-Barremian age. Adjacent basaltic flows (Loncomahuida Basalt) overlying ignimbrite flows, which have given a K/Ar whole-rock age of 135 ± 5 Ma (unpublished) may be related to the tuff cones. The radiometric value corresponds to the Valanginian-Hauterivian, an age older than the Apeleg Formation. If the correlation between the tuff cones and the basaltic lava is correct, then it is possible that the radiometric date is reflecting excess Ar. Basaltic and andesitic plugs are also emplaced in the Apeleg Formation. This volcanic complex represents a Surtseyan-type of eruption.

A tuff succession of green and reddish ignimbrites, tuffites and ash-fall deposits, exposed 25 km S of Chile Chico (Estero de Los Flamencos), overlies Lower Cretaceous marine deposits of the Cerro Colorado Formation and is unconformably overlain by the Ligorio Márquez Formation (upper Paleocene-lower Eocene). Rootlets present in the tuffite beds indicate sub-aerial deposition. Ignimbrite flows have yielded K/Ar dates (biotite) of 128 ± 3, 125 ± 4, and 123 ± 3 Ma, and considering its stratigraphic position, overlying Lower Cretaceous strata, these dates may represent near-eruption ages (Barremian). Twenty kilometres to the W, marine conditions prevailed up to the Lower Aptian (Suárez and De La Cruz, 1996). Therefore, these sub-aerial tuff beds were either deposits of an active volcanic island or deposits of a coastal volcano adjacent to the Austral Basin.

Weak Mid-Cretaceous tectonism and uplift: disappearance of the marine Aysén Basin

Locally, an angular unconformity exists between the Apeleg Formation and the overlying volcanic deposits of the Divisadero Group and the Upper Cretaceous basaltic lava, in adjacent Argentina. North of Villa Ortega, and approximately 30 km N of Coyhaique, steeply dipping strata of the Apeleg Formation are unconformably overlain by subhorizontal beds of the the Divisadero Group. In other localities rocks of the Divisadero Group have given K/Ar dates of 116 to 100 Ma, and in neighbouring exposures, the Apeleg Formation contains Hauterivian ammonites (*Favrella*; Suárez *et al.*, 1996; Bell and Suárez, 1997). Therefore, the timing of the deformation is confined between the Hauterivian and the Aptian. In the area W of Alto Rio Senguerr, in neighbouring Argentina, sub-horizontal basaltic flows that gave a K/Ar whole-rock date of 60 ± 2 Ma (unpublished) unconformably overlie weakly folded Apeleg sandstone beds, indicating a late Lower to early Upper Cretaceous age for a weak tectonic episode predating the widespread eruption of the Divisadero Group.

Aptian-Albian subaerial volcanism: Divisadero Group

Widespread sub-aerial volcanism of the Divisadero Group (Heim, 1940; Niemeyer *et al.*, 1984; Suárez *et al.*, 1996), mainly acid and explosive, occurred in the Aysén Domain during the Aptian-Albian. It is mainly composed of ignimbrite flows of a calc-alkaline affinity.

Turonian-Senonian calc-alkaline sub-aerial volcanism

Remnants of Upper Cretaceous acid domes, basaltic lava and basaltic to dacitic lava and plugs, with calc-alkaline affinities and radiometric dates of Turonian-Senonian are exposed E of Coyhaique, indicating that subduction processes continued during this time.

Casa de Piedra Volcanic Complex

Rhyolitic and dacitic domes, exposed 40 km E of Coyhaique have given K/Ar dates ranging from 81 ± 2 to 75 ± 1 Ma. One of these domes was emplaced in an ignimbrite of the Divisadero Group that gave a K/Ar date of 116 ± 4 Ma (Suárez *et al.*, 1996).

Morro Negro Basalts

A succession of basaltic lava flows exposed in the area of Morro Negro, on the Chile-Argentina border, E of Coyhaique, have given K/Ar dates of 76.1 ± 5.3 and 64.2 ± 2.6 Ma and may be equivalent to those to the W of Alto Rio Senguerr that gave K/Ar dates ranging between 79.1 ± 4.7 and 73.6 ± 5.4 Ma (Butler *et al.*, 1991).

Laguna del Toro Volcanic Complex

Volcanic breccias, lava flows, plugs and dykes of basaltic and andesitic composition, exposed 25 km E of Coyhaique, and overlying the Divisadero Group with a slight angular unconformity, have a calc-alkaline affinity indicative of a subduction-related origin. These rocks yielded K/Ar whole-rock dates of 75 ± 3 to 67 ± 3 Ma (Suárez *et al.*, 1996).

All the rock units in the Aysén Basin were affected by very low to low grade metamorphism in the zeolite, prehnite-pumpellyite and greenschist facies (Aguirre *et al.*, 1999). Differences in grade are related to the age of the rock sequences, with the younger metamorphosed in the zeolite facies and the older (Jurassic) in greenschist facies.

The Mid Tertiary Traiguen Basin: a fore-arc extensional basin

Exposed in the islands within and around the Moraleda Canal, which in Aysén represents the Central Valley, there is a volcano-sedimentary sequence, which has been denominated the Traiguen Formation. It consists of a marine sequence of predominantly volcanic breccia, which includes thick zones of greywacke, shale, turbiditic beds, and abundant flows of pillow basalt. It was in part deposited unconformably over metamorphic rocks, and in its western border over the Early Cretaceous components of the North Patagonian Batholith. It is intruded by the Miocene (20 Ma) components of the Northern Patagonian Batholith at Magdalena Island, and it hosts a basic dyke swarm of mainly early Miocene age.

The pillow basalt flows are geochemically intermediate between alkali basalt and subduction-related basalt (Hervé *et al.*, 1995). They were metamorphosed in a low-pressure environment, varying from zeolite facies with prehnite and pumpellyite in the southern areas to greenschist facies in the northern areas, in like maner to the ocean floor type metamorphism, without pervasive deformation. However, the sequence is folded in N-S trending ample folds, and was overthrust by its basement during basin inversion in some areas. The folding took place before the intrusion of the 20 Ma granitic rocks.

This unit was deposited in an extensional basin, spatially related to the Liquiñe-Ofqui dextral strike slip fault system. Extension thinned the continental crust, allowed the extrusion of the mafic rocks, and provided a high heat flow environment revealed by the metamorphic mineral assemblages.

The age of deposition of this sequence has not been determined with precision. Miocene foraminifera were mentioned as occurring on Traiguen Island by Niemeyer *et al.* (1984). Rb/Sr geochronology on pelitic rocks indicated ages between 40 and 20 Ma (Hervé *et al.*, 1995). However, as no stratigraphic column of the unit has been established, it is not known if all the strata were deposited at these times. Similar rock sequences have been recently identified in the surroundings of Estrecho Concepción (49 °S), but the relationship with Traiguen Formation is not known.

The basic dyke swarm is spatially related to the Traiguen Formation, but its limits extend beyond the present outcrops of the unit into the surrounding North Patagonian Batholith. Hervé *et al.* (1996b) established that the main trend of the dykes is N30°E, and they related this extension to oblique subduction of the Pacific Ocean crust during the period 40 to 25 Ma, as indicated by Cande and Leslie (1986).

At Melchor Island volcanic domes with glassy and brecciated margins, and well-developed divergent prismatic disjunctions are present near the western margin of the Traiguén Basin. These lavas are mostly porphyritic with a phenocryst assemblage composed of zoned plagioclases with anorthite to bytownite cores and labradorite rims, augite, orthopyroxene frequently altered to chlorite-smectite, and titanomagnetite. Such a mineral association is typical of calc-alkaline rocks. The chemical analyses classified these lavas as high-silica andesite to dacite (58 - 63% silica) belonging to a low-K series. Such a signature is in accordance with their tectonic position, in a fore-arc basin.

In the region of the Gulf of Penas (46 °S - 48 °S), and to the W of the southern end of the Liquiñe-Ofqui Fault Zone, several thick units of massive sedimentary rocks are exposed. According to Diemer *et al.* (1997), the oldest unit is of Early Cretaceous age and crops out at Puerto Barroso, in the southern part of the Taitao Peninsula. Eocene beds are known from Good Island, S of the Gulf of Penas, and Miocene beds from Chaicayan Island, N of the Gulf of Penas. These sequences are probably related to extensional accommodation of the fore-arc region to strike-slip movements along the Liquiñe-Ofqui Fault Zone System, although no Early Cretaceous activity has been otherwise observed.

Cenozoic back-arc basins

Two different sedimentary basins have been proposed for the Cenozoic in eastern central Patagonian Cordillera (Marshall and Salinas, 1990) currently separated by the Lake General Carrera, a feature lying transversal to the Andean trend, and probably representing a major geological boundary.

The so-called "Mayo River Embayment" to the N, and the "Magallanes Basin" to the S (not a proper name because the term has precedence when referring to the Mesozoic basin). These "basins" differ in their stratigraphy and in their subsequent deformational history, which is possibly a consequence of a different origin and evolution.

The stratigraphy of the north-eastern "Magallanes Basin" includes three different stratigraphic successions:

I) The Cosmelli Basin, in the W, is a upper Paleocene-lower Eocene to middle Miocene succession about 1000 m thick exposed in a double N-S trending monocline (15 km EW axis; 30 km NS axis):

- Ligorio Márquez Formation (Suárez *et al.*, in press), upper Paleocene-lower Eocene, fluvial deposits;
- San José Formation as defined by Flint *et al.* (1994), middle Eocene-early upper Oligocene(?), fluvial deposits;
- Guadal Formation, upper Oligocene-lower Miocene, marine deposits;
- "Galera Formation", correlated with the Santacrucian of early-middle Miocene age (*e.g.*, Marshall and Salinas, 1990), fluvial deposits.

II) Laguna Los Flamencos area (Cerro Rocoso of Ray, 1996), fluvial deposits;

- Ligorio Márquez Formation, upper Paleocene-lower Eocene; unconformably overlying Eocene basalt. A low angle unconformity, places the basalt flows in direct contact over older units to the N;
- Guadal Formation, upper Oligocene-lower Miocene;
- Basalt, Mio-Pliocene.

III) Rio Jeinimeni area:

- Guadal Formation, marine deposits;
- "Galera Formation";
- Miocene-Recent basalt.

This stratigraphic column differs from those accepted (Niemeyer *et al.*, 1984; Ray, 1996) as follows: the Chile Chico and Divisadero formations are not exposed in the Chile Chico-Cosmelli areas, as these in fact belong to the Ibáñez Group (Suárez and De La Cruz, 1996, 1997). An upper Paleocene-lower Eocene fluvial unit is defined (Ligorio Márquez Formation), and on the basis of preliminary work a middle Eocene age is assigned to the base of San José Formation. The stratigraphic unit designated by Niemeyer *et al.* (1984) as the "Primer Nivel Marino con Ostrea" and assigned an Upper Cretaceous-Paleocene age should be eliminated as in fact this interval corresponds to two formations: the Cerro Colorado (Lower Cretaceous) and Ligorio Márquez (upper Paleocene-lower Eocene). The San José Formation of Cerro Rocoso (Ray, 1996) has not been found, and includes the Cerro Colorado and Ligorio Márquez formations. Furthermore, in the Cosmelli Basin the Ligorio Márquez Formation underlies the San José Formation and these units can be seen to have different lithologies and age based on the paleoflora.

The Cenozoic stratigraphy of the south-eastern "Mayo River Embayment" starts with the Eocene basalt flows, interbedded acid tuff that unconformably overlie the Early Cretaceous Divisadero volcanics. In turn these volcanic rocks have an erosional contact with acid tuff, tuffite and fluvial deposits (>660 m), correlated with "The Galera Formation". Younger fluvial conglomerate units of Cerro Galera, may be equivalent to one of the conglomeratic formations in neighbouring Argentina: the Pedregoso Formation (middle Miocene) or the Chalia Formation (late Miocene-Pliocene) (Dal Molin and Franchi, 1996). Of the two, the latter is preferred. The "Mayo River Embayment" stratigraphy differs from that of the "Magallanes Basin", particularly for the post-Eocene sequence. However, it is not known whether the pre-Eocene history was really different, because the possibility exists that the units of the pre-Eocene "Magallanes Basin", were also deposited and later eroded, as is known to have occurred in the northwestern part of the "Magallanes Basin". Here Eocene basalt flows directly overlie the Ibáñez Group, as the Ligorio Márquez and Cerro Colorado formations and the Estero Los Flamencos tuff beds having been eroded. The main post-Eocene difference between the "Mayo River Embayment" and the southern basin is that no marine transgression followed the extrusion of the Eocene basalt flows that indicates a more elevated terrain.

Furthermore, the tectonic evolution to the N and S of Lake General Carrera seems different. The sequence in the "Magallanes Basin" is more complex than that of the flat-lying eastern "Mayo River Embayment" in which there occur monoclines related to faults (Marshall and Salinas, 1990; Ray, 1996). The tectonic evolution to the S of Lake General Carrera is complex. Thrusts and folds in the Mesozoic strata and pre-upper Paleocene-lower Eocene folds and faults have been defined (Suárez and De La Cruz, 1996, in press). Later thrusts and folds, of probable intra-Galera Formation age have been described by Flint *et al.* (1994), which indicates a middle Miocene contractional tectonism. In the same area, a double monocline has been identified, and the possible occurrence of growth faults within Galera Formation will corroborate the syn-tectonic deposition of the Galera sandstone beds (unpublished). To the S of Chile Chico, adjacent vertical and horizontal beds of the Galera Sandstone may also indicate an intra-Galera unconformity (Lagabrielle *et al.*, in press). A N-S normal fault may be observed to the N of Lake General Carrera, indicating a later shift from contractional to extensional tectonism. Active faulting reported in the Chile Chico area has been related to subduction of the Chile Ridge (Lagabrielle *et al.*, in press).

The Cenozoic tectonic history of the area to the S of Lake General Carrera comprises a poorly dated pre-late Paleocene-early Eocene and post-Early Cretaceous compressional event (Suárez and De la Cruz, 1996). Although the exact timing of this tectonism cannot be determined, it may represent one of the two main pre-Eocene deformational episodes recognised in the Patagonian Cordillera: a Middle-Cretaceous event, and a Late Cretaceous-Eocene event. In the latter case, the structural evolution of the area S of Lake General Carrera may represent one of progressive deformation with the earliest stages during the Latest Cretaceous?-Paleocene, followed by middle Miocene, intra-Galera tectonism along the orogenic front S of Chile Chico as well as along an out-of-sequence structures in the area of San José (Flint *et al.*, 1994); subsequently changing to extensional tectonism with probable active normal faulting (Lagabrielle *et al.*, in press). Accordingly, the sedimentary upper Paleocene-lower Eocene and late early to early middle Miocene basins would represent wedge-top and foreland basins. The origin of the San José depocentre is not clear, and widely extensive process may have caused the Guadal marine basin, that had a wide geographical distribution.

The Mesozoic to Cenozoic Patagonian Batholith: a subduction related magmatic arc

This batholith is a continuous body, about 1000 km long and up to 200 km wide, sub-parallel to the present coastline, and consisting of Late Jurassic to Pliocene intrusive rocks. Some satellite bodies occur to the W of the main batholith on the Taitao Peninsula, and to the E of it at Paso las Llaves (Vargas and Hervé, 1994), Puerto Sanchez, Cerro San Lorenzo (Welkner, 1999), Monte Fitz Roy and Torres del Paine (Skarmeta and Castelli, 1997). The Patagonian Batholith can be subdivided in two segments, the Northern Patagonian Batholith (NPB) N of the Gulf of Penas (47 °S), and the Southern Patagonian Batholith (SPB) to the South. The trend of the NPB is NNE and that of the SPB is NNW.

Pankhurst *et al.* (1999a) have recently described the Northern Patagonian Batholith in some detail. It is a calc-alkaline series of rocks, in which hornblende-biotite tonalite and granodiorite predominate. The age distribution of individual plutons within the batholith shows longitudinal zonation, with Early Cretaceous units near the margins (*c.* 120 Ma in the E, *c.* 135 Ma in the W), while the centre of the batholith includes Eocene, Miocene and Pliocene plutons (Fig. 6). The discrete episodes of granite emplacement were found to correspond to changes in the kinematics of the subduction regime, granitoid emplacement apparently being favoured by more orthogonal angle and/or sudden increases in the rate of subduction. The younger plutons lie near the main trace of the Liquiñe-Ofqui Fault Zone, which appears to have controlled their emplacement. They are syn-tectonic plutons, many of them showing strong magmatic and tectonic foliation. Hervé *et al.* (1996a) provided evidence based on the Al-in hornblende geobarometer that the Early Cretaceous margins were emplaced at shallower levels that those of the younger medial plutons. All the granitoid rocks of the Northern Patagonian Batholith show similar major and trace element geochemical patterns, regardless of age. However, this is not true of their isotope composition at the time of emplacement. The Early Cretaceous plutons have relatively high initial $^{87}Sr/^{86}Sr$ ratios (0.7045 - 0.7050) and significantly negative $\varepsilon_{Nd(T)}$ values, indicating the involvement of relatively mature crust, either in the source region or in magmatic processes. The Tertiary plutons, on the other hand, have lower initial $^{87}Sr/^{86}Sr$ ratios (down to 0.7038) and less negative $\varepsilon_{Nd(T)}$ values, similar to those exhibited by the Miocene to Recent volcanic rocks of the Southern Andes. Pankhurst *et al.* (1999b) argued that the Sr/Nd trends were consistent with lower or middle crustal involvement rather than contamination in upper crustal magma chambers. They proposed remelting of underplated mafic rocks in the lower crust as the main petrogenetic process: the Tertiary magmas being derived by more rapid remelting of juvenile mafic rocks very soon after underplating.

The Southern Patagonian Batholith has been studied only in some cross sections by Weaver *et al.* (1990) and by Bruce *et al.* (1991). The known ages extend from 165 Ma to 11 Ma, with a peak between 120 and 70 Ma that coincides with the Cretaceous global maximum in sea-floor spreading and the climax of Andean Orogeny. Plutons of gabbro to granite composition were intruded, mainly at mid-crustal depths. It is a calc-alkaline suite in which initial $^{87}Sr/^{86}Sr$ ratios tend to decrease with time as in the Northern Patagonian Batholith. However, the authors cited above interpreted this as a consequence of plutons being intruded into earlier plutons near the central axis of the batholith rather than into country rocks, as on the margins. This magmatic inflation of the batholith was thought to have shielded younger plutons from the radiogenic country rocks in a manner that became increasingly effective over time. This would suggest that the state of stress within the magmatic arc was neutral or extensional, at least during periods of magmatic intrusion, a conclusion supported by the general lack of deformed plutons in the batholith. Although there may be some Cenozoic bodies near the median part of the batholith, the main body of the Southern Patagonian Batholith does not have Miocene or Pliocene constituents as the Northern Patagonian Batholith does.

The western satellite body of the northern Taitao Peninsula, is a shallow level monzogranitic body of Late Cretaceous (70 Ma; U/Pb zircon, Pankhurst *et al.*, 1999b) age. There is no pluton within the main body of the batholith with this age. The eastern satellite bodies are more varied. Small plutons of Middle Jurassic age occur at El Faldeo (47°05'S), dated 155 ± 10 Ma (Parada *et al.*, 1997; U/Pb in zircon), which have rather high Sr initial ratios (0.708), indicating a significant crustal signature, which is not present in the Patagonian Batholith.

An exception to the rule is the case of the San Lorenzo

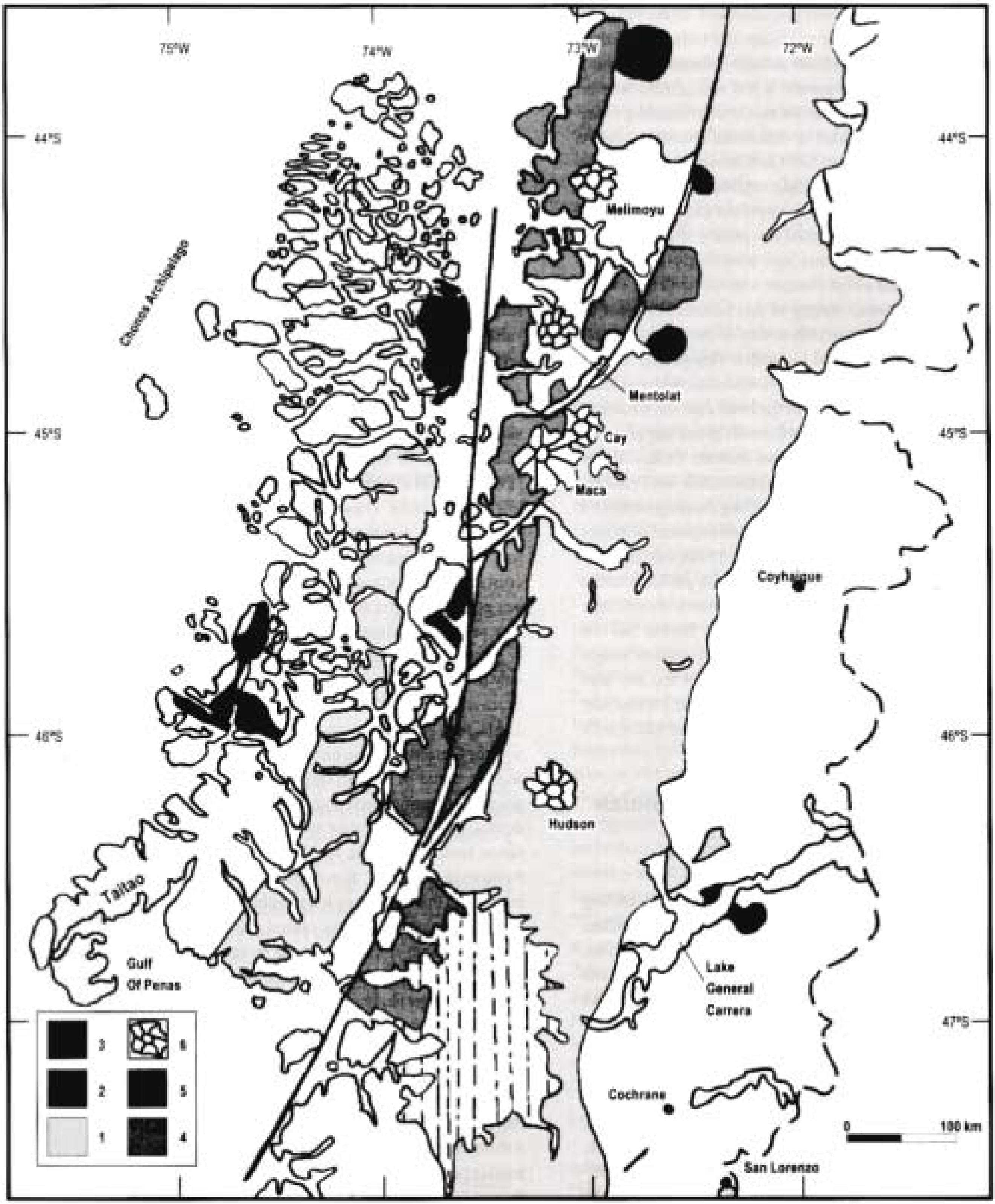


FIGURE 6 - Geological sketch map with the age distribution of plutons within the North Patagonian Batholith and its satellite plutons in Aysén. Age groups: 1. Early Cretaceous. 2. Late Cretaceous. 3. Eocene. 4. Early Miocene. 5. Late Miocene and Pliocene. 6. Stratovolcanoes of the Southern Volcanic Zone. The main branches of the Liquiñe-Ofqui Fault Zone are shown to reveal its close spatial associations with the Miocene plutons. In the back arc region, at Cerro San Lorenzo and the central portion of General Carrrera Lake, episodic plutonism from the Middle Jurassic to the Pliocene has taken place in restricted areas.

Complex. It occurs in a restricted area around Mount San Lorenzo, the second highest peak in the Patagonian Andes. Welkner (1999) described the presence of Late Jurassic, Late Cretaceous and late Miocene intrusives. U/Pb dating (Pankhurst *et al.*, in press) and that of the third phase have confirmed the age of the first phase by Rb/Sr geochronology (Pankhurst, unpublished). Within these mainly calc-alkaline bodies, there is an hyperalkaline facies of Late Cretaceous age containing riebeckite and astrophyllite.

At Puerto Cristal, a Middle Jurassic (D. Quiroz, personal communication) pluton, on the northern shore of Lake General Carrera, an Early Cretaceous (100 Ma; K/Ar, Hasegawa *et al.*, 1971) monzonite and a Late Cretaceous (70 Ma; Townley, 1996, Ar/Ar) granite occur. Near by, at Paso de las Llaves, on the southern shore of the lake, there is exposed a 10 Ma old pluton (Pankhurst *et al.*, 1999b) with miarolitic cavities that indicate a shallow emplacement (less than 3 km) (Vargas and Hervé, 1994). The Faldeo-San Lorenzo and Puerto Cristal-Paso Las Llaves are two centres where restricted plutonic activity has taken place intermittently for a long period of time.

Isolated satellite plutons constitute some of the most conspicuous mountais near the international Argentine - Chilean border. These are the Fitz Roy and the Torres del Paine plutons, where shallow calc-alkaline magmatism occurred in the Miocene.

Small plutons of Pliocene age also warrant a special mention. These are exposed in the southern part of the Taitao Peninsula, and were considered by Mpodozis *et al.*, (1985) to be related to the subduction of pieces of ocean ridge beneath the continental margin. They occur near a near-by trench and are adakitic in composition (Bourgois *et al.*, 1996).

The Late Cretaceous to Pliocene Patagonian basalts: intracontinental magmatism

In the back-arc region of the Patagonian Andes, and extending far into the Patagonian foreland, basaltic activity occurred from the Late Cretaceous to the late Cenozoic. We will refer here only to those exposures near the mountain belt in Chile.

At Morro Negro, a sequence of flat lying basalt flows was dated by K/Ar (Butler *et al.*, 1991) and an average age of 71 ± 5 Ma was obtained. A dacitic sample collected at the base of the volcanic sequence gave a slightly older Late Cretaceous age (77 Ma). The Morro Negro lava is basalt to basaltic andesite with olivine and abundant plagioclase phenocrysts. These mafic lava flows present on a Pearce (1983) spidergram, negative anomalies for Ta and Nb, and enrichment in LILE; typical of subduction-related magmas.

A subhorizontal sequence of basalt extrusives, approximately 150 m thick, is exposed in the Balmaceda Basin. At the base of the volcanic pile, the presence of pillow structures and Surtseyan-type surge deposits indicate that volcanic activity began in a subaqueous environment (Demant *et al.*, 1996). However, this episode ended rapidly, and the following eruptions correspond to sub-horizontal lava flows with well developed columnar jointing. These flows tend to be thicker toward the top of the pile. Basaltic plugs with abundant fresh olivine, cropping out E of Cerro Divisadero and in the Río Oscuro Valley, can be considered as part of the Tertiary basaltic episode. Whole-rock K/Ar radiometric ages on Balmaceda basalt have given Eocene ages (Baker *et al.*, 1981; Butler *et al.*, 1991). Indeed, the basalt flows are overlain by Miocene to Pliocene continental sediments which contain intercalated acidic pyroclastic units (Suárez *et al.*, 1994).

Similar flat-lying basaltic mesa (about 100 m thick) also occurs 200 km N of Balmaceda, in the Winchester River and Cisnes River valleys (Demant *et al.*, 1996). Most of the flows are olivine basalt, but this mineral is frequently transformed to iddingsite. A peralkaline rhyolitic dome (El Chueco) is associated with the basaltic sequence in the Cisnes River valley. These basalts show slightly negative anomalies in Ta and Nb on a Pearce (1983) spidergram, whereas the Balmaceda basalt has regular trends and enriched REE patterns more akin to OIB. No radiometric ages are available for the basalt from Cisnes River, but overlying lower Miocene continental sediments (Marshall and Salinas, 1990) and upper Miocene to Pliocene glacial or fluvio-glacial deposits clearly post-date the volcanic activity which can be therefore considered as broadly contemporary with the Balmaceda sequence.

Late Miocene to Recent undersaturated basalt, covering Eocene Plateau basalt flows are exposed near Chile Chico, on the southern side of Lake General Carrera. These alkali basalt flows contain frequent mantle xenoliths (Niemeyer, 1978; Baker *et al.*, 1981; Flint *et al.*, 1994). Similar rocks crop out at Meseta Buenos Aires (Charrier *et al.*, 1979), where they are interbedded with fossiliferous marine sediments.

Farther to the E in Argentina, Ramos and Kay (1992) have described Ne-normative alkali basalt and hawaiite near Lake Strobel. The K/Ar ages are between 16 and 4 Ma. The geochemistry of the lava and the development of this plateau basalt between 46°30'S and 49 °S at a latitude corresponding to a gap in the volcanic arc, have been considered (Ramos and Kay, 1992; Gorring *et al.*, 1997) as an evidence of their genetic relationship with the asthenosphere window created by the subduction of the active Chile mid-oceanic ridge.

Generally speaking, a progressive change in the geochemical signatures of the plateau basalt is observed with time. Late Cretaceous basalt flows at Morro Negro have patterns, which typify the influence of a lithosphere mantle source. However, the imprint of an asthenosphere component increased progressively during the Tertiary as shown by the fading of the Ta-Nb anomaly and the enrichment in light REE emphasized by the increase of the (La/Yb) ratios. Isotopic data (Stern *et al.*, 1990) also highlighted a clear difference between eastern cratonic basalt with Sr, Nd and lead isotopic ratios similar to Oceanic Island Basalt (OIB) and western basalt occurring near the volcanic arc, which have transitional characteristics and isotopic values similar to the Andean orogenic lavas.

The Taitao Ophiolite : a Late Cenozoic Ridge Collision Related Setting

The existence of this lithological association, unique to the Andean segment considered here, was first reported by Forsythe and Nelson (1985). Further studies by Mpodozis *et al.* (1985), Forsythe *et al.* (1986), Kaeding *et al.* (1990), Allen *et al.* (1991), Nelson *et al.* (1993), Lagabrielle *et al.* (1994) and Bourgois *et al.* (1996) have clarified the nature of the ophiolite and of the time and spatially related near-trench magmatism and sedimentation.

The Taitao Suite is a complete ophiolitic sequence: ultramafic tectonites, layered and massive gabbro and peridotite, sheeted dykes, pillow lava and terrigenous clastic sedimentary rocks deposited in a shallow marine to sub-aerial environment. Geochemically the ophiolite has a mixed oceanic and arc signature. Two possible origins have been proposed for the ophiolite: obduction of a block of oceanic crust during ridge collision, or an *in situ* generation in a collision-related rift or transtensional basin within the fore arc. The age of the ophiolite is still under discussion, a 3 to 5 Ma age was proposed by Mpodozis *et al.* (1985), whereas a Miocene age has been proposed by Bourgois *et al.* (1996), *based on a K/Ar amphibole age.*

Lagabrielle *et al.* (1994) have presented evidence that the early Pliocene to early Pleistocene volcanic sequences, which have a close spatial relationship with the Taitao Ophiolite in the western part of the Taitao Peninsula (Fig. 7), are not part of the ophiolite suite, but rather the result of near trench magmatism related to the then recently subducted segment of the Chile Rise. During their transit to the surface, the magmas would have been contaminated with the metasediments of the Chonos Metamorphic Complex, a process which coupled with fractional crystallization, would have given rise to the observed geochemical variety in the volcano-sedimentary units.

Pliocene (3.0 - 5.5 Ma) silicic plutons intruded the ophiolite suite and the surrounding fore-arc rocks in a position that is roughly 200 km to the W of the present day magmatic arc. These plutons are geochemically calc-alkaline and I-type, but they appear to be a mixture of a mantle-derived basaltic parent together with a certain amount of crustal contaminant (Kaeding *et al.*, 1990) derived by heating of the base of the fore arc by the subducting ridge in a typical near-trench setting.

The active volcanic belt : response to present day subduction

The active volcanic belt in the Patagonian Andes is composed of large stratovolcanoes and small monogenetic cones. Their spatial distribution differs N and S of the Chile Triple Junction. Three stratovolcanoes per degree of latitude occur to the N, whilst only four post glacial edifices are present along the 900 km S of the triple point (Fuenzalida, 1974). The stratovolcanoes (Fig. 8) N of the Chile Triple Junction are, from N to S, Hornopiren, Huequi, Chaiten, Michimahuida, Melimoyu, Mentolat, Maca, Cay and Hudson. To the S of the Hudson volcano, encompassing the triple point, and for nearly 200 km, there is a gap in the volcanic arc and no recent stratovolcanos are known. Then, five volcanoes constitute the Austral Volcanic Zone which are from N to S, Lautaro, Aguilera, Reclus, Burney and Mount Cook. The last named volcano lies in the Fuegan Andes. The products of the stratovolcanoes are typically basaltic to andesitic in composition; only the Chaiten volcano has a significant rhyolitic component (Onuma *et al.*, 1985). They constitute a typical calc-alkaline volcanic belt. The Hudson Volcano, the only historically active volcano of the southern Andes, has had two large eruptions in 1976 and 1991 (Deruelle and Bougois, 1993; Naranjo and Stern, 1998). The eruption in 1991 produced a huge volume of tephra, which covered extensive areas of the main Andean Belt and the adjoining extra-Andean Patagonia.

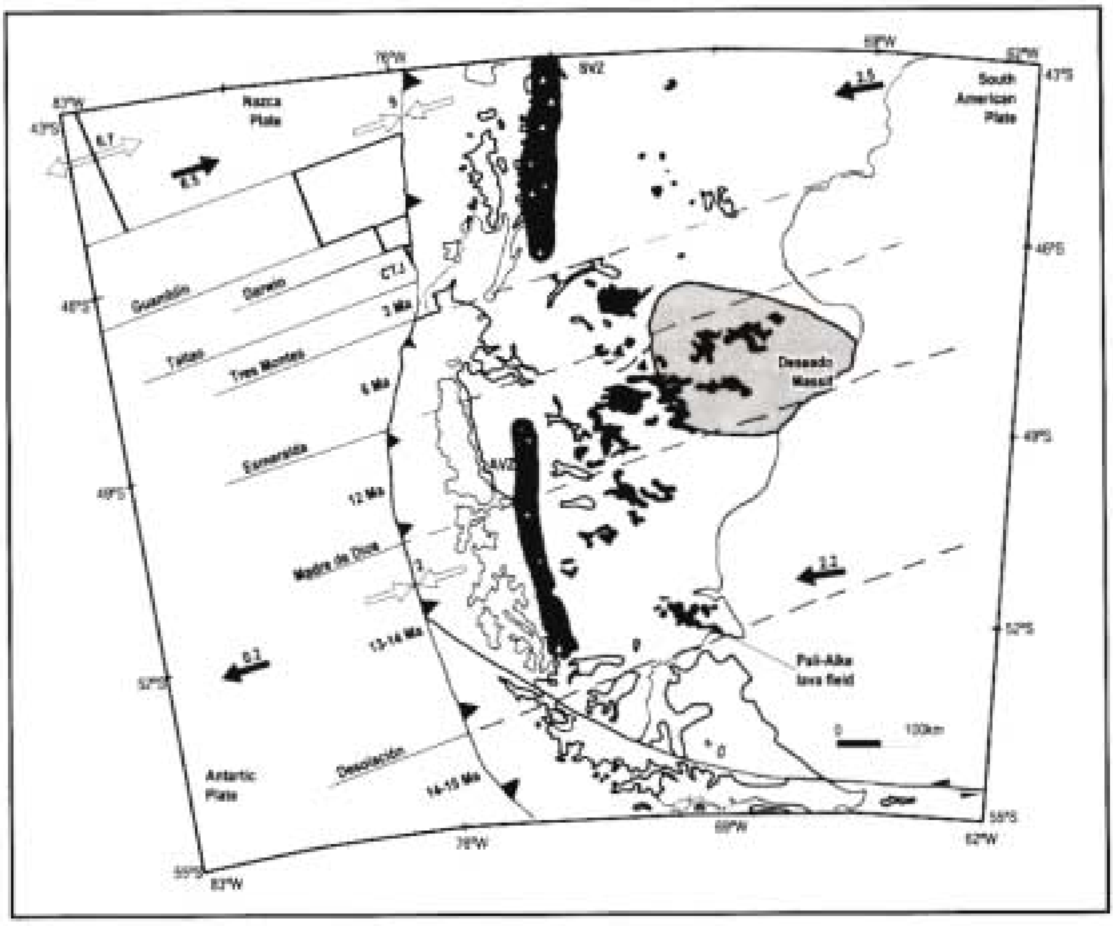
Small monogenetic cones are widely distributed in the Patagonian Andes. Many are situated on the flanks or in the neighbourhood of the stratovolcanoes, but others are also dispersed between them. They are usually situated along faults or in shear zones related to the Liquiñe-Ofqui Fault Zone(LOFZ). López-Escobar *et al.* (1995) have indicated a relationship between the position and the characteristics of the erupted magmas and the tectonic regime in the area between 40 °S and 44 °S. They point out that volcanoes generating more evolved magmas, dacite or even rhyolite, *are found only in compressional NW trending faults related* to the LOFZ, whereas those volcanoes situated along the main structure, or along NE extensional faults have erupted exclusively basaltic lava.

The area between Puyuhuapi, the Aysén Fjord and Lake General Carrera, is a good example of the distribution and characteristics of these small monogenetic centres. Demant *et al.* (1994, 1998) have shown that a complex association of calc-alkaline (Isla Colorada, Fiordo Aysén), alkaline (Puyuhuapi) and E-MORB type basalts (Murta) exists in this part of the Holocene volcanic belt.

Colorada Island, an emergent volcano occurring near the southern entrance of the Aysén Fjord, is the southernmost edifice of a Holocene volcanic field situated at the base of the Maca Stratocone. On the northern shoreline of the fjord this volcanic field also has a complex hyaloclastic vent (surtseyian cone), 1 km in diameter, and two breached scoria cones. Two other vents and associated lava flows, lying 4 km to the N and NW, are much less covered by the vegetation and therefore seem very recent. These edifices are composed of plagioclase-rich lava, mostly basalt and basaltic andesite in composition (51 - 54% silica). The lava has low-*Ti* and high alumina contents as well as strong negative anomaly in Nb, chemical characteristics typical of the calc-alkaline series.

The Puyuhuapi Volcanic Centre is composed of several cones situated at the northern end of the Ventisquero Fjord, the morphological expression of the Liquiñe-Ofqui Fault Zone in the region. Three of these volcanic vents have dammed the valley giving rise to Lake Risopatrón, and one of them has the typical morphology of a hyaloclastic cone built in a subaqueous environment. Other cones are situated on the relief, which defines the western margin of the valley. The absence of glacial fingerprints on the lava indicates that these outcrops are Holocene in age. Puyuhuapi lavas are relatively aphyric and contain only small olivine phenocrysts. They have a relatively primitive basaltic composition (MgO>7%). Puyuhuapi basalt flows have most of the characteristics of alkali basalt (Demant *et al.*, 1994; González Ferrán *et al.*, 1996) and differ from the other monogenetic centres situated to the N of Puyuhuapi along the LOFZ.

The Murta Basalt occur about 30 km SSE of the Hudson Volcano and to the E of the LOFZ. The presence of lava tubes,



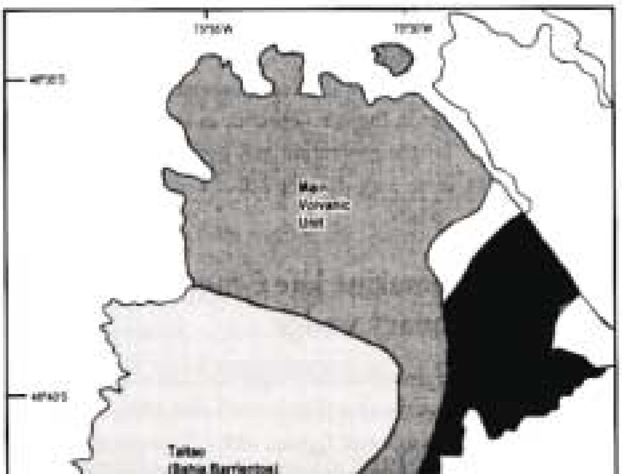


FIGURE 7 - The Taitao (Bahia Barrientos) ophiolite is one of the most singular rock bodies in the area. 1. Chonos Metamorphic Complex. 2. Main body of the ophiolite with ultramafic rocks, gabbro and a dyke complex. 3. Main Volcanic units, pillow basalt and sediments, Pliocene-Pleistocene. 4. Chile Margin unit, volcanic and sedimentary rocks. 5. Cabo Raper and Seno Hooper adakitic intrusives (Pliocene). Modified after Le Moigne et al. (1996).

FIGURE 8 - Map showing the distribution of volcanic centres in the southern portion of the South American Plate, and the main tectonic features in the Nazca and Antarctic oceanic plates. In the South American Plate, fields with triangles are the location of the main volcanic centres in the southern part of the Southern Volcanic Zone (SVZ) and in the Austral Volcanic Zone (AVZ). Note Cook Volcano near the lower end of the map. In black, the distribution of the main patagonian basalt fields. In the oceanic plates, the names of the fracture zones, (thin ENE lines) are given, as well as the age in which the intervening segment of the mid-ocean ridge was subducted beneath the leading edge of the South American Plate. The Chile Triple Junction (CTJ) is located, as are the segments of the Chile rise not yet subducted. Filled black arrows indicate the absolute motion of plates, white arrows indicate relative velocities (modified after Gorring et al., 1997).

grading laterally to pillow lava and palagonitic tuff, and irregular prismatic jointing indicates that this volcanism occurred under a thick carapace of ice. The very well preserved morphology of the lava flows, and their position near the valley floor, suggest that this volcanic event happened during the late glacial episode. These basalt flows are locally highly porphyritic and contain large plagioclase feldspar and clinopyroxene megacrysts (Demant *et al.*, 1998). The Murta Basalt has regular patterns on a MORB-normalized trace elements diagram (Pearce, 1983) characterized by slight enrichment in incompatible elements (K, Rb, Ba, Th) and slight negative anomalies in Nb and Ta. The chondrite-normalized REE patterns of the Murta Basalt is relatively flat ($La_n / Yb_n = 3.5$) resembling that of E-MORB.

The diversity of the petrographical types present in this part of the Patagonian Andes, is clear evidence of the influence of the subduction of the Chile Ridge beneath the South American Margin in the petrogenesis of these magmas. It compares with what can be observed in the Taitao Peninsula or at the Chile Ridge (Lagabrielle *et al.*, 1994; Le Moigne *et al.*, 1996). The Austral Volcanic Zone (AVZ) has been little studied due to inaccessibility of some of the volcanoes. In fact, even the precise location of many of them was only been established after 1960 (Martinic, 1988). On average, the volcanic products of the AVZ are mainly andesite and some dacite. Harambour (1988) has shown a progressive increase of SiO_2 and K_2O in the volcanic rocks from S to N. According to Futa and Stern (1988) these variations are related to a change in the stress regime of the continental crust. On passing from a compressional to a transcurrent environment towards the S, there would occur a faster transit of the magmas to the surface, thus reducing the possibility of magmatic differentiation in the magma chamber. Stern and Kilian (1996) have shown the adakitic character of the andesite and dacite of the Austral Volcanic Zone, their geochemical modelling indicated a 35 to 90% contribution of the subducted oceanic crust to the magmas. Crustal assimilation and fractional crystallization processes and the mass contribution of the continental crust increase from S to N. The adakitic nature of the magma and its variations are explained in terms of the variation in the angle of subduction of the oceanic lithosphere, from very a low angle in the volcanic arc gap zone, increasing to the S to intermediate values below Mount Burney (51°S - 52 °S), to low again at the southermost part of South America, where oblique subduction and slab fragmentation are occurring, coupled with a low subduction velocity (2 cm/yr).

Mineral deposits: a poorly known potential

The Patagonian Andes are not rich in mineral deposits compared to northern segments of the Andes. In addition, all of the known mineral deposits are small. This can be clearly observed in maps that show the geographical position of the major ore-bodies, *e.g.*, Petersen (1990), who however does not report some well known occurrences in the Patágonian Andes. This appears to be strange, as the region has been subject to long periods of subduction, with accompanying magmatic and deformational processes, which are considered a positive feature for the generation of diverse types of mineral deposits (Nelson, 1996). This author defines six metallo-tectonic terranes in the mountain belt, with geological conditions favourable to the generation of different types of mineral deposits in this supra-subduction zone setting.

Townley (1996) established that the mineral deposits in the back-arc region of Aysén (44°S-47°S) revealed two metallogenic epochs, of Late Jurassic and Early Cretaceous ages, and three metallogenic provinces or groups. The southernmost El Faldeo Group (S of 47°05'), hosted by Late Jurassic volcanic and sedimentary rocks and/or by the Eastern Andes Metamorphic Complex, is dominated by Zn-Pb skarn and polymetallic veins, locally overprinted by late Au-rich epithermal mineralization. Some features of porphyry copper type mineralization are also observed. The Fachinal Group (46°05'S - 47°05'S) is characterized by epithermal mineralization of precious metals and Zn-Pb skarns, in Lower Cretaceous acid volcanic rocks and the basement respectively. The El Toqui Group (44°30'S - 45°50'S) has Early Cretaceous mineralization of Au-rich Zn-Pb skarn type, with associated polymetallic Au-Ag rich epithermal veins. Lead isotopes indicate an orogenic source of lead, with significant addition of crustal components at El Faldeo and El Toqui groups. This study provides good evidence of the important influence of the basement rocks as a source for metals deposited by hydrothermal systems in the overlying Mesozoic sequences. The lack of such a basement in the fore-arc regions of the Patagonian Andes could be a factor in the general absence of mineral deposits in that setting. There is also some evidence that the Miocene plutons in the magmatic arc represent deeply emplaced rocks (Hervé *et al.*, 1996a), so that the overlying rock column, more favourable for the emplacement of magmatic related ore deposits, has been eroded. This leaves the back-arc setting, with an older basement and lower uplift/erosion rates in the late Cenozoic, as the most favourable prospect for the generation and preservation of mineral deposits, in keeping with the actual distribution of known mineral deposits.

The submarine late Cenozoic accretionary wedge

Marine geological and geophysical data acquired mainly in the last two decades (Cande and Leslie, 1986; Bourgois *et al.*, 1996; Bangs and Cande, 1997) have permitted the identification of the general geological characteristics of the slope and trench along the considered part of the continental margin. Besides which it has been possible to define the more important aspects of the sedimentary and tectonic processes that occur in this active environment.

In a N-S section along the central axis of the trench, the thickness of the sedimentary infill in the trench varies greatly, from 0 to 6 km, increasing to the S and to the N of the present triple junction. This may be controlled by the age vs depth relationship of the surface of the oceanic slab, which becomes younger towards the Chile Ridge. At the triple junction, the landward trench slope is narrower and steeper than along other sectors of the margin, and the trench axis has migrated shorewards. These features are interpreted as a result of subduction-driven tectonic erosion of this portion of the continental margin.

From the evidence presented, subduction-related tectonic erosion must have taken place in other sectors of the margin. The volume of the accretionary wedge (e.g., at the precollisional margin at 38 °S to 40 °S) has only accumulated in the latest Cenozoic, an indication that earlier accretionary material in this long-lasting active margin must have been tectonically eroded. However, in this case, tectonic erosion was probably triggered by low rates of sedimentation, unable to completely cover the relief of the subducting crust, which preceded the high rates of sedimentation that occurred during the late Pliocene-Pleistocene glaciation of the region. This has lead to a model in which sediment accretion and tectonic erosion are distributed along the margin under the independent controls of ridge, or fracture zone subduction, and sediment availability, giving rise to a regime of periodic accretion and erosion in any one part of the margin.

Sediment overflow to the W of the outer rise of the trench, has resulted in the deposition of sediments of continental provenance in the abyssal plains of the Antarctic Plate, thus enlarging the areal influence of orogenic and erosional processes in the continental margin.

The glaciers: remnants of the ice ages

The Patagonian Andes are host to the Northern and Southern Patagonian icefields continental glacier systems that are a remnant of the more extensive ice cover of the region in the late Cenozoic and Pleistocene (Fig. 9). The Southern Patagonia Icefield has an area of about 30 000 km², and is the third largest icefield in the world.

Continental glaciation in the Patagonian Andes has been active for the past 4.6 Ma (Rabassa and Clapperton, 1990). Between 2.1 and 1.0 Ma, there occurred numerous Patagonian glacial events, including the greatest advance of the Patagonian glaciation at 1.2 Ma (Mercer, 1976). The last glacial advance, the Llanquihue event of the past 100 000 years, retreated no more than 15 000 years ago (Heusser, 1990), when the ice front reached the western coast of the continent S of 42 °S.

According to Bangs and Cande (1997), glacial periods provide the trench with more sediment, and favour the generation of accretionary periods in the continental margin.

Cretaceous-Cenozoic Tectonic Evolution

The Patagonian Andes are the result of the tectonic interaction between the leading edge of the South America Plate and diverse oceanic plates that have been subducted from the W during geological time; at least since the Late Triassic. However, the present mountain belt, which lies parallel to the continental margin, is mainly the result of late Cenozoic tectonic processes, which allow a distinction between the northern and the southern segments, with the Nazca-Antarctica-South American Chile Triple Junction area as a dividing point. This scenario will be described first.

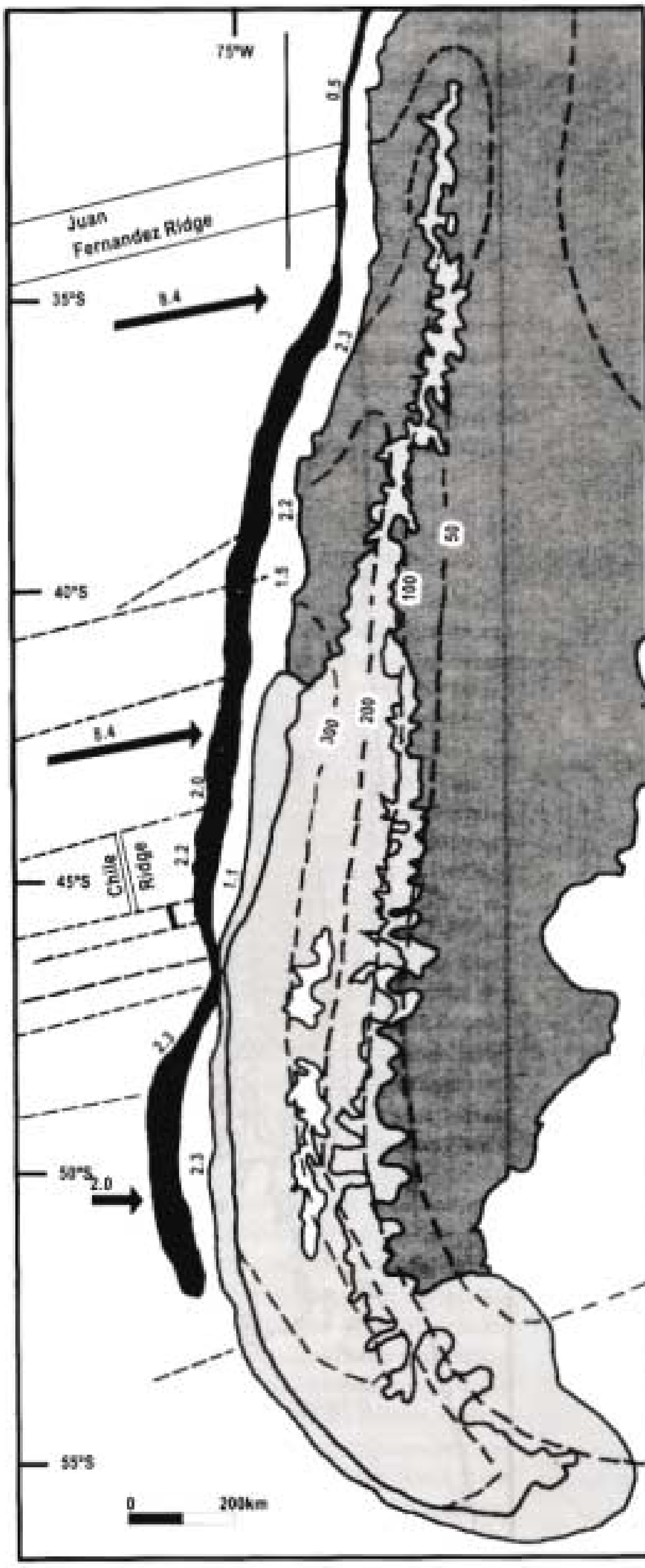


FIGURE 9 - Extent of ice coverage in southern South America at present (white), during the Llanquihue major event (20 000 years BP) and during the greatest glaciation. Width and greatest thickness of trench fill shown by darkest shading and adjacent numbers in kilometres. Arrows and adjacent numbers show relative plate directions (in km/Ma) between plates. Dashed contours indicate annual rainfall (in cm). The whole Andean Belt discussed in this paper was under the ice during the indicated ice maxima (modified after Bangs and Cande, 1997).

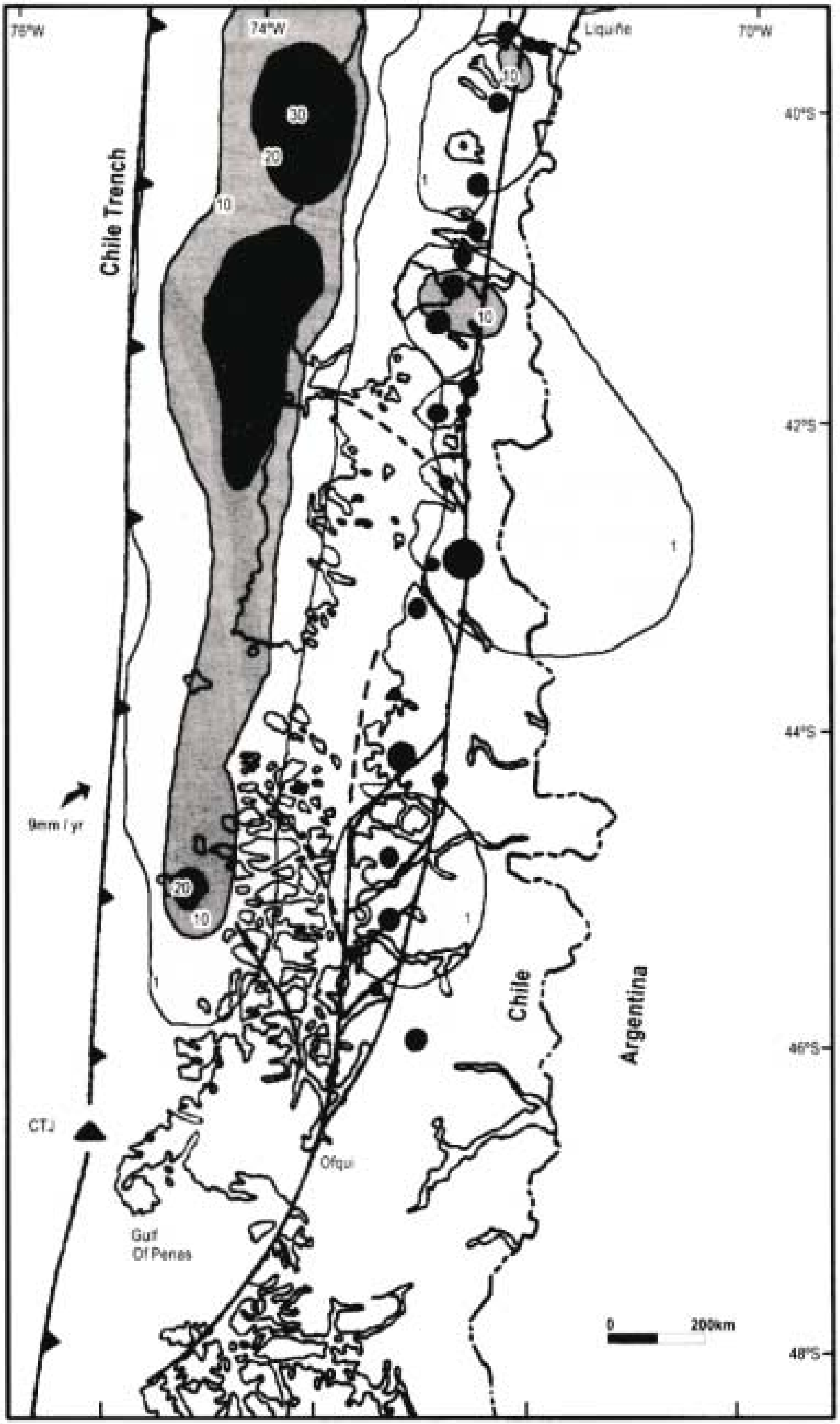


FIGURE 10 - The Liquiñe-Ofqui Fault Zone, with indication of the active volcanic centres (black dots). The size of the latter roughly proportional to the size of the volcanic edifices. Contoured areas are the amount of slip, in metres, occured in the Benioff Zone during the 1960 earthquake, after Barrientos and Ward (1990). The area without slip, coincides well with the sea-invaded central depression. Maximum slip at the western limit of the Coast Range. The eastern area with moderate slip is situated below the LOFZ and the volcanic arc (modified after Cembrano et al., 1996).

Northern Patagonian Andes: a transpressional fore arc

The N-S extent of the Northern Patagonian Andes coincides with that of the Liquiñe-Ofqui Fault Zone (LOFZ). This structure (Fig. 10) is a long intra-arc dextral strike-slip fault zone, about 1000 km long, that lies parallel the Nazca - South American plate boundary 300 km to the E, and curves towards it in its southern extremity. In its central part it has been described has having two main branches, connected by NE structures, the whole constituting an extensional strike-slip duplex. The transpressional character of the LOFZ is manifest by flower structures developed along the eastern main branch (Hervé, 1994), and by partitioning of the strike-slip component into oblique inverse faults near Puerto Cisnes (Arancibia *et al.*, 1999).

The LOFZ has been a long-lived structure, with pre-Late Cretaceous strike-slip sinistral motion in its northern part (Cembrano, 1998); mainly Miocene and Pliocene activity in its central part; and probably including Late Quaternary to Holocene activity near its southern end (Muir Wood, 1989). Seismic activity has been recorded near both ends, and indicates that the Chilean fore arc is undergoing shortening and a slight trench-parallel movement. Oblique subduction has been considered as the driving mechanism of the LOFZ activity (Hervé, 1976), but the indenter effect of ridge subduction has also been raised as a possible mechanism (Forsythe and Nelson, 1985; Nelson and Forsythe, 1989). Cembrano (1998) concluded that the influence of the indenter effect of subducting segments is more noticeable in the southern parts of the structure.

No regional thrust and fold belt has developed in the back-arc region of this part of the mountain belt (Ramos and Kay, 1992), and most of the margin-perpendicular shortening has been absorbed in the fore arc, particularly along the LOFZ (Cembrano, 1998).

The local existence of a thrust and fold belt has been pointed out by Flint *et al.* (1994) affecting late Oligocene-early Miocene marine deposits of the Cosmelli Basin at the southwestern end of Lake General Carrera. Hasegawa *et al.* (1971) mapped the thrusting of basement rocks over the Ibañez Formation (Jurassic) at the northwestern shore of the lake. This area is near the transition from the Northern to the Southern Patagonian Andes. Marshall and Salinas (1990) present stratigraphic and paleontologic evidence that the last uplift phase of the Cordillera began *c.* 18 Ma, and that the Patagonian Andes started to produce a noticeable rain shadow effect in the back arc region about 15 Ma in the middle Miocene.

Southern Patagonian Andes: development of a fold and thrust belt

In this segment of the mountain belt, a well-developed late Cenozoic thrust and fold belt is present in the back-arc region, as has been described by Ramos (1989). This Patagonian Fold-and-Thrust Belt (FTB) involves eastward thrusting of the Paleozoic mud pile over early Cenozoic deposits, and also many thrusts within the Mesozoic sequences of the Magallanes Basin. The timing of this thrusting has been related by Ramos (1989) to the subduction of several segments of the East Pacific Rise, starting at the southern tip of the continent 14 Ma ago. The fore-arc region in this segment bears no evidence of the existence of a major shear zone, similar to the LOFZ, which could have absorbed shortening perpendicular to the orogen, which here appears to have been mainly absorbed by the fold-and-thrust belt.

Paleomagnetism and block rotations

Several paleomagnetic studies, mainly in the fore-arc region but also some in the arc or back-arc regions have been carried out along the Patagonian and Fuegan Andes. Most of these studies have reported block rotation, but no meridional displacements beyond the error limits of the paleomagnetic method (5 °) have been established.

Studies in the vicinities of the Liquiñe-Ofqui Fault Zone in the northern Patagonian Andes (Cembrano *et al.*, 1992; Beck *et al.*, 1993, 1998; Rojas *et al.*, 1994) have shown a pattern of anticlockwise rotation in the fore-arc and clockwise rotations in the back-arc, with the magnitude of the rotation increasing from both sides towards the fault zone. This has been considered as a manifestation of the buttressing of the northwards translation of the coastal fore-arc block. The rocks in the fore-arc are Miocene or younger, and those in the back arc are Early Cretaceous but their magnetic signature was acquired in the Late Cretaceous, at least at Lake Verde (Beck *et al.*, 1998).

The Late Cretaceous and Cenozoic basalt in the back-arc region, away from the LOFZ shows no indication of major block rotations (Butler *et al.*, 1991) and is fixed to the stable Patagonian Platform.

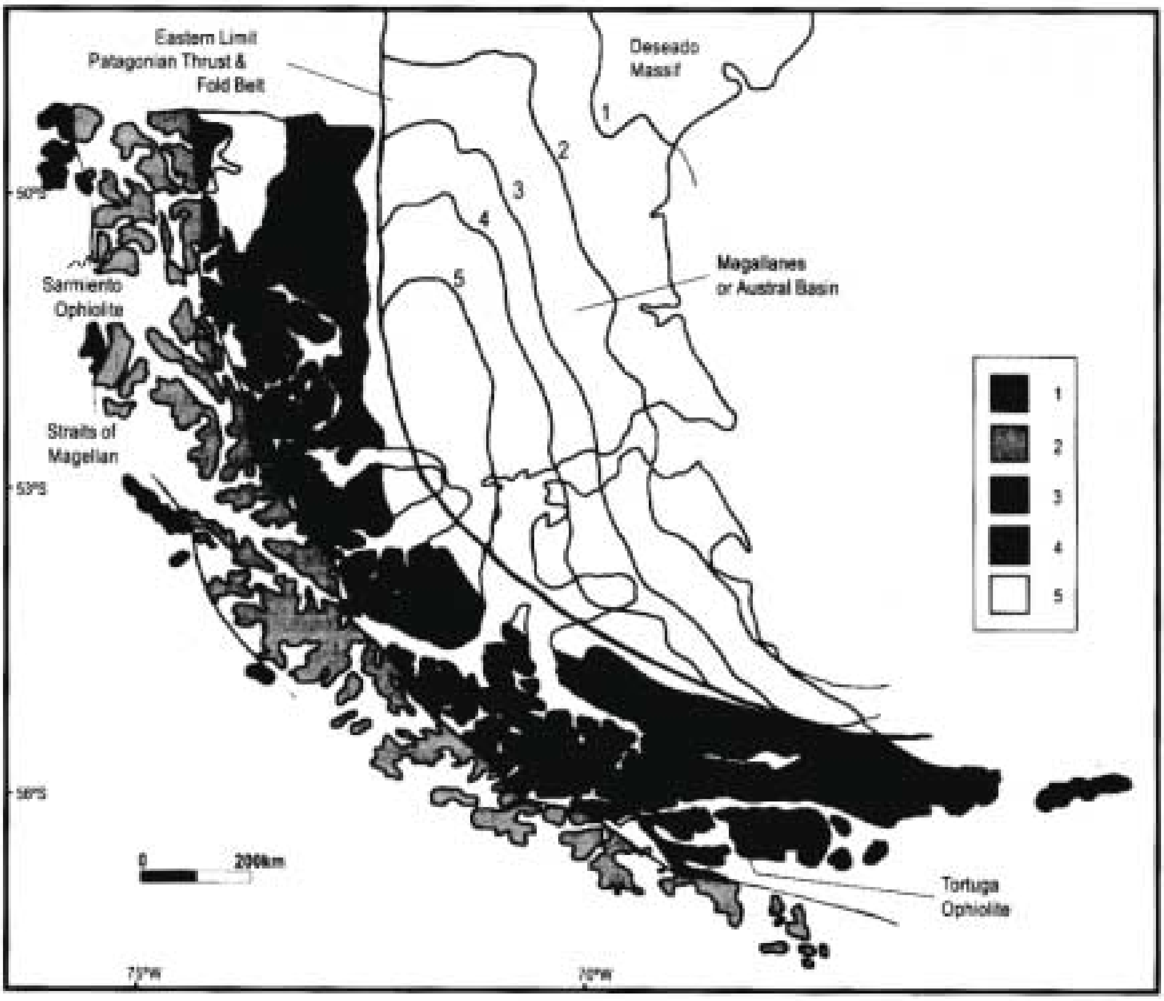
In the southern Patagonian Andes there exist evidence to support a large (*c.* 87 °) counterclockwise post Early Cretaceous rotation of the late Paleozoic Tarlton Limestone and Denaro Complex, in addition to a small northward transport (less than 5 °). However, the magnetization of the rocks under study seem to have been acquired during the Early Cretaceous, so the earlier rotation or translation of the Madre de Dios Complex has not been identified in this preliminary study.

In the Fuegan Andes, Cunningham *et al.* (1991) have shown that the Lower Cretaceous volcanic sequences in the Hardy Peninsula have been rotated counterclockwise by as much as 90°. This is in accordance with the general view that the Fuegan Andes were rotated left-laterally along the Magellan Fault Zone and associated structures during the oroclinal bending of the southernmost Andes. This resulted in the disruption of the South America-Antarctic Peninsula connections and the generation of the Drake Passage and the Scotia Arc.

These block rotations are interpreted by Beck *et al.* (1998) as related to a general state of subcrustal ductile shearing, probably caused by north-oblique subduction of the Nazca Plate. In this context, the Liquiñe-Ofqui Fault Zone could be merely another manifestation of this state of shear.

The Fuegan Andes

The southernmost stretch of the Andes, S of the Straits of Magellan, will be referred to here as the Fuegan Andes, as it is mainly developed along the western and southern margin of Tierra del Fuego. Many of the geological units here are continuous with those of the Patagonian Andes. The



topographic culmination is the ice-capped Darwin Cordillera. Forsyth (1975) suggested that this part of South America is part of the Scotia Microplate, separated from the South American Plate by the Magellan Fault Zone-Malvinas Chasm Fracture Zone along the North Scotia Ridge.

This part of the Andes is strongly curved, changing from the N-S trend of the Patagonian Andes towards the EW trend of the submarine North Scotia Ridge. It is bounded on its continental side by foothills and the Atlantic Lowlands, underlain by a thick sequence of Late Mesozoic to Cenozoic sedimentary rocks that were deposited in the Magallanes or Austral Basin.

The main geological units in the Fuegan Andes are a metamorphic basement; silicic to intermediate volcanic rocks of Middle to Late Jurassic age (Tobifera Formation); a Late Mesozoic to Early Cenozoic batholith which constitutes the present margin of the continent; a Late Jurassic to Early Cretaceous ophiolite sequence; Early Cretaceous sediments (Yaghan Formation); and minor Cenozoic to Holocene volcanic rocks (Fig. 11).

The pre-Jurassic basement: Cretaceous metamorphism

A complex of highly deformed psammo-pelitic schist and gneiss and associated metabasite and metachert underlies the Mesozoic units in the Darwin Cordillera area. They are considered to be part of the Late Paleozoic accretionary complexes that crop out along most of the coastal region of Chile S of 34 °S. However, there is ample structural and geochronological evidence to show that this complex was penetratively deformed and metamorphosed during the Middle Cretaceous tectonic events (Nelson *et al.*, 1980).

The metamorphic grade is variable, and reaches the amphibolite facies at the Darwin Cordillera, N of the Beagle Channel, where garnet, staurolite, kyanite and sillimanite were first identified by Nelson *et al.* (1977). A clockwise P-T path culminating at *c.* 8 kbar and 600 °C in the kyanite stability field was recognized, with a subsequent almost isothermal exhumation into the sillimanite field, followed by rapid cooling of the complex. It is assumed that the prograde metamorphism is related to the second deformational event, which took place in the mid-Cretaceous and is related to the closure of the Rocas Verdes Basin. To the S of the Beagle Channel, the rocks of the metamorphic complex are of chlorite and biotite grade. This large increase in metamorphic condition has been interpreted as an indication of the existence of a structural discontinuity along the Beagle Channel: an extensional detachment fault, or mainly a strike-slip structure are two possible interpretations.

The relationship of this metamorphic basement in the Andean region to the older gneissic basement, which underlies the eastern parts of the Austral Basin deposits, is not known.

FIGURE 11 - Geological sketch map of the Fuegan Andes as referred to in this paper. Main geological units: 1. Fore-arc accretionary complex of supposed late Paleozoic age and Darwin Cordillera metamorphosed in the Cretaceous. 2. Mesozoic to Cenozoic Fuegan Batholith. 3. Mesozoic volcanic and sedimentary units. 4. Late Jurassic-Early Cretaceous ophiolites of the Rocas Verdes Basin. 5. Cenozoic rocks with contoured thickness (in km) of the sediments of the Magallanes or Austral Basin (modified after Wilson, 1991).

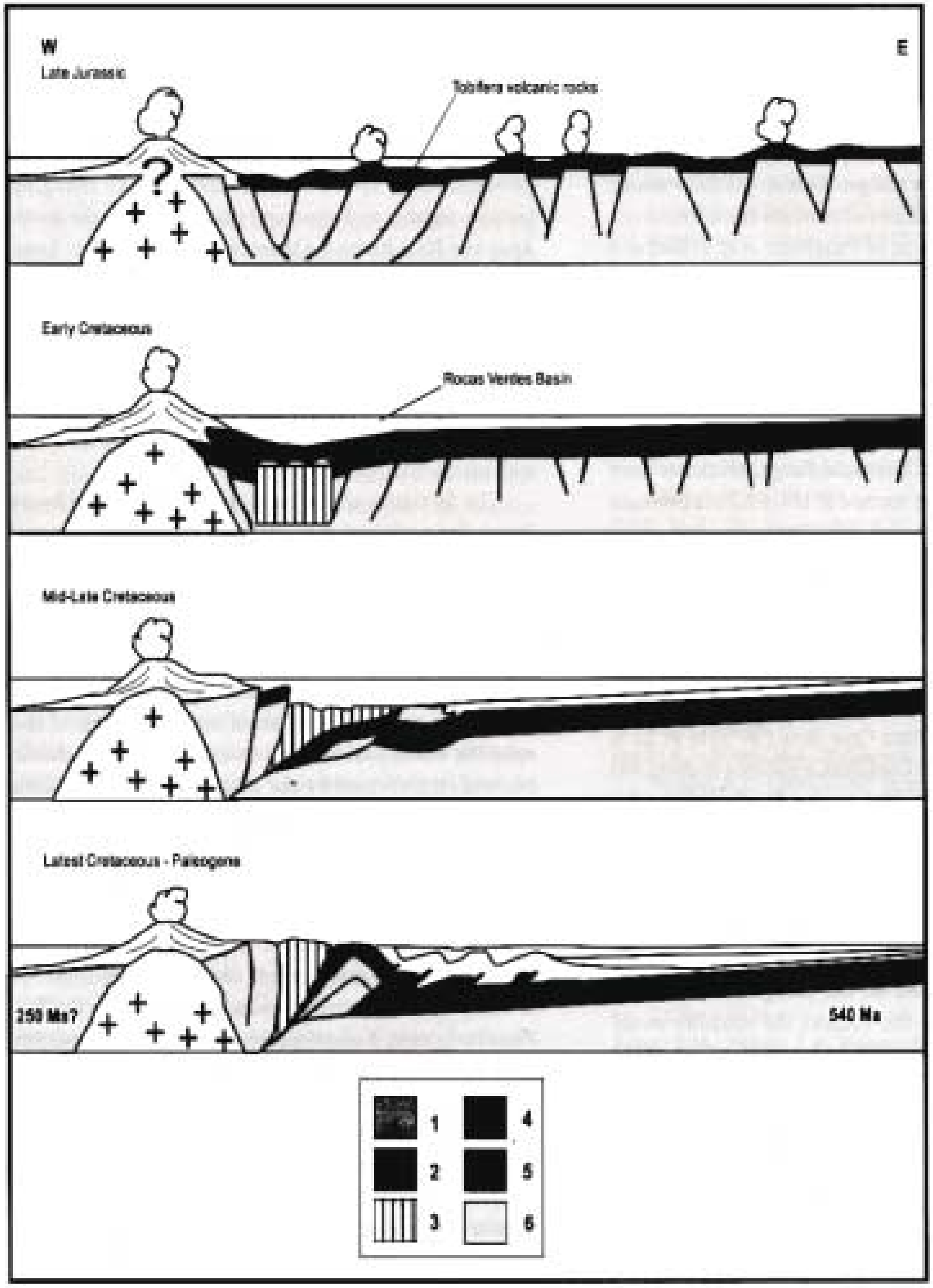


FIGURE 12. - Diagrammatic cross sections depicting the geological evolution of the Fuegan Andes from the Late Jurassic to the Paleogene. 1. Syn-orogenic sediments in the foreland basin. 2. Volcaniclastic sedimentary infill in the Rocas Verdes Basin. 3. The ophiolite floor of the back-arc Rocas Verdes Basin. 4. Early Cretaceous sediments on the west-facing platform of the Magallanes or Austral Basin. 5. Tobifera volcanic rocks. 6. Metamorphic basement of late Paleozoic (250 Ma?) age near the arc, and early Phanerozoic (540 Ma) to the E (modified after Wilson, 1991).

The Tobifera Formation: the product of an extensional volcanotectonic rift zone

A widespread blanket of silicic volcanic and volcaniclastic rocks, was generated over most of Patagonia, including the Fuegan Andes, during the Middle to Late Jurassic, in response to the occurrence of a regional extensional event associated with the break-up of Gondwana (Bruhn *et al.*, 1978). These formations are most clearly seen in extra-Andean Patagonia (*e.g.*, Chon-Aike Formation) and in the Patagonian Andes (Ibañez Formation, referred to above). The predominant rock types are rhyolitic ignimbrite, and rhyolitic or andesitic tuff and lava. In the extreme S, such rocks are assigned to the Tobífera Formation. This formation is exposed within the Andean Cordillera, where it is deformed and often altered. However, it also occurs at depth throughout the Magallanes Basin, where its occurrence has been proved by drilling. Recent accounts of the petrology and geochemistry of this volcanic province, which also continues southwards into the Antarctic Peninsula, have been given by Pankhurst *et al.* (1998) and Riley and Leat (1999). SHRIMP U/Pb zircon dating shows that the locus of eruption migrated westwards from the Atlantic Coast at *c.* 190 - 180 Ma to reach the Andes by about 160 - 150 Ma (Pankhurst *et al.*, in press). S-type and subvolcanic granites associated with the Jurassic volcanic rocks are often assumed to be cogenetic with them, *e.g.*, the Darwin Granite Suite of Tierra del Fuego, which has been dated by the U/Pb zircon method at 164 ± 1.7 Ma (Mukasa and Dalziel, 1996).

The ophiolite complex: the floor of an Early Cretaceous back arc basin

The ophiolite suites in the Fuegan Andes (Dalziel *et al*, 1974) crop out in a series of narrow, discontinuous lenses of mafic igneous rocks from Cape Horn (56 °S) to as far N as 50 °S in the Sarmiento Cordillera, a distance of about 800 km. These rocks are flanked on the Pacific side by the Patagonian Batholith, a calc-alkaline body averaging 60 km wide. The batholith is intruded into the Paleozoic metamorphic rocks and its unconformable cover of Jurassic silicic volcanic rocks of the Tobífera Formation. The batholith is interpreted as the root of a subduction-related magmatic arc developed on the margin of the South American continent. In this context, the ophiolite would represent the mafic floor of a back-arc marginal basin, separating the magmatic arc from stable continental South America. It has usually been thought that this basin formed in latest Jurassic time, but U/Pb dating by Stern *et al.* (1992) suggests an age of 141 - 137 Ma (earliest Cretaceous) for the oceanic floor of the basin.

The pseudostratigraphy of the ophiolite suite consists of gabbro, over a kilometre thick, a sheeted dyke complex that is overlain by an extrusive unit of considerable thickness consisting of pillow lava, pillow breccia and tuff (Saunders *et al.*, 1979). Leucocratic plutonic rocks are also present, some apparently representing fused stopped blocks of sialic crust, but others corresponding to plagiogranites, in irregular or sheet-like bodies tens to hundreds metres across. Only minor ultramafic rocks have been found associated to the ophiolite.

The extrusive components of the ophiolite are overlain by and intercalated with Lower Cretaceous (Aptian) shale in the northern outcrops, and by thick, up to 2 km, volcaniclastic greywacke, including some andesite flows, of the Yaghan Formation in the Beagle Channel area. The latter are thought to represent infill into the ophiolite basin from the western southern active volcanic arc.

The Magallanes or Austral Basin: a long lived foreland basin

A thick wedge of sedimentary rocks, thinning towards the E, was deposited over the South America continental crust in the foreland of the Andes during the Mesozoic and the Cenozoic. The western regions are involved in the fold-and-thrust belt, which constitutes the eastern margin of the Fuegan and southern Patagonian Andes.

Wilson (1991) has distinguished four stages in the development of the basin, and links them to changing tectonic regimes and structural patterns within the South American Plate during the Mesozoic and the Cenozoic. Basin formation in the Jurassic and Early Cretaceous occurred during regional crustal extension. The initial development of a deep marine through coincided with the start of the silicic Tobífera volcanism along a 1000 km long sector of the Pacific Margin. In the Ultima Esperanza area, the upper volcanic levels of the Tobífera Formation are interbedded with marine mudstones.

The formation of the back-arc ophiolite in the Rocas Verdes Basin (Dalziel *et al.*, 1974) in the Early Cretaceous, where coarse grained sedimentation took place, was accompanied by the deposition of several hundred metres of dark mudstone rhythmically interbedded with thin sandstone beds of the Zapata Formation in the west-facing continental slope to the E. In the Late Albian to Cenomanian, the compressional deformation and obduction of the ophiolite basin caused the deposition of coarse clastic material, represented by the sand rich turbiditic Punta Barrosa Formation, to shift eastwards into a flexural foreland basin. The sediment supply came now mainly from the western magmatic arc and the tectonically rising Main Cordillera of the Fuegan Andes. From the Cenomanian to the Maastrichtian, a fan-like depositional system in a tectonically stable environment allowed the deposition of at least 3 km of sedimentary rocks, represented by the Tres Pasos Formation. A major erosional unconformity separates these units from mid-Eocene shallow marine, deltaic and fluvial deposits environment, which persisted until the Neogene in the Magallanes Basin.

The Fuegan Andes underwent an important orthogonal compression during late Eocene times, indicated by a strong angular unconformity in the sedimentary sequence (Olivero and Malumián, 1999), and by the inception of the foreland basin stage in the Malvinas Basin (Galleazzi, 1996).

The Fuegan Batholith: the roots of a magmatic arc

The plutonic rocks of this part of the batholith were emplaced into Late Jurassic-Early Cretaceous sedimentary and volcanic rocks formed in an island arc-marginal basin system (Hervé *et al.*, 1984). Three main plutonic groups have been recognized with crystallization ages ranging from 141 to 34 Ma. The earliest gabbroic rocks (141 - 103 Ma) are exposed as relicts within younger granitoid plutons. The Beagle Canal Plutonic Group consists of granitoid, which show a penetrative syn-magmatic foliation, and is dated at 113 to 81 Ma. The southern Seno Año Nuevo Plutonic Group is dated as Paleogene (60 - 34 Ma), and it is composed mainly of tonalite and quartz monzodiorite having an isotropic texture. They contain inclusions of the other two plutonic groups, and are considered to be post-tectonic relative to the deformation and closure of the Rocas Verdes Basin, which took place in the Mid-Cretaceous. This spatial age-zoning of the plutonic rocks of the Fuegan Batholith, which become younger oceanwards, has no counterpart in the Patagonian Batholith, but is similar to the trend seen in the Antarctic Peninsula area.

Miocene and Recent volcanic rocks

In the southernmost Fuegan Andes, S of 54°45'S, Puig *et al.* (1984) described the presence of Miocene (21 to 18 Ma) dacite, andesite and alkali basalt. The postglacial low-K calc-alkaline andesite of Cook Island about 400 km S of Mount Burney represents the southernmost recent volcanism recorded in South America. The presence of these volcanic rocks indicates that subduction of the oceanic litosphere beneath South America was active in the Miocene and also at present, although the subducted plate changed from the Kula to the Antarctic Plate 14 Ma ago. The presence of alkali basalt in this fore arc position is not clearly understood.

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REFERENCES

Aguirre, L., Cortés, J., Morata, D. and Hervé, F. (1999). Low-grade metamorphism of Mesozoic and Cenozoic volcanic sequences in Patagonia, Chile (43°-46°S). *Revista Geológica de Chile*, **24**(2), 187-201.

Allen, R.B. (1982). Geología de la Cordillera Sarmiento, Andes Patagónicos entre los 51° y 52°15' Lat. S, Magallanes, Chile. *Servicio Nacional de Geologia y Minería, Boletín*, **38**, 46 p.

Allen, M., Nelson, E., Elthon, D. and Forsythe, R. (1991). Geochemical indicators of depleted and enriched sources for sheeted dikes of the Taitao ophiolite, southern Chile. *Actas, VI Congreso Geológico Chileno, Viña del Mar*, 1, 615 p.

Arancibia, G., Cembrano, J. and Lavenu, A. (1999). Transpresión dextral y partición de la deformación en la Zona de Falla de Liquiñe-Ofqui, Aysén, Chile (44°-45°S). *Revista Geológica de Chile*, **26**(1), 3-22.

Baker, P.E., Rea, W.J., Skarmeta, J., Caminos, R. and Rex, D.C. (1981). Igneous history of the Andean cordillera and Patagonian plateau around latitude 46°S. *Philosophical Transactions of the Royal Society, London*, **A303**, 105-149.

Bangs, N.L. and Cande, S.C. (1997). Episodic development of a convergent margin inferred from structures and processes along the southern Chile margin. *Tectonics*, **16**(3), 489-503.

Barrientos, S.E. and Ward, S.N. (1990). The 1960 Chile earthquake; inversion for slip distribution from surface deformation. *Geophysical Journal of the Interiors*, **103**, 589-598.

Beck, M.E., Jr., Rojas, C. and Cembrano, J. (1993). On the nature of buttressing in margin-parallel strike-slip fault systems. *Geology*, **21**(8), 755-758.

Beck, M.E., Jr., Burmester, R.F. and Steele, B.C. (1998) Paleomagnetism of probably remagnetized late Mesozoic volcanic rocks near Lake Verde, Aisén, Southern Chile. *Revista Geológica de Chile*, **25**(2), 153-163.

Bell, C.M. and Suárez, M. (1997). The Lower Cretaceous Apeleg Formation of the Aisén basin, Southern Chile. Tidal sandbar deposits of an epicontinental sea. *Revista Geológica de Chile*, **24**(2), 203-226.

Biddle, K.T., Uliana, M.A., Mitchum, R.M., Jr., Fitzgerald, M.G. and Wright, R.C. (1986). The stratigraphic and structural evolution of the central and eastern Magallanes Basin, southern South America. In: Foreland basins, eds. Allen, P.A. and Homewood, P. *International Association of Sedimentologists Special Publication*, **8**, 41-61.

Bourgois, J., Martin, H., Le Moigne, J. and Frutos, J. (1996). Subduction erosion related to spreading-ridge subduction: Taitao peninsula (Chile margin triple junction area). *Geology*, **24**(8), 723-726.

Bruce, R.M., Nelson, E.P., Weaver, S.G. and Lux, D.R. (1991). Temporal and spatial variations in the southern Patagonian batholith; Constraints on magmatic arc development. In: Andean Magmatism and its tectonic setting: Boulder, Colorado, eds. Harmon, R.S. and Rapela, C.W. *Geological Society of America, Special Paper*, **265**, 1-12.

Bruhn, R.L., Stern, C.R. and de Wit, M.J. (1978). Field and geochemical data bearing on the development of a mesozoic volcano-tectonic rift zone and back-arc basin in southernmost South America. *Earth and Planetary Science Letters*, **41**, 32-46.

Butler, R.F., Hervé, F., Munizaga, F., Beck, M.E. Jr., Burmester, R.F. and Oviedo, E. (1991). Paleomagnetism of the Patagonian plateau basalts, southern Chile and Argentine. *Journal of Geophysical Research*, **96**, 6023-6034.

Cande, S.C. and Leslie, R.B. (1986). Late Cenozoic tectonics of the Southern Chile Trench. *Journal of Geophysical Research*, **91**, 471-496.

Cecioni, G. (1956). Primeras noticias sobre la existencia del

Paleozoico Superior en el Archipiélago Patagónico entre los paralelos 50° y 52°S. Universidad de Chile, *Fac. Cienc. Fís. y Mat., An.*, **13**, 183-202.

Cembrano, J., Beck, M.E., Jr., Burmester, R.F., Rojas, C., García, A., Hervé, F. (1992). Paleomagnetism of Lower Cretaceous rocks from east of the Liquiñe-Ofqui fault zone, southern Chile: evidence of small in-situ clockwise rotations. *Earth and Planetary Sciences Letters*, **113**(4), 539-551.

Cembrano, J., Hervé, F. and Lavenu, A. (1996). The Liquiñe Ofqui fault zone: a long-lived intra-arc fault system in southern Chile. *Tectonophysics*, **259**, 55-66.

Cembrano, J. (1998). *Kinematics and timing of intra-arc deformation at the Southern Andes plate boundary zone.* Ph.D. Thesis, Dalhousie University, Halifax, Nova Scotia, Canada, 231 p.

Charrier, R., Linares, E., Niemeyer, H. and Skarmeta, J. (1979). K-Ar ages of basalt flows of the Meseta Buenos Aires in southern Chile and their relation to the southeast Pacific triple junction. *Geology*, **7**, 436-439.

Chinn, D.S. and Isacks, B.L. (1983). Accurate source depths and focal mechanisms of shallow earthquakes in western South America and in the New Hebrides island arc. *Tectonics*, **2**, 529-563.

Cifuentes, I. (1989). The 1960 Chilean earthquakes. *Journal of Geophysical Research*, **94**, 665-680.

Corvalán, J., (1989). Geologic-Tectonic Framework of the Andean Region. In: Geology of the Andes and its relation to hydrocarbon and mineral resources, eds. Ericksen, G.E., Cañas Pinochet, M.T. and Reinemund, J.A. *Circum-Pacific Council for Energy and Mineral Resources Earth Science Series*, Houston, Texas, **11**.

Covacevich, V., De La Cruz, R., Suárez, M. (1994). Primer hallazgo de fauna del Berriasiano Inferior (Neocomiano) en la Formación Ibáñez, Región XI, Aisén. *Actas VII Congreso Geológico Chileno, Concepción*, **1**, 425-429.

Cunningham, W.D., Klepeis, K.A., Gose, W.A. and Dalziel, I.W.D. (1991). The Patagonian Orocline: New Paleomagnetic Data from the Andean Magmatic Arc in Tierra del Fuego, Chile. *Journal of Geophysical Research*, **96**(B10), 16,061-16,067.

Dal Molin, C.N. and Franchi, M. (1996). Reinterpretación estratigráfica de las sedimentitas terciarias del suroeste del Chubut. *Actas XIII Congreso Geológico Argentino y III Congreso de Exploración de Hidrocarburos*, **I**, 473-478.

Dalziel, I.W.D. and Elliot, D.H. (1973). The Scotia Arc and Antarctic margin. In: *The Ocean Basins and Margins*, eds. Nairn, A.E.M. and Stehli, F.G., pp. 171-246. Plenum Pub. Corp., New York.

Dalziel, I.W.D., de Wit, M.J. and Palmer, K.F. (1974). Fossil marginal basin in the southern Andes. *Nature*, **250**, 291-294.

Dalziel, I.W.D. and Palmer, K.F. (1979). Progressive deformation and orogenic uplift at the southern extremity of the Andes. *Geological Society of America Bulletin*, Part I, **90**, 259-280.

Daniel, A.J., N.J. Kusznir and Styles, P. (1996). A temperature and gravity model of spreading Centre subduction, with application to southern Chile. *30th International Geological Congress, Beijing, Abstracts*, **1**, 128.

Davidson, J., Mpodozis, C., Godoy, E., Hervé, F., Pankhurst, R.J. and Brook, M. (1987). Late Paleozoic accretionary complexes on the Gondwana margin of southern Chile: Evidence from the Chonos Archipelago. In: Gondwana Six, Structure, Tectonics and Geophysics American Geophysical Union, ed. McKenzie, G.D. *Geophysical Monograph*, **40**, 221-227.

De La Cruz, R., Suárez, M., Covacevich, V. and Quiroz, D. (1996). Estratigrafía de la zona de Palena y Futaleufú (43°15'-43°45' Latitude S), X Región, Chile. *XIII Congreso Geológico Argentino y III Congreso de Exploración de Hidrocarburos, Actas*, **1**, 417-424.

Demant, A., Hervé, F., Pankhurst, R.J. and Magnette, B. (1994). Alkaline and Calc-Alkaline Holocene Basalts from Minor Volcanic Center in the Andes of Aysén, Southern. Chile. *VII Congreso Geológico Chileno, Concepción, Actas*, **II**, 1326-1330.

Demant, A., Belmar, M., Hervé, F., Pankhurst, R.J., Suárez, M. (1996). Pétrologie et géochimie des basaltes de Murta: une éruption sous-glaciaire dans les Andes patagoniennes (46° lat. S), Chili. Relation avec la subduction de la ride du Chili. *Comptes Rendus de l'Academie des Sciences Paris*, **327**, 795-801.

Deruelle, B. and Bourgois, J. (1993). Sur la derniere eruption du volcan Hudson (sud Chili, aout 1991). *Comptes Rendus de l'Academie des Sciences Paris*, **316**, 1399-1405.

de Wit, M.J. and Stern, C. (1978). Pillow talk. *Journal of Volcanology and Geothermal Research*, **4**, 55-80.

de Wit, M.J. and Stern, C. (1981). Variations in the degree of crustal extension during formation of a back-arc basin. *Tectonophysics*, **72**, 229-260.

Diemer, J.A., Forsythe, R.D., Englehard, T.D. and Porter, C. (1997). An Early Cretaceous forearc basin the Golfo de Penas region, southern Chile. *Journal of the Geological Society, London*, **153**.

Douglass, R.C. and Nestell, M.K. (1976). Late Paleozoic foraminifera from southern Chile. *U.S. Geological Survey Special Paper*, **858**, 49 p.

Escobar, F. et al. (1980). *Mapa Geológico de Chile, escala 1:1,000,000.* Instituto de Investigaciones Geológicas, Santiago.

Fang, Z-j., Boucot, A., Covacevich, V. and Hervé, F. (1998). Discovery of Late Triassic fossils in the Chonos Metamorphic Complex, Southern Chile. *Revista Geológica de Chile*, **25**(2), 165-173.

Flint, S.S., Prior, D.J., Agar, S.M., Turner, P. (1994). Stratigraphic and structural evolution of the Tertiary Cosmelli Basin and its relationship to the Chile triple junction. *Journal of the Geological Society, London*, **151**, 251-268.

Forsyth, D.W. (1975). Fault plane solutions and tectonics of the South Atlantic and Scotia Sea. *Journal of Geological Research*, **80**, 1429-1443.

Forsythe, R. and Mpodozis, C. (1979). El archipiélago Madre de Dios, Patagonia Occidental, Magallanes: rasgos generales de la estratigrafía y estructura del basamento pre-Jurásico Superior. *Revista Geológica de Chile*, **7**, 13-29.

Forsythe, R.D., Dalziel, I.W.D., Mpodozis, C., Hall, B. and Barrientos, S. (1981). Geologic studies in the outer Chilean fiords, R/V Hero cruise 79-5. *Antarctic J. of the United States*, **15**(5), 109-111.

Forsythe, R.D. (1982). The Late Paleozoic to Early Mesozoic evolution of southern South America: A plate tectonic interpretation. *Journal of the Geological Society of London*, **139**, 671-682.

Forsythe, R.D. and Mpodozis, C. (1983). Geología del Basamento pre-Jurásico Superior en el Archipiélago Madre de Dios, Magallanes, Chile. *Servicio Nacional de Geología y Minería, Boletín*, **39**, 63 p.

Forsythe, R.D. and Nelson, E. (1985). Geological manifestation of ridge collision: Evidence from the Golfo de Penas, Taitao basin, southern Chile. *Tectonics*, **4**, 477-495.

Forsythe, R.D., Olsson, R.K., Hohnson, C. and Nelson, E. (1985). Stratigraphic and micropaleontologic observations from the Golfo de Penas-Taitao basin, Southern Chile. *Revista Geológica de Chile*, **25-26**, 3-12.

Forsythe, R.D., Nelson, E.P., Carr, M.J., Kaeding, M.E., Hervé, M., Mpodozis, C., Soffia, J.M. and Harambour, S. (1986). Pliocene near trench magmatism in southern Chile: A possible manifestation of ridge collision. *Geology*, **14**, 23-27.

Forsythe, R.D. and Allen, R.B. (1980). The basement rocks of Peninsula Staines, Region XII, Province of Ultima Esperanza, Chile. *Revista Geológica de Chile*, **10**, 3-15.

Frutos, J., Oyarzún, J. and Pincheira, M. (1986). *Geología y Recursos Minerales de Chile.* Editorial Universidad de Concepción, Chile.

Fuenzalida, R. (1974). The Magellan Fault Zone. In: *Proc. Symp. on "Andean Antarctic Problems"*, International Association of

Volcanology and Geochemistry of the Earth Interior, Napoli, p. 373-391.

Fuenzalida, R. and Etchart, H. (1975). *Geología del territorio de Aysén comprendido entre los 43°45' y los 45° latitud Sur.* Instituto de Investigaciones Geológicas, Santiago, Chile, 99 p.

Futa, K. and Stern, C.R. (1988). Sr and Nd isotopic and trace element compositions of quaternary volcanic centers of the Southern Andes. *Earth and Planetary Sciences Letters*, **88**, 253-262.

Galeazzi, J.S. (1996). Cuenca de Malvinas. In: Geología y Recursos Naturales de la Plataforma Continental Argentina, eds. Ramos, V.A. and Turic, M.A. *XIII Congreso Geológico Argentino y III Congreso de Exploración de Hidrocarburos, Relatorio*, Buenos Aires, **15**, 273-309.

García, A., Beck, M.E., Jr., Burmester, R.F., Munizaga, F., Hervé, F. (1988). Paleomagnetic reconnaissance of the Región de Los Lagos, southern Chile, and its tectonic implications. *Revista Geológica de Chile*, 15(1), 13-20.

Garrido, I. (1987). *Geología y petrología de Isla Teresa Archipiélago de los Chonos, Aysén. La cristalinidad de la illita como indicador de grado metamórfico.* Memoria para optar al título de Geólogo. Depto. de Geología, Universidad de Chile, Santiago (unpublished).

González Ferrán, O., Innocenti, F., Lahsen, A., Manetti, P., Mazzuoli, R., Omarini, R. and Tamponi, M. (1996). Alkali basalt volcanism along a subduction related magmatic arc: the case of Puyuhuapi Quaternary volcanic line, southern Andes (44°20'S). *XIII Congreso Geológico Argentino*, 549-565.

Gorring, M.L., Kay, S.M., Zeitler, P.K., Ramos, V.A., Rubiolo, D., Fernández, M.I. and Panza, J.L. (1997). Neogene Patagonian plateau lavas: continental magmas associated with ridge collision at the Chile triple junction. *Tectonics*, **16**, 1-17.

Harambour, S.M. (1988). Sobre el hallazgo del mítico volcán Reclus, ex Mano del Diablo, Hielo Patagónico Sur, Magallanes, Chile. *Revista Geológica de Chile*, **15**(2), 173-180.

Hasegawa, K., Matsushita, K., Nishimura, T., Hervé, F. and Akagi, Y. (1971). Geological note at Las Chivas District, Aysén Province, Chile. *Bulletin of the Geological Survey of Hokkaido*, Sapporo, **44**(3), 25-29.

Heim, A. (1940). Geological observations in the Patagonian Cordillera. *Eclogae Geologicae Helveticae, Zürich*, **33**(1), 25-51.

Hervé, F., Davidson, J., Godoy, E., Mpodozis, C. and Covacevich, V. (1981). The Late Paleozoic in Chile stratigraphy, structure and possible tectonic framework. *Annais da Academia Brasileira de Ciências*, **53**(2), 361-373.

Hervé, F., Godoy, E., Garrido, I., Hormazábal, L., Brook, M., Pankhurst, R.J. and Vogel, S. (1988). Geocronología y condiciones de metamorfismo del complejo de subducción del archipiélago de Chonos. *V Congreso Geológico Chileno, Tomo II, E157-E173 (Extended Abstracts).* IGCP Project 211: Late Paleozoic of South America.

Hervé, F. (1993). Paleozoic metamorphic complexes in the Andes of Aysén, Southern Chile (West of Occidentalia). *Proceedings of the First Circum-Pacific and Circum-Atlantic Terrane Conference*, Guanajuato, Mexico, 64-65.

Hervé, F., Pankhurst, R.J., Suárez, M. and De La Cruz, R. (1993). Basic magmatism in a mid-Tertiary transtensional basin, Isla Magdalena, Aysén, Chile. *Second International Symposium on Andean Geodynamics (ISAG)*, Oxford, ORSTOM editions, 367-369.

Hervé, F. (1994). The southern Andes between 39° and 44°S latitude. The geological signature of a transpressive tectonic regime related to a magmatic arc. In: *Tectonics of the Southern Central Andes*, eds. Reuter, K.J., Scheuber, E., and Wigger, P.J., pp. 243-248. Springer-Verlag.

Hervé, F., Greene, F. and Pankhurst, R.J. (1994). Metamorphosed fragments of oceanic crust in the upper Paleozoic Chonos accretionary complex, southern Chile. *Journal of South American Earth Sciences*, 7(3/4), 263-270.

Hervé, F., Pankhurst, R.J., Drake, R. and Beck, M. (1995). Pillow metabasalts in a mid-Tertiary extensional basin adjacent to the Liquiñe-Ofqui fault zone: The Isla Magdalena area, Aysén, Chile. *Journal of South American Earth Sciences*, **8**(1), 33-46.

Hervé, F., Pankhurst, R.J., Demant, A. and Ramírez, E. (1996). Age and Al-in hornblende geobarometry in the North Patagonian Batholith, Aysén, Chile. *Third International Symposium on Andean Geodynamics (ISAG)*, Saint Malo, France, 579-582.

Hervé, F., Munizaga, F., Cembrano, J. (1996). El enjambre de diques máficos Cenozoico del margen continental de Aysén, sur de Chile. *Actas XIII Congreso Geológico Argentino*, 413-420.

Hervé, F., Aguirre, L., Godoy, E., Massone, H., Morata, D., Pankhurst, R., Ramírez, E., Sepúlveda, V. and Willner, A. (1998). Nuevos antecedentes acerca de la edad y las condiciones P-T de los complejos metamórficos en Aysén. *X Congreso Latinoamericano de Geología y VI Congreso Nacional de Geología Económica, Actas*, **II**, 134-137.

Hervé, F., Fanning, M., Bradshaw, J., Bradshaw, M. and Lacassie, J.P. (1999a). Late Permian SHRIMP U-Pb detrital zircon ages constrain the age of accretion of oceanic basalts to the Gondwana margin at the Madre de Dios archipelago, southern Chile. *4th International Symposium on Andean Geodynamics, Abstracts*, Göttingen.

Hervé, F., Prior, D., López, G., Ramos, V.A., Rapalini, A., Thomson, S., Lacassie, J.P. and Fanning, M. (1999b). Mesozoic blueschists from Diego de Almagro, Southern Chile. *2nd South American Symposium on Isotope Geology, Actas*, Córdoba.

Hervé, F., Aguirre, L., Sepúlveda, V. and Morata, D., Contrasting geochemistry and metamorphism of pillow basalts in metamorphic complexes from Aysén, S. Chile. *Journal of South American Earth Sciences* (in press).

Hervé, M. (1976). Estudio geológico de la falla Liquiñe-Reloncaví en el área de Liquiñe; antecedentes de un movimiento transcurrente (Provincia de Valdivia). *Actas I Congreso Geológico Chileno*, B39-B56.

Hervé, M., Suárez, M. and Puig, A. (1984). The Patagonian Batholith S of Tierra del Fuego, Chile: timing and tectonic implications. *Journal of the Geological Society of London*, **141**, 909-917.

Heusser, C.J. (1990). Chilotan piedmont glacier in the Southern Andes during the last glacial maximum. *Revista Geológica de Chile*, 17(1), 3-18.

Hormazábal, L. (1991). *Geología del Basamento Metamórfico del Grupo Islote Smith e Islas aledañas, entre los 45°45°15'Latitud Sur, Aysén, Chile.* Memoria para optar al título de Geólogo, Depto. de Geología, Universidad de Chile, Santiago (unpublished)

Jarrard, R.D. (1986). Relations among subduction parameters. *Reviews of Geophysics*, **24**(2), 217-284.

Kaeding, M., Forsythe, R. and Nelson, P. (1990). Geochemistry of the Taitao Ophiolite and near-trench intrusions from the Chile Margin Triple Junction. *Journal of South American Earth Sciences*, **3**(4), 161-177.

Kay, S.M., Ramos, V.A. and Márquez, M. (1993). Evidence in Cerro Pampa volcanic rocks for slab-melting prior to ridge-trench collision in southern South America. *Journal of Geology*, **101**, 703-714.

Lacassie, J.P. and Godoy, E. (1999). Superposición de plegamiento en el interfluvio Baker-Bravo, Complejo Metamórfico Andino Oriental, Aysén, Chile. *XIV Congr. Geol. Argentino.*

Lagabrielle, Y., Le Moigne, J., Maury, R.C., Cotten, J. and Bourgois, J. (1994). Volcanic record of the subduction of an active spreading ridge, Taitao Peninsula (southern Chile). *Geology*, **22**, 515-518.

Lagabrielle, Y., Bourgois, J., Suárez, M., De La Cruz, R., Garel, E., Dauteuil,

O. and Gutscher, M.A. From compressional to extensional tectonic regime at the front of the Patagonian Andes, 46°-47°S: a response to the subduction of the Chile spreading Ridge? *4th ISAG, Gottingen, Germany*, 1999. (in press).

Lagally, U. (1975). *Geologische Untersuchungen mi Gebiet Lake General Carrera-Lake Cochrane, Prov. Aisen/Chile unter besorender Berücksichtigung des Grundgebirges und seiner Tektonik.* Ludwig-Maximilians-Universität München.

Leanza, A.F. (1972). *Andes Patagónicos Australes.* Geología Regional Argentina, Academia Nacional de Ciencias, Córdoba, p. 689-706.

Le Moigne, J., Lagabrielle, Y., Whitechurch, H., Girardeau, J., Bourgois, J. and Maury, R.C. (1996). Petrology and geochemistry of the ophiolitic and volcanic suites of the Taitao Peninsula, Chile Triple junction area. *Journal of South American Earth Sciences*, 9, 43-58.

Ling, H.Y., Forsythe, R.D. and Douglass, C.R. (1985). Late Paleozoic microfaunas from Southernmost Chile and their relation to Gondwanaland forearc development. *Geology*, 13, 357-360.

Ling, H.Y. and Forsythe, R.D. (1987). Late Paleozoic pseudoalbaillellid radiolarians from Southernmost Chile and their geological significance. In: McKenzie, G.D. (ed.). Gondwana Six: Structure, Tectonics and Geophysics. *AGU Geophysical Monograph*, Washington 40, 253-260.

López-Escobar, L., Cembrano, J. and Moreno, H. (1995). Geochemistry and tectonics of the Chilean Southern Andes basaltic Quaternary volcanism (37°-46°S). *Revista Geológica de Chile*, 22(2), 219-234.

Marshall, L. and Salinas, P. (1990). Stratigraphy of the Río Frías Formation (Miocene), along the alto Río Cisnes, Aisén, Chile. *Revista Geológica de Chile*, 17, 57-87.

Martinic, M. (1988). Actividad volcánica histórica en la región de Magallanes. *Revista Geológica de Chile*, 15(2), 181-186.

Mercer, J.H. (1976). Glacial history of the southernmost South America. *Quaternary Research*, 6, 125-166.

Miller, H. (1975). El basamento de la provincia de Aisén (Chile) y sus correlaciones con las rocas Premesozoicas de la Patagonia Argentina. *VI Congreso Geológico Argentino*, 1, 125-141.

Miller, H. (1979). Unidades estratigráficas y estructurales del basamento andino en el Archipiélago de los Chonos, Aisén, Chile. *II Congreso Geológico Chileno*, Arica, 1, A103-A120.

Miller, H. and Sprechmann, P. (1978). Eine devonische Faunula aus dem Chonos-Archipel, Region Aisén, Chile und ihre stratigraphische Bedeutung. *Geologische Jahrbuch*, **B28**, 37-45.

Mpodozis, C., Hervé, M., Nasi, C., Soffia, J., Forsythe, R. and Nelson, E. (1985). El Magmatismo Plioceno de Península Tres Montes y su relación con la Evolución del Punto Triple de Chile Austral. *Revista Geológica de Chile*, **25-26**, 13-28.

Mpodozis, C. and Ramos, V.A. (1989). The Andes of Chile and Argentina. In: Geology of the Andes and its relation to hydrocarbon and mineral resources, eds. Ericksen, G.E., Cañas Pinochet, M.T. and Reynemund, J.A. *Circum Pacific Council for Energy and Mineral Resources Earth Sciences Series*, Houston, Texas, **11** (Chapter 5).

Muir Wood, R. (1989). Recent faulting at Laguna San Rafael Aisen province southern Chile. *Comunicaciones*, **40**, 57-68.

Mukasa, S.B. and Dalziel, I.W.D. (1996). Southernmost Andes and South Georgia Island, North Scotia Ridge: zircon U-Pb and muscovite $^{40}Ar/^{39}Ar$ age constraints on tectonic evolution of southwestern Gondwanaland. *Journal of South American Earth Sciences*, **9**, 349-365.

Murdie, R.E. (1994). *Seismicity and Neotectonics associated with the subduction of an Active Ocean Ridge-Transform System in Southern Chile.* PhD Thesis. University of Liverpool (unpublished).

Naranjo, J.A. and Stern, C.R. (1998). Holocene explosive activity of Hudson Volcano, southern Andes. *Bulletin of Volcanology*, **59**, 291-306.

Nelson, R., Forsythe, R., Hervé, F., Suárez, M., Valenzuela, E. and Wilson., T. (1977). Observaciones estructurales en la Cordillera Darwin, Prov. Antártica y Tierra del Fuego: crucero 77/4 del R/V Hero. Depto. de Geología, U. de Chile, Santiago, *Comunicaciones*, **21**, 32-35.

Nelson, E.P. (1982). Post-tectonic uplift of the Cordillera Darwin orogenic core complex: evidence from fission track geochronology and closing temperature-time relationships. *Journal of the Geological Society of London*, **139**, 755-761.

Nelson, E.P., Dalziel, I.W.D. and Milnes, A.G. (1980). Structural geology of the Cordillera Darwin: Collision style orogenesis in the southernmost Chilean Andes. *Ecologae Geologicae Helveticae*, 73, 727-751.

Nelson, E. and Forsythe, R. (1989). Ridge collision at convergent margins: implications for Archean and post-Archean crustal growth. *Tectonics*, **161**, 307-315.

Nelson, E., Forsythe, R., Diemer, J., Allen, M. and Urbina, O. (1993). Taitao ophiolite: a ridge collision ophiolite in the forearc of southern Chile (46° S). *Revista Geológica de Chile*, **20**, 137-166.

Nelson, E., Forsythe, R. and Arit, I. (1994). Ridge collision tectonics in terrane development. *Journal of South American Earth Sciences*, 7(3/4), 271-278.

Nelson, E.P. (1996). Suprasubduction Mineralization: Metallo-tectonic Terranes of the Southernmost Andes. In: Subduction: Top to the Bottom, eds. Bebout, G.E. et al. *AGU Geophysical Monograph*, Washington, **96**, 315-330.

Niemeyer, H. (1978). Nódulos máficos y ultramáficos en basaltos alcalinos de la Meseta Buenos Aires, Lago General Carrera, Provincia de Aysén, Chile. *Revista Asociación Geológica Argentina*, **33**, 63-75.

Niemeyer, H., Skarmeta, J., Fuenzalida, R., Espinosa, W. (1984). Hojas Península de Taitao y Puerto Aisén. *Servicio Nacional de Geología y Minería, Carta Geol. Chile*, **60-61**, 80 p.

Olivero, E.B. and Malumián, N. (1999). Eocene stratigraphy of southeastern Tierra del Fuego island, Argentina. *American Association of Petroleum Geologists*, **83**(2), 295-313.

Onuma, N., Tagiri, M., Takahashi, M., Notsu, K., Hasegawa, A., Lahsen, A., López-Escobar and Moreno, H. (1985). *Geochemical investigation of the Southern Andes Volcanic Belt, 1982-1984.* Overseas Scientific Research N°59043009, Ibaraki University and University of Chile.

Ozawa, T. and Kanmera, K. (1984). Tectonic terranes of Late Paleozoic rocks and their accretionary history in the circum-Pacific region viewed from Fusulinacean paleobiogeography. In: Proceedings of the Circum-Pacific Terrane Conference: Stanford University Publications, eds. Howell, D.G. et al. *Geological Sciences*, **18**, 158-160.

Pankhurst, R.J. and Rapela, C.R. (1995). Production of Jurassic rhyolite by anatexis of the lower crust of Patagonia. *Earth Planetary Science Letters*, **134**, 23-36.

Pankhurst, R.J., Leat, P.T., Sruoga, P., Rapela, C.W., Márquez, M., Storey, B.C. and Riley, T.R. (1998). The Chon-Aike silicic igneous province of Patagonia and related rocks in West Antarctica: a silicic large igneous province. *Journal of Volcanology and Geothermal Research*, **81**, 113-136.

Pankhurst, R.J., Weaver, S.D., Hervé, F. and Larrondo, P. (1999a). Mesozoic-Cenozoic evolution of the North Patagonian batholith in Aysén, southern Chile. *Journal of the Geological Society, London*, **156** (in press).

Pankhurst, R.J., Riley, T.R., Fanning, C.M. and Kelley, S.P. (1999b). Episodic silicic volcanism in Patagonia and the Antarctic Peninsula: chronology of magmatism associated with the break-up of Gondwana. *Journal of Petrology* (in press).

Parada, M.A., Palacios, C. and Lahsen, A. (1997). Jurassic extensional tectono-magmatism and associated mineralization of the El

Faldeo polymetallic district, Chilean Patagonia: geochemical and isotopic evidence of crustal contribution. *Mineralium Deposita*, **32**, 547-554.

Pardo-Casas, F. and Molnar, P. (1987). Relative motion of the Nazca (Farallon) and South American plates since Late Cretaceous times. *Tectonics*, **6**, 233-248.

Pearce, J.A. (1983). The role of sub-continental lithosphere in magma genesis at destructive plate margins. In: *Continental basalts and mantle xenoliths*, eds. Hawkesworth, C.J. and Norry, M.J., pp. 230-249. Shiva.

Petersen, U. (1990). Geological framework of Andean mineral resources. In: Geology of the Andes and its Relation to Hydrocarbon and Mineral Resources, eds. Ericksen, G.E., Cañas Pinochet, M.T. and Reinemund, J.A. *Circum-Pacific Council for Energy and Mineral Resources Earth Science Series*, **11**, 213-232.

Pino, M. (1976). *Reconocimiento geológico de los Departamentos Cochrane y Baker, Undécima Región, Aysén.* Tesis de prueba para optar al título de Geólogo, inédito, Depto. de Geología, Univ. de Chile.

Puig, A., Hervé, M., Suárez, M. and Saunders, A.D. (1984). Calc-alkaline and alkaline Miocene and calc-alkaline recent volcanism in the southernmost Patagonian Cordillera, Chile. *Journal of Volcanology and Geothermal Research*, **20**, 149-163.

Rabassa, J. and Clapperton, C.M. (1990). Quaternary glaciations of the southern Andes. *Quaternary Science Reviews*, **9**, 153-174.

Ramos, V.A. (1982). Las ingresiones pacificas del Terciario en el Norte de la Patagonia. *III Congreso Geológico Chileno, Concepción, Actas*, **I**(A), 262-288.

Ramos, V.A. (1984). La Patagonia: ¿un continente a la deriva? *IX Congreso Geológico Argentino*, Bariloche, **2**, 311-328.

Ramos, V.A. (1989). Andean Foothills Structures in Northern Magallanes Basin, Argentina. *The American Association of Petroleum Geologists Bulletin*, **73**(7), 887-903.

Ramos, V.A. and Kay, S. (1992). Southern Patagonian plateau basalts and deformation: backarc testimony of ridge collisions. *Tectonophysics*, **205**, 261-282.

Ray, F.M. (1996). *Stratigraphical and structural evolution of Tertiary backarc basins in southern Chile.* Ph.D. Thesis, University of Liverpool, England, 208 p. (unpublished).

Riccardi, A. (1971). Estratigrafía en el oriente de la Bahía de la Lancha, Lago San Martín, Santa Cruz, Argentina. *Revista del Museo de la Plata (nueva serie), Sección Geología*, **VII**, 245-318.

Riccardi, A.C. and Rolleri, E.O. (1980). Cordillera Patagónica Austral. In: Turner, J.C.M. (ed.) Geología Regional Argentina. *Segundo Simposio, Academia Nacional de Ciencias*, Córdoba, **II**, 1173-1306.

Riccardi, A.C. (1988). The Cretaceous System of Southern South America. *The Geological Society of America, Memoir*, **168**.

Riley, T.R. and Leat, P.T. (1999). Large volume silicic volcanism along the proto-Pacific margin of Gondwana: lithological and stratigraphical investigations from the Antarctic Peninsula. *Geological Magazine*, **136**, 1-16.

Rojas, C., Beck, M.E., Jr., Burmester, R.F., Cembrano, J., Hervé, F. (1994). Paleomagnetism of the mid-Tertiary Ayacara Formation, southern Chile: counterclockwise rotation in a dextral shear zone. *Journal of South American Earth Science*, **7**(1), 45-56.

Saunders, A., Tarney, J., Stern, C. and Dalziel, I.W. (1979). Geochemistry of Mesozoic margin basin floor igneous rocks from Southern Chile. *Geological Society of America Bulletin*, **90**, 237-258.

Skarmeta, J.J. and Castelli, J.C. (1997). Intrusión sintectónica del granito de las Torres del Paine, Andes Patagónicos de Chile. *Revista Geológica de Chile*, **24**(1), 55-74.

Stern, C.R., Frey, F.A., Futa, K., Zartman, R.E., Peng, K. and Kyser, T.K. (1990). Trace-elements and Sr, Nd, Pb and O isotopic composition of Pliocene and Quaternary alkali basalts of the Patagonian Plateau lavas of southernmost South America. *Contributions to Mineralogy and Petrology*, **104**, 294-308.

Stern, C.R., Mukasa, S.B. and Fuenzalida, R. (1992). Age and petrogenesis of the Sarmiento ophiolite complex of southern Chile. *Journal of South American Earth Sciences*, **6**, 97-104.

Stern, C. and Kilian, R. (1996). Role of the subducted slab, mantle wedge and continental crust in the generation of adakites from the Andean Austral Volcanic Zone. *Contributions to Mineralogy and Petrology*, **123**, 263-281.

Suárez, M., Prieto, X., Belmar, M. and Quiroz, D. (1994). Estratigrafía terciaria del sector nororiental de Aysén (44°-46° L.S.), Chile. *VII Congreso Geológico Chileno, Extended Abstracts*, 533-537.

Suárez, M. and De La Cruz, R. (1996). Estratigrafia y tectónica de la zona sureste del Lago General Carrera (46°30'-47°Lat. S), Cordillera Patagónica, Chile. *Actas XIII Congreso Geológico Argentino y III Congreso de Exploración de Hidrocarburos*, Acta I, 425-432.

Suárez, M. and De La Cruz, R. (1997). Cronología magmática de Aysén Sur, Chile (45°-48°30'L.S.). *Actas Octavo Congreso Geológico Chileno, Antofagasta, Chile*, **II**, 1543-1547.

Suárez, M., De La Cruz, R., Bell, C.M. (1996). Estratigrafía de la región de Coyhaique (Latitud 45°-46°S), Cordillera Patagónico, Chile. *Actas XIII Congreso Geológico Argentino y III Congreso de Exploración de Hidrocarburos*, I, 575-590.

Suárez, M., Demant, A. and De La Cruz, R. Volcanismo calcoalcalino en W Provincia Chon Aike: Grupo Ibáñez, Jurásico Superior-Cretácico Inferior temprano, Cordillera Patagónica de Aysén, Chile (45°30'-46°30'S). *Actas IV Congreso Geológico Argentino, Salta* (in press).

Suárez, M., De La Cruz, R. and Troncoso, A. Depósitos fluviales del Paleoceno (Formación Ligorio Márquez), Región de Aysén, Chile (46°30'S). *Actas XIV Congreso Geológico Argentino, Salta* (in press).

Thomson, S., Hervé, F., Stöckhert, B. and Brix, M. (1999a). Consequences of ridge collision, oblique subduction and arc-parallel strike-slip faulting in the southern chilean andes (44° to 46°S) using fission track data. *Terra Nostra, Abstracts*, **11**, 426.

Thomson, S.N., Hervé, F., Brix, M.R. and Stöckhert, B. (1999b). Fission-track thermochronology of the Southern Chilean andes (42°S to 48°S). *Fourth International Symp. Andean Geodynamics*, Göttingen.

Townley, B. (1996). *Ore deposits, tectonics and metallogenesis of the continental Aysén region, Chile.* Ph.D. Thesis, Ontario, Canadá, Queen's University, 246 p. (unpublished).

Trouw, R.A., Simões, L.S.A. and Valladares, C.S. (1998). Metamorphic evolution of a subduction complex, South Shetland Islands, Antarctica. *Journal of Metamorphic Geology*, **16**, 475-490.

Vargas, G. and Hervé, F. (1994). Mineralogía y condiciones de formación de las cavidades miarolíticas en el stock de Paso de Las Llaves, Aysén. *VII Congreso Geológico Chileno, Concepción. Actas*, II, 1259-1263.

Weaver, S.G., Bruce, R., Nelson, E.P., Brueckner, H.K. and LeHuray, A.P. (1990). The Patagonian batholith at 48° S latitude, Chile; Geochemical and isotopic variations. In: Kay, S.M. and Rapela, C.W. (eds). Plutonism from Antarctica to Alaska. Boulder, Colorado. *Geological Society of America Special Paper*, **241**, 33-50.

Welkner, D. (1999). *Geología del área del Cerro San Lorenzo: Cordillera Patagónica Oriental, XI Región de Aysén, Chile (47°25' a 47°50'S).* Memoria para optar al título de Geólogo, Depto. de Geología, Universidad de Chile (unpublished).

Willner, A., Hervé, F. and Massone, H.-J. Pressure-temperature evolution of the Chonos Metamorphic Complex, Southern Chile. *Journal* *of Petrology* (in press).

Wilson, T.J. (1991). Transition from back-arc to foreland basin development in the southernmost Andes: Stratigraphic record from the Ultima Esperanza District, Chile. *Geological Society of America Bulletin*, **103**, 98-111.

Wilson, T.J., Hanson, R.E. and Grunow, A.M. (1989). Multistage melange formation within an accretionary complex, Diego Ramírez Islands, southern Chile. *Geology*, **17**, 11-14.

Yoshida, K. (1981. *Estudio geológico del curso superior del río Baker, Aysén, Chile (47°05'a 47°42' Lat. S., 72°28' a 73°15' Long. W)*. Ph.D. Thesis, Depto. de Geología, Univ. de Chile (unpublished).

Tectonic Evolution of the Andes

Victor A. Ramos and Antenor Aleman

The Andes is the type section of a non-collisional orogen that formed a mountain chain by the subduction of oceanic crust under a continental plate, as proposed by Dewey and Bird (1970). However, the analysis of the different sectors of the Andes shows a great variety of processes and a vast complexity, including mountain-building processes, which do not fit into the original definition. Therefore, the Andes will be divided following the proposal of Gansser (1973) to examine these tectonic processes (Fig. 1). The proposed division, although descriptive, is appropriate to analyze the distinctive history of the different Andean segments.

The Andean history will be divided in four major stages. The first is related to the reconstruction of the proto-margin of Gondwana, and consists of amalgamation and collision of different terranes against the late Proterozoic margin of Gondwana. The second stage in the Late Paleozoic is linked to the formation of the Gondwanides, the first mountain chain developed along the Pacific margin by an Andean-type subduction; and the Alleghanides, related to the closure of the Iapetus Ocean and the formation of the Pangea Supercontinent. The third stage is related to a generalized extension during Pangea break-up that predates the opening of the South Atlantic and related oceans, and is punctuated by the collision of island arcs in the Northern Andes. The last stage is responsible for the present orogen, and includes a great variety of tectonic processes from collision of island arcs, seismic and aseismic ridges, as well as normal subduction of oceanic crust under the South American Plate, that defines the proper Andean-type.

The proto-margin of Gondwana

It is well established that the Gondwana Supercontinent was formed by accretion and amalgamation of different continental blocks and terranes as the result of the Panafrican-Brasiliano Orogeny (Brito Neves and Cordani, 1991; Hoffman, 1991; see recent review of Brito Neves *et al.*, 1999). Most of the western Gondwana margin was formed at that time, although accretion of different small cratonic blocks to this proto-margin, mainly during the Early Paleozoic (Ramos, 1988a; Restrepo-Pace, 1992; Toussaint, 1993), contributed to the final configuration of the present Pacific margin. Some of these cratonic blocks were exotic to the continent, whereas others were parautochthonous or perigondwanan. Perigondwanan terranes are those that were part of western Gondwana during Proterozoic times, but were split away and subsequently accreted to the margin (Ramos and Basei, 1997a; Keppie and Ramos, 1999).

The different terranes identified in the basement of the Andes are shown in Figure 2. In order to reconstruct the proto-margin of Gondwana it is necessary to subtract all the allochthonous terranes accreted during Phanerozoic times. Firstly, the oceanic terranes accreted during the Cenozoic in the Northern and Caribbean Andes such as the Choco Terrane of the Serranía de Baudó and its northern extension in Panamá (Dengo and Covey, 1993), and the Bonnaire Block of Venezuela (Kellogg and Bonini, 1982). This block formed by a series of nappes and oceanic rocks, detached and thrust from the Caribbean Plate, contains slivers of metamorphic rocks that have been defined as the Caribbean Terrane by Bellizzia and Pimentel (1994). This terrane was incorporated to South America during Late Cretaceous-early Paleogene times.

Secondly, it is required to subtract the Mesozoic Piñón-Daua terranes of the Ecuadorian and Colombian Andes (Alemán and Ramos, this volume), and their extension in the Amotape-Tahuín Terrane (Feininger, 1987; Reynaud *et al.*, 1999). However, due to the important sinistral strike-slip displacements produced by the oblique subduction during Meso-Cenozoic times, a palinspastic restoration is needed, similar to the proposed reconstruction of the Alleghenides by Pindell (1985). In this reconstruction the Santa Marta and Santander massives, as well as part of the Central and Eastern Cordilleras should be located farther S (present coordinates) of their present position.

This restoration permits the reconstruction of a large piece of basement named the Chibcha and related terranes (Etayo-Serna and Barrero, 1983; Alemán and Ramos, this volume) or Central Andean terranes by Restrepo-Pace (1992). One common feature of the basement of these terranes is the Grenville signature, based on the Pb-isotope composition (Ruiz *et al.*, 1999), as well as the geochronology and Nd-isotope data of the Colombian terranes (Restrepo-Pace *et al.*, 1997). This large terrane is presently the basement of Eastern Cordillera, and it was interpreted as an allochthonous terrane derived from North America by Forero Suárez (1990), that during the early closure of the Iapetus Ocean was transferred to the proto-margin of Gondwana in pre-Emsian times (pre-late Early Devonian). The magmatic arc at that time was developed on the Paleozoic terrane. The Borde Llanero-Guaicáramo Fault that uplifted the Eastern Cordillera foothills on the foreland basin deposits approximately coincides with the suture between the Paleozoic allochthons and the proto-margin of Gondwana.

Paleozoic rocks in the Ecuadorian Andes consist of semipelitic schist and gneiss, highly deformed and intruded by S-type Triassic granitoid plutons (Aspden and Litherland,

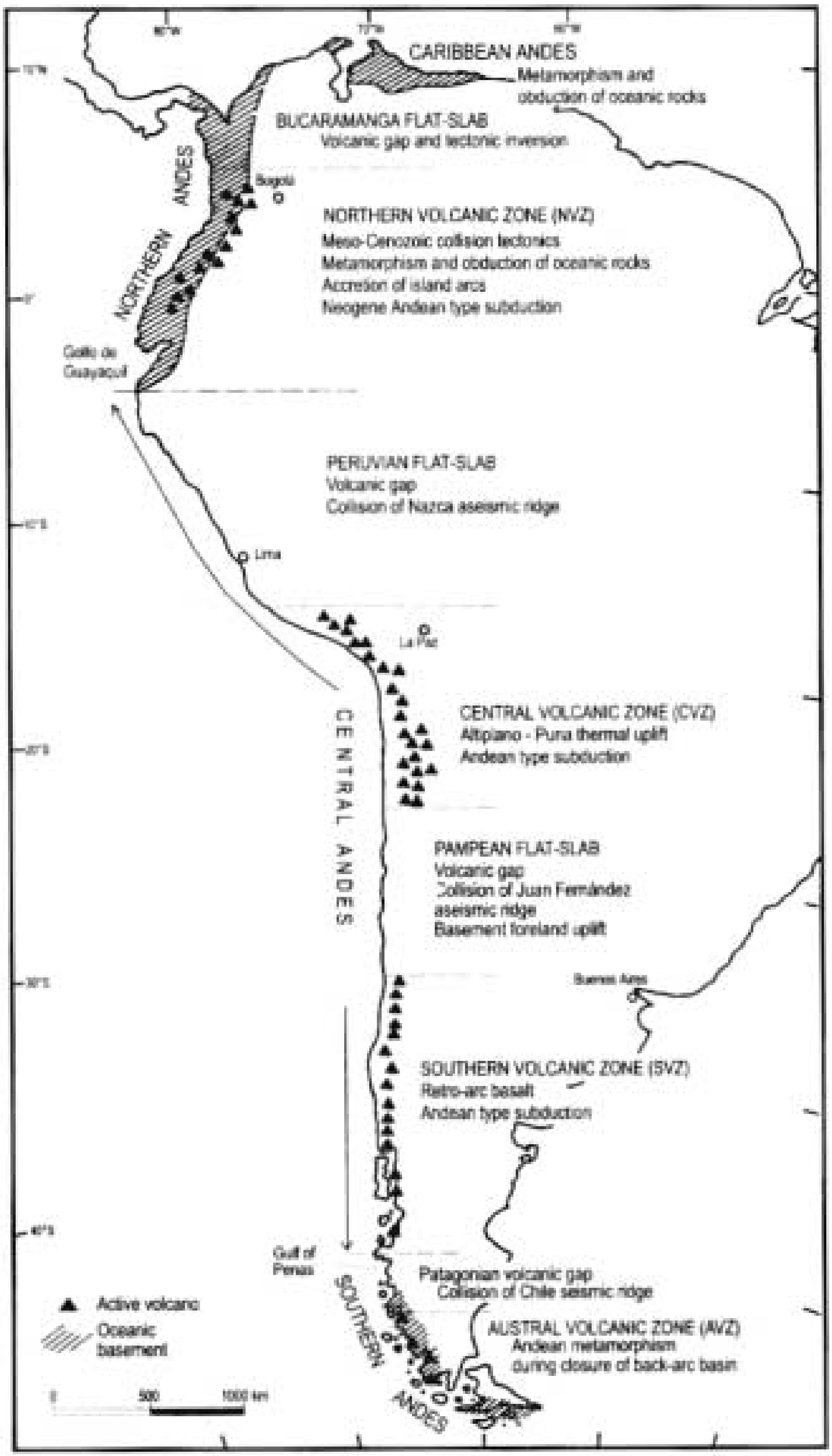
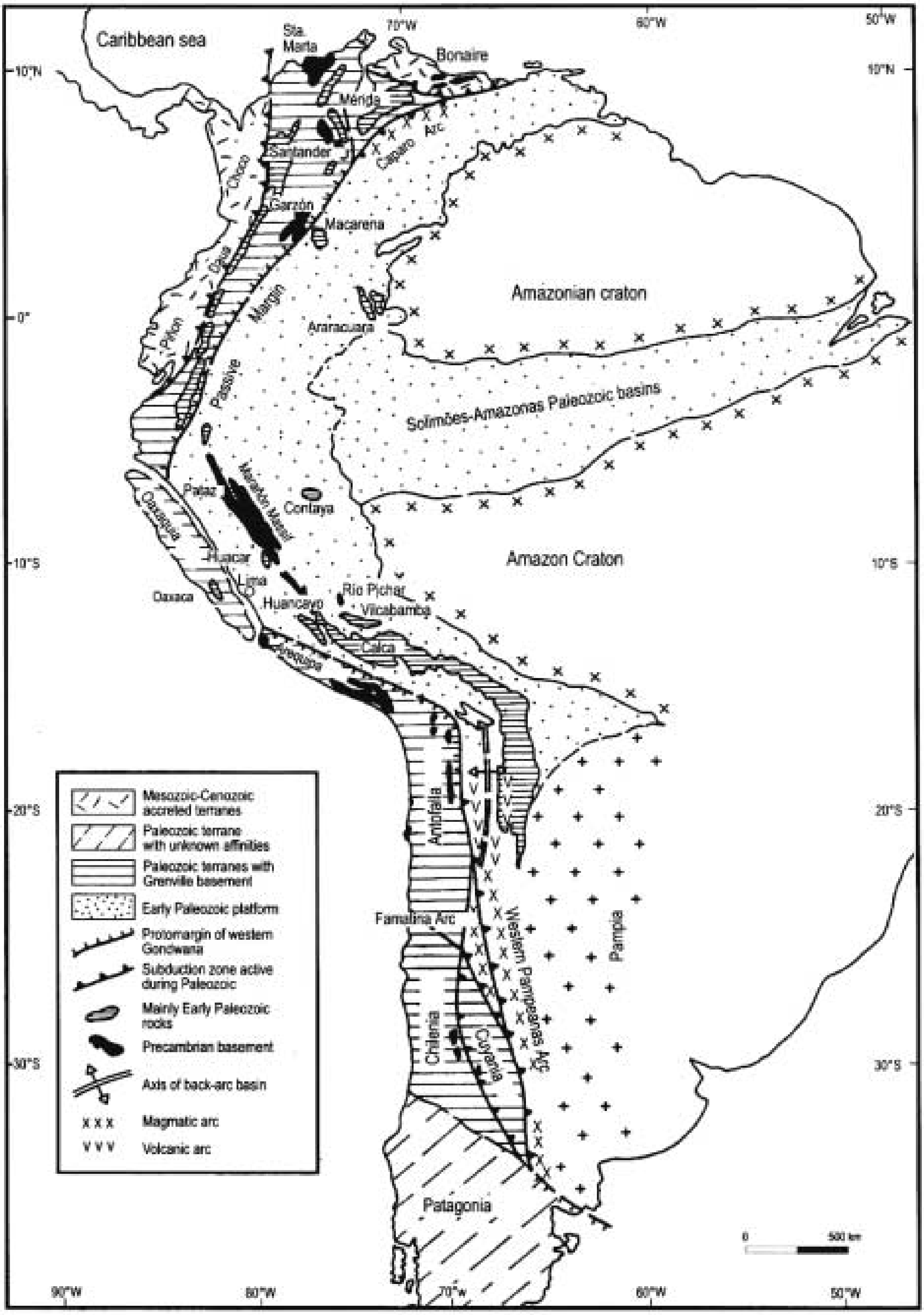


FIGURE 1: Main segments of the Andes modified after Gansser (1973), with indication of major tectonic processes involved during their formation.

FIGURE 2: Reconstruction of the proto-margin of western Gondwana with indication of the main Precambrian and Early Paleozoic blocks underlying the Andean cover (modified after Bellizzia and Pimentel, 1994; Restrepo-Pace, 1992; Restrepo-Pace et al., 1997; Keppie and Ortega Gutiérrez, 1999; Ramos, 1988a).


Caribbean sea
Sta. Marta
Bonaire
Mérida
Santander
Caparo Arc
Chocó
Garzón
Macarena
Dagua
Margin
Passive
Piñón
Araracuara
Amazonian craton
Solimões-Amazonas Paleozoic basins
Pataz
Marañón Massif
Contaya
Amotape
Huacar
Lima
Oaxaca
Huancayo
Rio Pichar
Vilcabamba
Amazon Craton
Calca
Arequipa
Antofalla
Famatina Arc
Western Pampeanas Arc
Pampia
Chilenia
Cuyania
Patagonia
Mesozoic-Cenozoic accreted terranes
Paleozoic terrane with unknown affinities
Paleozoic terranes with Grenville basement
Early Paleozoic platform
Protomargin of western Gondwana
Subduction zone active during Paleozoic
Mainly Early Paleozoic rocks
Precambrian basement
Axis of back-arc basin
Magmatic arc
Volcanic arc
0
500 km
90°W
80°W
70°W
60°W
50°W
10°N
0°
10°S
20°S
30°S

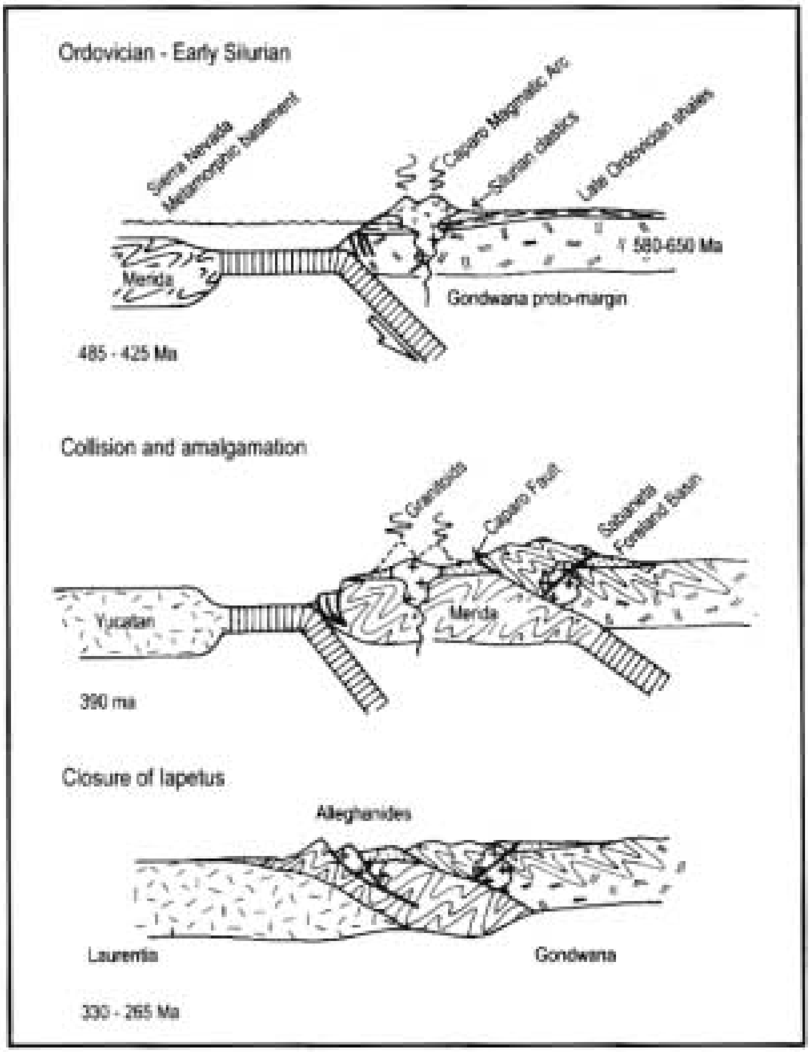


FIGURE 3: Schematic evolution of the Mérida Terrane, and the pre-Late Permian accretion of the Yucatán Block during the closure of the Iapetus Sea (modified after Bellizzia and Pimentel, 1994; Pindel, 1985).

1992). They could be relicts of the allochthonous terranes described farther N in Colombia, deformed during the successive collisions that affected the Eastern Cordillera or Cordillera Real.

Farther S in Peru, the basement of the Arequipa Massif is truncated, either by tectonic erosion due to subduction (Von Huene and Scholl, 1991), or by detachment of a large piece of basement, as was proposed by Dalziel (1994). The location of the Oaxaquia Terrane in Figure 2, is one of the alternatives proposed by Keppie and Ortega Gutiérrez (1999) to explain the gap along the margin between the Arequipa Massif and the Chibcha and related terranes of Colombia.

This alternative shows a fringe of Grenville terranes overlain by Early Paleozoic deposits. Some of them such as the Oaxaquia Terrane, have as in Oaxaca, a Late Cambrian-Early Ordovician cover with *Parabolina argentina* (Robison and Pantoja-Alor, 1968), a typical Gondwana trilobite very frequent in northern Argentina (Harrington and Leanza, 1957). The correlation between Northern Argentina, and the Oaxaca fauna is also enhanced by the fauna found by Moya *et al.* (1993) in northern Puna of Argentina over the Arequipa-Antofalla Terrane.

Another point to be noted, is the Paleozoic cover of these terranes. There are Cambro-Ordovician trilobite-bearing sequences as in the Garzón Massif (Ruiz *et al.*, 1999) or graptolite-bearing as in Central Cordillera (Mojica *et al.*, 1988), partially preserved in metamorphic facies and associated with Ordovician magmatic rocks. These terranes are bounded to the E by a Cambrian platform in Venezuela and Colombia (Bordonaro, 1992), and by more extensive and well-described Ordovician platform deposits from northern Argentina to Venezuela. This shallow water platform developed on the Gondwana autochthonous basement faced deep-water deposits as proposed in the paleogeographic reconstruction of Moya (1988) and Aceñolaza (1992).

Based on these characteristics this fringe of basement terranes of Grenville affinities are interpreted as part of the Rodinia Supercontinent (Hoffman, 1991; Dalziel, 1991 and subsequent papers), that were attached to the Amazonian Craton after collision and amalgamation. The active margin was located in the Sunsas Belt of Amazonia, whereas Laurentia was the passive margin as proposed by Sadowski and Bettencourt (1994). The separation between Laurentia and Amazonia left behind some Grenville terranes in South America. This late Proterozoic rifting and the subsequent Early Paleozoic extension were responsible for separation of part of the terranes. For example Oaxaquia, left western Gondwana after the Early Ordovician, while other blocks remained as perigondwanan terranes (Ramos and Basei, 1997b). Part of the Grenville sutures were the foci of new extension in Cambro-Ordovician times developing a passive margin, and the resulting subsidence formed the extensive clastic platform that characterized the Ordovician paleogeography from Venezuela to northern Argentina (Aceñolaza, 1992).

There is an exception along this Early Paleozoic passive margin in the Caparo Arc developed in eastern Venezuela (Bellizzia and Pimentel, 1994). This area shows magmatic rocks emplaced in deformed Ordovician and Silurian fossiliferous deposits that are unconformably overlying autochthonous Precambrian basement (650 - 580 Ma) of western Gondwana (Marechal, 1983). The syn-tectonic granitoid rocks range from Early Ordovician to Early Silurian (495-425 Ma), whereas the post-tectonic granites are Devonian (Fig. 3). This magmatic activity recorded the approximation and collision of the Mérida Terrane by the end of the Ordovician and beginning of the Silurian (Bellizzia and Pimentel, 1994). Later on, once amalgamated to western Gondwana, the Mérida Terrane was affected by deformation and the Sabaneta Formation was deposited in a peripheral foreland basin. Subsequent deformation followed the closure of the Iapetus Ocean and the collision of Laurentia with the northwestern sector of Gondwana to form the Alleghenides in the Late Permian (Pindel, 1985).

Some of the detached basement blocks such as the Chibcha and related terranes, were separated by oceanic crust, and later after subduction (Fig. 4) re-amalgamated to western Gondwana (Restrepo-Pace *et al.*, 1997).

Extension along the Arequipa Massif and its southern prolongation in the Antofalla basement was probably controlled by the previous Grenville-age suture (Suárez-Soruco, 1999). This extension formed an important Ordovician to Devonian basin along southern Peru, Bolivia and northern Argentina (Sempere, 1995). In northern Bolivia there is no evidence of Ordovician oceanic crust, whereas to the S there are oceanic rocks (Allmendinger *et al.*, 1982; Bahlburg and Hervé, 1997).

The history of the southern Arequipa-Antofalla Block is complex, as it records a late Proterozoic collision during Pampean Orogeny (approximately 530 Ma) (Omarini *et al.*, 1999), subsequent extension during Cambrian and Ordovician times (Ramos, 1988a), and final amalgamation during Late Ordovician (Coira *et al.*, 1982, 1999; Dalziel and Forsythe, 1985), resulting in the Ocloyic Orogeny (Ramos, 1986; Bahlburg and Hervé, 1997).

The western Puna magmatic arc was active from Late Cambrian to Middle Ordovician times (Palma *et al.*, 1987; Bahlburg, 1990). The inception and northern extension of this Early Paleozoic subduction zone, is not that evident, and it could be underlying the thick Andean cover.

The Early Paleozoic proto-margin of northwestern and western Argentina is better constrained. The development of a calc-alkaline series of granitoid rocks of Late Cambrian to Middle Ordovician age in western Sierras Pampeanas, has permitted the identification of an Early Paleozoic magmatic arc (Ramos, 1988a; Toselli *et al.*, 1996; Pankhurst *et al.*, 1998). The distribution and magmatic characteristics show two different belts: one with volcanic and plutonic rocks in the Famatina Terrane (Toselli *et al.*, 1996; Quenardelle and Ramos, 1999), and another restricted to western Sierras Pampeanas (Rapela *et al.*, 1999). Figure 5 summarizes the tectonic evolution and subsequent amalgamation of these terranes. This magmatic arc was related to the accretion of

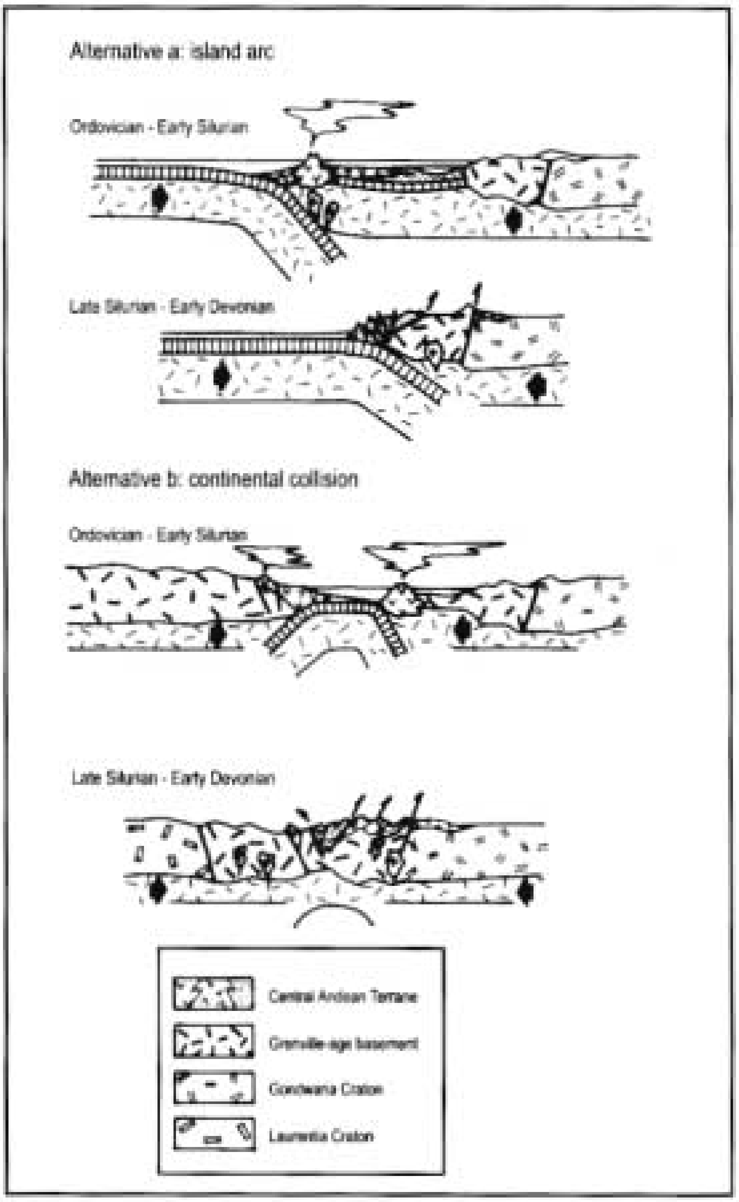


FIGURE 4: Different scenarios to explain the magmatic activity in the Chibcha (Central Andean) and related terranes of Colombia (modified after Restrepo et al.*, 1997).*

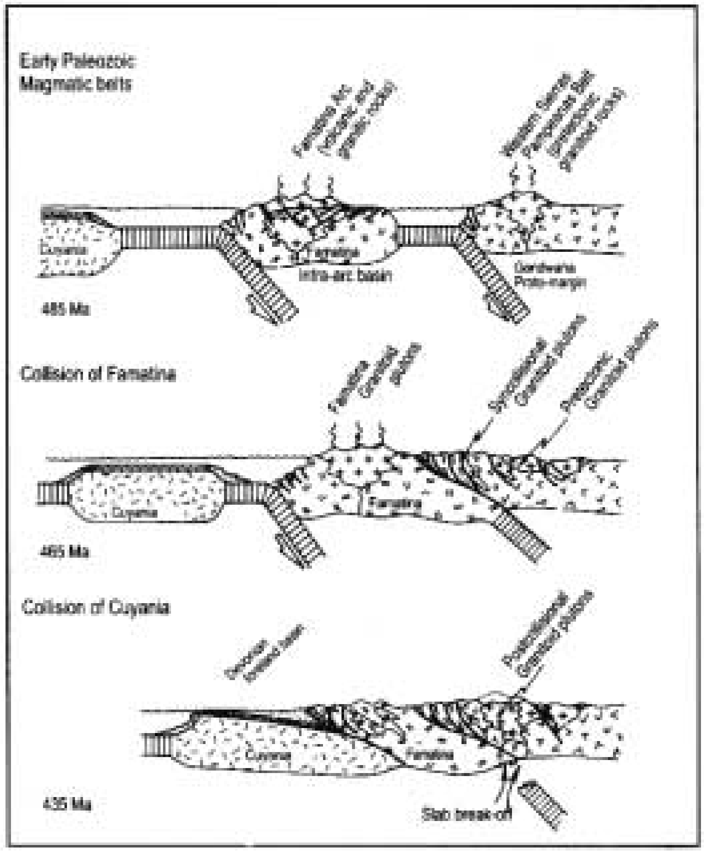


FIGURE 5: Tectonic evolution of western-central Argentina during Early Paleozoic times (modified after Quenardelle and Ramos, 1999).

FIGURE 6: Major acid and bimodal provinces developed along the western Gondwana margin during Late Paleozoic-Early Jurassic times. On the cratonic side mafic rocks dominated, while more acid rocks formed on a fringe of Early Paleozoic terranes along the proto-continental margin (modified after Mpodozis and Kay, 1992; Hervé, 1988).

the Cuyania composite terrane, containing the Precordillera, an allochthonous carbonate platform derived from Laurentia and accreted to the Gondwana proto-margin during Middle Ordovician times (Astini and Thomas, 1999). Another Grenville basement block, known as the Chilenia Terrane, collided against the western margin of Precordillera during Devonian times (Ramos *et al.*, 1986).

Farther S in the Patagonian Andes the Precambrian basement is scarce. Most of the gneiss and medium grade metamorphic rocks assumed to be Precambrian have U/Pb Paleozoic ages (Ramos and Aguirre-Urreta, this volume). These rocks are equivalent to the accretionary prism of Late Paleozoic age preserved along the Pacific coast (Hervé, 1988).

Alleghanian and Gondwanan Orogenies

The Late Paleozoic rocks of the Andes record two important mountain-building events: the Alleghanian Orogeny in the Northern Andes and the Gondwanan Orogeny in the Southern Andes.

The Alleghany Orogen was proposed by Woodward (1958) to encompass what in the old literature was known as the "Appalachian Revolution", responsible for the final deformation and uplift of the Appalachian System. As a consequence of the final closure of the Iapetus Ocean, Laurentia collided against western Gondwana, the present Africa and South America continents. The Alleghanides were diachronously formed between Late Carboniferous and Permian, encompassing a time span from 330 to 265 Ma (Hatcher *et al.*, 1989).

The Gondwanan Orogeny was first proposed by Keidel *(1916, 1921)*, to include the Late *Paleozoic* deformation of Sierra de la Ventana in the extra-Andean province of Buenos Aires, and the previously adjacent Cape Town System (Keidel, 1916, 1917). The Gondwanides were reconstructed from eastern Australia, through western Antarctica, South Africa to finally join the Ventania System and the Precordillera de Cuyo in the present foothills of the Andes (Du Toit, 1937; Groeber, 1938).

The Northern Andes

The final collision and amalgamation of the Mérida and other terranes into the proto-margin of Gondwana, was one of the most evident effects of the Alleghanian Orogeny in the Andean basement of Venezuela (Figs. 2 and 4). This deformation predated the Late Permian and was interpreted by Pindell (1985) as the result of final docking of the Yucatán Terrane, trapped between Laurentia and Gondwana during the closure of the Iapetus Ocean. Early Carboniferous rocks of the Mérida Cordillera underwent important deformation (Marechal, 1983). The compressive deformation was followed by transtension with widespread granitic plutonism from Early Permian to Triassic (290-225 Ma). This transtension was related to counterclockwise rotation between western Gondwana and North America, as proposed by Rapalini and Vizán (1993) on paleomagnetic grounds.

Farther S in the Colombian Andes, the Alleghanian event is characterized by strike-slip deformation related to the final closure of the Iapetus Ocean, associated with some S-type granitoid plutons (McCourt and Feininger, 1984).

The Central Andes

The Eo-Hercynian deformation of southern Peru and Bolivia proposed by Mégard *et al.* (1971), is temporally restricted to a Late Devonian-Early Carboniferous transpression (Díaz Martínez, 1996). This transpression has been detected farther S in the Tarija Basin (Fernández Seveso *et al.*, 1993) and in the basins of west-central Argentina. In the Paganzo and Río Blanco basins, at both sides of the proto-Precordillera, Limarino *et al.* (1999) recognized during the Late Paleozoic two different stages. A foreland basin stage as a result of the Chanic deformation at the end of the Devonian and beginning of the Early Carboniferous, that persisted until the Late Carboniferous; an extensional phase associated with alkaline basalt in the Early Permian. The first stage could be related to transpression as the result of oblique convergence and block rotation as identified by Rapalini and Vilas (1991) based on paleomagnetic data.

The Late Paleozoic along the Pacific margin was characterized by an important magmatism. It is represented by alkali-basalt and associated leucogranite and rhyolite of the Mitu Group (260-190 Ma) in southern Peru and northern Bolivia (Kontak *et al.*, 1985). These Permian to Triassic rocks are associated with red-beds in rift basins.

There is also a series of important batholiths and volcanic rocks formed during Late Paleozoic-Early Triassic times in southern Peru, Chile and central Argentina (Kontak *et al.*, 1985; Kay *et al.*, 1989). The magmatic suites have been divided in a subduction related Carboniferous-Early Permian series, and an extensional mainly granitic and rhyolitic series known as the Choiyoi Province (Mpodozis and Kay, 1992; Llambías and Sato, 1995).

Along the Principal Cordillera a strong angular unconformity is related to the San Rafael Orogeny (Ramos *et al.*, 1996b), which caused strong deformation in the Late Carboniferous-Early Permian rocks, separated by an angular unconformity of the Choiyoi volcanic rocks. This diastrophism produced an important penetrative deformation in the Late Paleozoic rocks of the Frontal Cordillera, but in the adjacent Precordillera the effects are only seen by an increase in subsidence rates in the late Early Permian basins as described by Fernández Seveso *et al.* (1993).

The Late Paleozoic marked the beginning of the subduction in the Pacific margin for the first time along the present trench. Most authors have related the Late Paleozoic deformation to changes in the intensity and direction of the convergence vector as inferred by paleomagnetic data (Ramos, 1988b; Kay *et al.*, 1989). Another tectonic alternative was proposed by Mpodozis and Kay (1992). These authors related the deformation of the San Rafael Orogeny to the docking of an enigmatic *terrane X*, which is not presently preserved due either to tectonic erosion or strike-slip displacements along the continental margin.

The subduction related magmatism was followed by an extensional regime. The extensional magmatic activity can be tracked (Fig. 6), as well as the younger Jurassic associated rocks, from Peru all along the western margin of Gondwana trough Antarctica to eastern Australia (Mpodozis and Kay, 1992).

Southern Andes

The Gondwanides Orogeny on the western slope of the Patagonian Andes is characterized by an extensive accretionary prism where high pressure-low temperature metamorphic rocks are cropping out (Hervé, 1988). These rocks farther N are partially preserved due to the tectonic erosion of the fore-arc region (Stern, 1991). Late Paleozoic magmatic rocks are poorly exposed, mainly in the eastern slope of the cordillera, where scarce tonalite and other granitoid rocks have Late Carboniferous-Early Permian ages (Ramos, 1983).

Farther S along the present coast, minor exotic terranes have been identified in the Madre de Dios region by Mpodozis and Forsythe (1983), who confirmed the hypotheses proposed previously by Helwig (1972). Mafic and ultramafic oceanic rocks compose the accreted rocks, and fusulinid-bearing platform carbonates associated with flysch and pelagic chert facies, which were docked into the Gondwanan subduction complex. Recent paleomagnetic studies demonstrate conspicuous rotation after their emplacement (Rapalini *et al.*, 1999). These oceanic exotic rocks may have been derived from platform carbonate patches coeval with the Copacabana Limestone developed at the latitude of present Bolivia.

The Early Mesozoic extension

Most of the Andes were dominated by extension at the end of the Permian and Triassic times. This extension was the precursor of the Pangea break-up. The Northern Andes continental margin (Fig. 7) was the conjugate rifted margin of the Yucatán Block and of the extensive present platform of U.S. Gulf and eastern Mexico. The reconstruction of the North American margin is difficult due to the intense crustal attenuation distributed along 760 km-wide zone (Pindell, 1985). However, the South American rifted margin is even more difficult to reconstruct due to the superimposed deformation produced by the Cretaceous and younger oceanic terrane accretion in Colombia and Ecuador, and the Caribbean deformation during emplacement of the Bonaire Block. These regions were displaced during Cenozoic times by important wrenching that obliterated the original margin.

The Northern Andes continental margin was developed during Triassic-Jurassic times in a series of *en échelon* rifts of NE-SW trend, such as the Espino, Uribante, Barquismeto, Mérida, and Machiques (Fig. 7), among others, which concentrated active extensional faulting until Late Jurassic times (Parnaud *et al.*, 1995). The tectonic inversion of the Machiques Rift exposed in the Sierra de Perija felsic to mafic igneous rocks, interfingered with red-beds, as well as the characteristic red-bed sequences of the La Quinta Formation of the Barquismeto and Mérida rifts, cropping out in the Serranía de Trujillo and the Mérida Andes (Lugo and Mann, 1995).

Small plutons along the Magdalena Valley of Colombia are precursor of the Late Triassic extension. Conglomerate, breccia, and sandstone form the initial fill of half-graben systems such as the Bogotá Trough, with a northeastern trend similar to previously described rifts farther N (Mojica and Dorado, 1987; Mojica *et al.*, 1996).

Highly deformed S-type granites of Late Triassic age are coeval with continental to volcaniclastic sequences in Ecuador (Litherland *et al.*, 1994). Subsequent terrane collision and closure of back-arc basins have obliterated original paleogeography.

The foreland region of Peru shows an intricate pattern of Triassic rifts, superimposed on the Late Paleozoic extension. Thick sequences of evaporites and red-beds of Late Triassic to Jurassic age of the Pucará Group constitute the initial syn-rift deposits (Mathalone and Montoya, 1995). This rifting postdates the Mitu Group rifts and alkaline magmatism.

Farther S, minor alkaline basalt of Entre Ríos (233 Ma) and Tarabuco (171 Ma) in Bolivia are associated with the Triassic and Early Jurassic rifting (Sempere, 1995). Near Lake Titicaca the red-beds fill the Mitu Group grabens, and along the foreland Subandean areas fluvial deposits of the Serere Formation of Late Triassic-Early Jurassic age are linked to mild extension.

The Triassic rift system of the Andes of Argentina and Chile has a dominant NW-SE trend (Charrier, 1979; Uliana and Biddle, 1988). The Triassic extension followed the rifting that began with the Choiyoi Volcanics, and was developed over the fringe of Paleozoic terranes accreted to cratonic South America (Kay *et al.*, 1989). The age of the normal faulting progressed from N to S. It was Late Carboniferous to Permian in northern Chile, Late Permian in Central Argentina and Triassic to Early Jurassic in Patagonia, predating the opening of the South Atlantic Ocean (Mpodozis and Ramos, 1990).

The rift system was filled by red-beds, interfingered with volcanic rocks of bimodal composition in northern and central Chile (Suárez and Bell, 1993; Mpodozis and Cornejo, 1997). Thick sequences of red-beds and lacustrine deposits, interfingered with alkaline basalt (233 Ma) are well represented in the Cuyo Basin (Kokogián *et al.*, 1993). Farther S these occur in the subsurface of the Neuquén Embayment along both sides of the Colorado River and in the Huincul area.

The geometry of the Triassic-Early Jurassic rifting of the Andes as a whole, denotes three different regions controlled by the extension and characteristics of the basement terranes (Fig. 7). The Northern Andes trend NE-SW in an *en échelon* pattern, matching the extension with a counterclockwise rotation developed during the development of the continental margin. The central region of Peru and Bolivia has a rift system with a dominant NW-SE trend, developed in the hanging-wall of the suture between the Arequipa Terrane and the Amazonian Craton. This suture developed in Grenville times indicates that Arequipa subducted beneath the Amazonian Craton, that was the active continental margin at that time (Sadowski and Bettencourt, 1994). The NW-SE rifting direction in the southern region was heavily controlled by the basement fabric. The inception of most of the rifts has been located along the suture hanging-wall of the previous Paleozoic accreted terranes (Ramos and Kay, 1991).

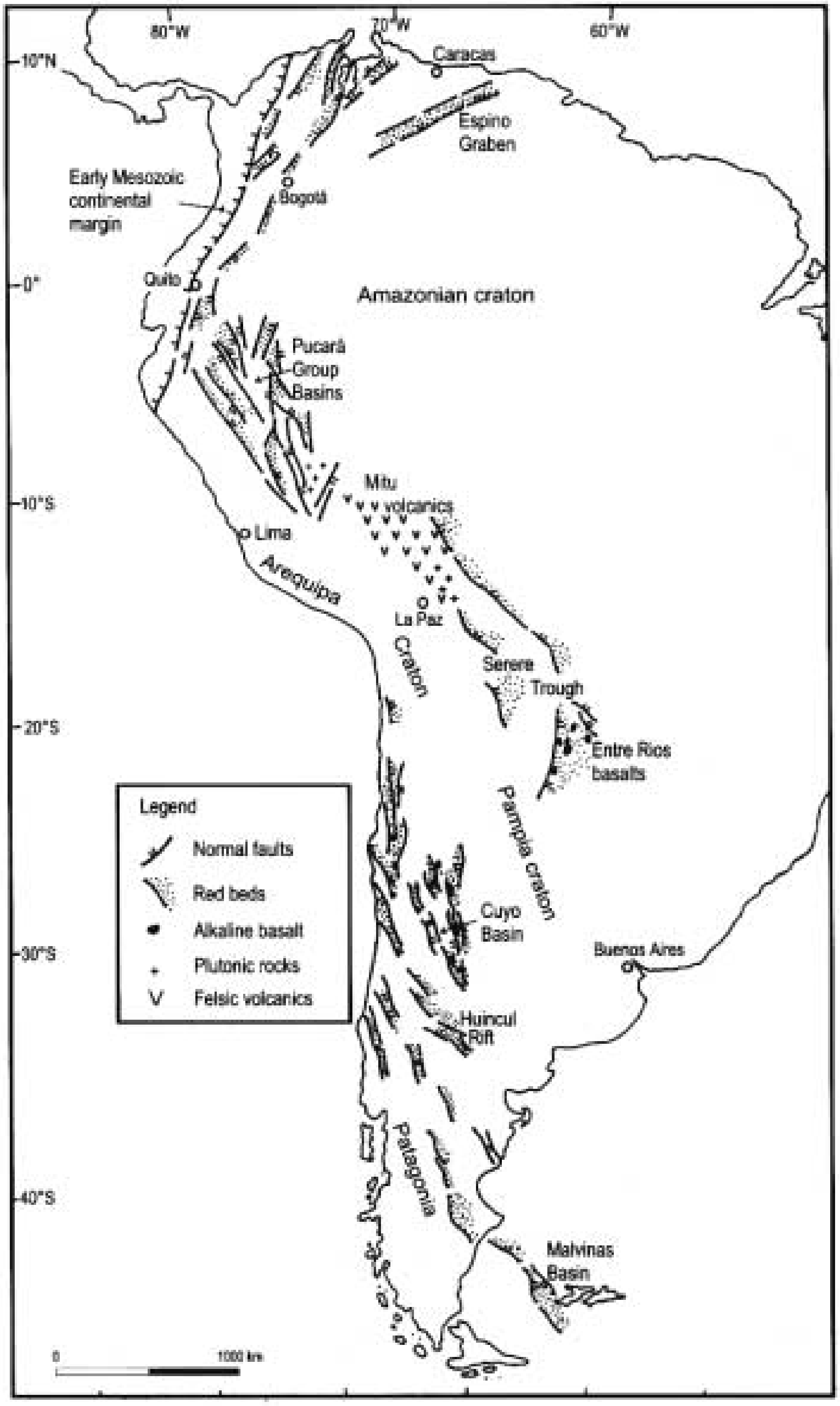


FIGURE 7: Major rift trends developed during the Pangea break-up along the margin of the Northern and Central Andes, as well as the generalized extension of the Southern Andes during Late-Triassic and Jurassic times. Configuration of the continental margin prior to the accretion of oceanic terranes is indicated in the Northern Andes. (modified after Parnaud et al., 1995; Mojica and Dorado, 1987; Daly, 1989; Mathalone and Montoya, 1995; Uliana and Biddle, 1988).

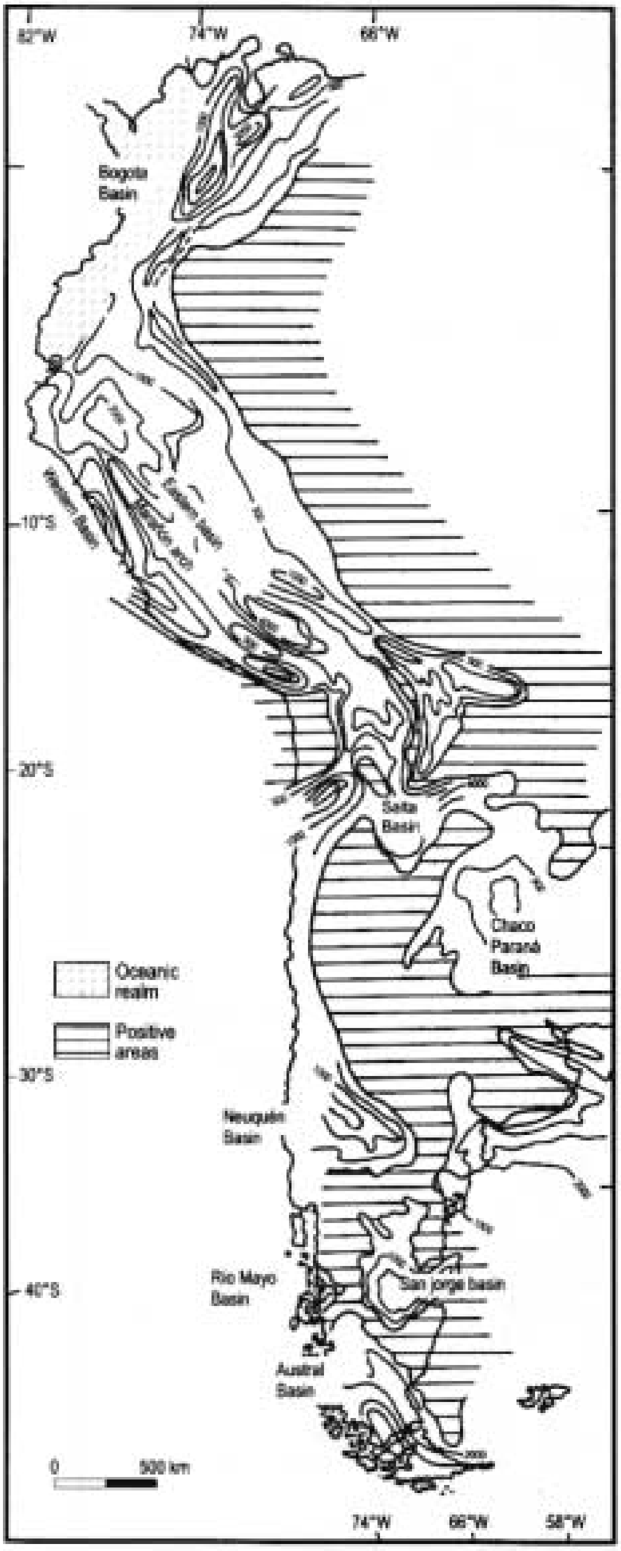


FIGURE 8: Cretaceous paleogeography of the Andes with isopach contours in metres (modified after Macellari, 1988; Ramos and Aguirre-Urreta, 1994 Salfity and Marquillas, 1994).

Passive margin and retro-arc Thermal Subsidence Stage

The entire Andean system records during Jurassic and Cretaceous times the development of a complex series of fore-arc, intra-arc and retro-arc basins that have been recently reviewed by several authors (Zambrano, 1987; Mojica and Dorado, 1987; Salfity, 1994; Benavides-Cáceres, 1999; Villamil, 1999; Jaillard *et al.*, this volume). The crustal fabric, heavily controlled by the previous accretionary history governs this complexity. Although the transition from active extensional faulting to thermal subsidence had occurred in Cretaceous times, the details and precise timing vary from one place to another. Some of these sag sequences have remained without deformation until Tertiary times, while others have had important precursor orogenies, during the Cretaceous, predating the Andean deformation.

Soon after the Late Jurassic most of the Northern Andes continental margin went into thermal subsidence developing a passive margin (Fig. 8), where the Cretaceous marine sediments were accommodated (Pindell, 1985). Isopach maps indicate that subsidence was maximum along the earlier depocenters such as Machiques, Barquismeto and Uribante rifts (Macellari, 1988). The end of the Early Cretaceous was marked by an important paleo-oceanographic change, consisting of a generalized drowning of the carbonate platform and deposition of intervals rich in organic matter in an anoxic environment that extended into the foothills at the Barinas-Apure basins (Erlich *et al.*, 1999). These environments were interrupted by the uplift of the Central and Eastern Cordilleras during Campanian to Maastrichtian times (Villamil, 1999). The platform deposits bounded the Brazilian Craton with important depocenters such as the Upper Magdalena Valley (Etayo-Serna, 1994) or the Bogotá Trough, where several thousand metres of Cretaceous sediments were deposited (Macellari, 1988).

Farther S in the Ecuador retro-arc area, marine platform sedimentation was important from Albian times (Macellari, 1988; Jaillard *et al.*, 1995). In Peru, two distinct longitudinal depocenters are recognized: the Eastern Basin along the Brazilian Craton and the Western Basin, separated by the Marañón Arch extended from 6°S to 14°S (Benavides-Cáceres, 1999). This shelf sedimentation was coeval with the arc volcanism along the margin.

Active rifting with continental deposition persisted from the Late Jurassic to the Early Cretaceous in Bolivia in a distal retro-arc setting (Sempere, 1995). Isolated depocenters of the Puca Group red-beds were interfingered with alkali basalt. The thermal sag sedimentation began in Albian times, when marine limestone beds overlap the Andean depocenters.

The active extensional faulting in eastern Bolivia and northwestern Argentina was younger than farther N in the Andes, and it was closely related to the opening of the South Atlantic Ocean. This Early Cretaceous extension was active from the volcanic arc in Chile (Mpodozis and Allmendinger, 1993), to central Argentina where the old Brasiliano sutures were reactivated, as well as along the Atlantic margin (Fig. 9).

This extension was coeval with the Salta rift basin in northern Argentina where thick syn-rift deposits and related

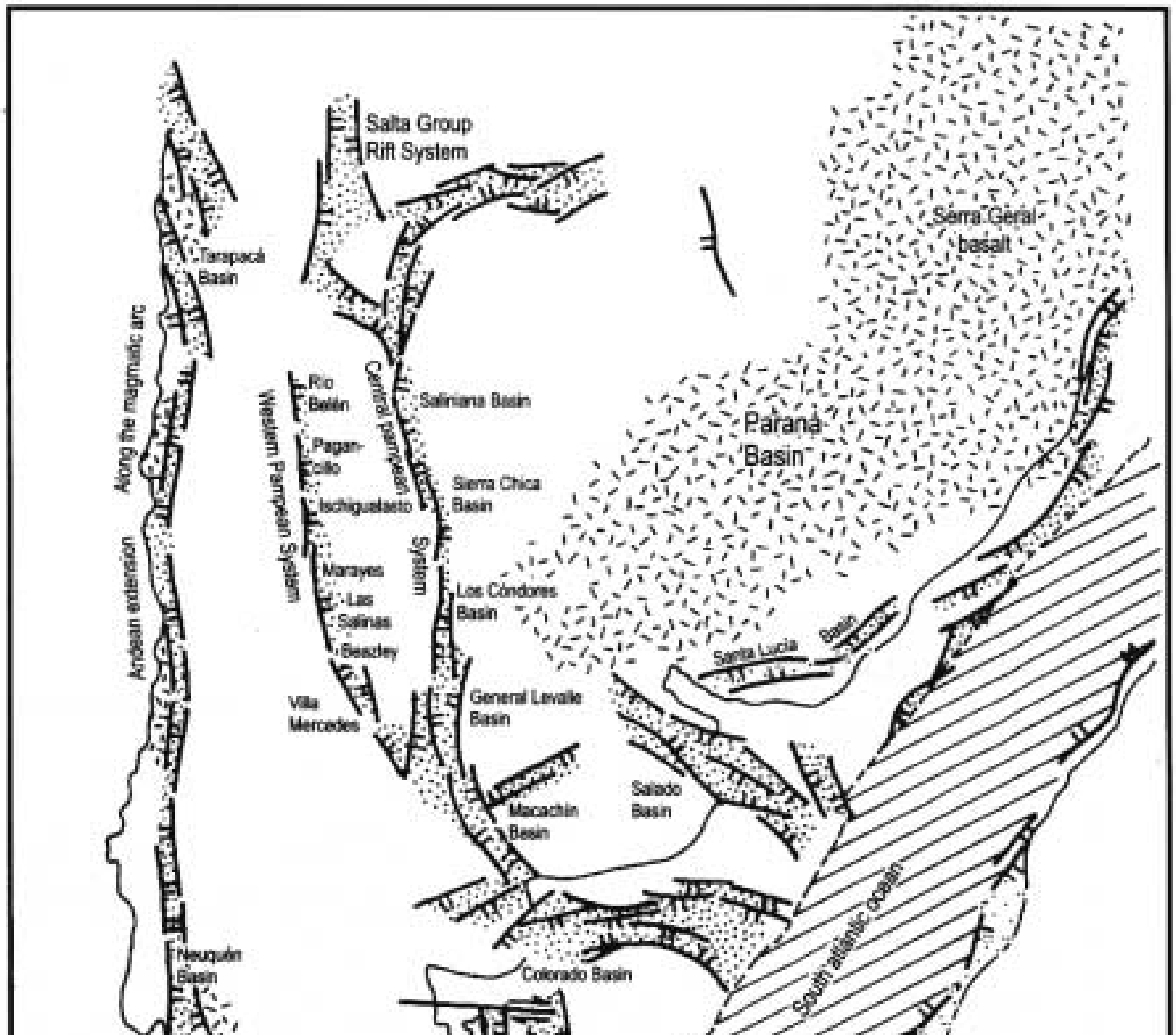


FIGURE 9: Early Cretaceous rifting in southern South America coeval with the early opening of the South Atlantic (modified after Salfity and Marquillas, 1994; Uliana and Biddle, 1988; Rossello and Mozetic, 1999).

alkali-basalt of the Pirgua Subgroup have been deposited in different sub-basins (Salfity, 1982; Salfity and Marquillas, 1994), with variable trend controlled by the Precambrian fabric of the Pampia Craton (Comínguez and Ramos, 1995). Thermal subsidence and widespread shallow marine to lacustrine sedimentation in these sub-basins took place in Campanian to Paleocene times.

Along the Pacific margin several retro-arc basins were filled with clastic and carbonate deposits (Fig. 8) such as the Neuquén Basin (Uliana and Legarreta, 1993), Río Mayo Basin (Ramos and Aguirre-Urreta, 1994), and the Austral or Magallanes Basin (Biddle *et al.*, 1986). The axis of the Austral Basin was the locus of the Rocas Verdes marginal basin (Dalziel *et al.*, 1974) that extended from Cordillera de Sarmiento at about 52°S to S of the Beagle Channel in Tierra del Fuego Island, at the southernmost Andes.

Accretion and collision along the Northern Andes

The western Northern Andes consist of several superimposed orogens as a result of the obduction of oceanic terranes (Restrepo and Toussaint, 1973), that may have been formed far from the continent. Their allochthonous nature was suspected since the early work of Barrero (1979) in western Colombia and by Henderson (1979) in Ecuador. Many authors have proposed different tectonic settings for the ophiolite sequences, tholeiitic basalt, and pelagic and hemipelagic sediments, that constitute these oceanic terranes (Feininger, 1982, 1987; McCourt *et al.*, 1984; Lebrat *et al.*, 1985; Mégard, 1987). The number and boundaries of these exotic terranes vary considerably from a maximum of 34 terranes (including the Paleozoic terranes, Etayo-Serna and Barrero, 1983) to two or three major composite terranes (Henderson, 1979; McCourt *et al.*, 1984; Restrepo and Toussaint, 1988). However, there is a general agreement in the allochthonous oceanic nature of western Colombia and Ecuador, and in the source area for these terranes in the Caribbean Plate during an early stage of its evolution (Burke, 1988; Meschede and Frisch, 1998).

The Caribbean Andes also record an important accretion of oceanic rocks derived from the Caribbean Plate that constitutes the termination of the Andes along the Atlantic Ocean.

There are some important differences among the oceanic terranes from N to S, due to the different petrological and geochemical characteristics and the timing of the accretion.

Accretion of the Caribbean Andes

The motion of the Farallon Plate changed during the Campanian from a northeasterly to a nearly eastward direction (Engebretson *et al.*, 1985). As a result, the relative movement between North America and South America became slightly convergent (Pindell and Barret, 1990), and due to the eastward displacement of the Caribbean Plate during Late Cretaceous to Eocene times, part of the oceanic floor of this plate was obducted in the Atlantic margin of northern South America.

The Caribbean Orogeny was related to the accretion of these oceanic rocks, emplaced orthogonally to the Mérida Andes during the latest Cretaceous, and continuing into the early Eocene. The late Eocene molasses were deposited in an E-W trending foreland basin that was reactivated during the Oligocene, developing a complex system of nappes, folded nappes, and deformed syn-orogenic deposits (Stephan, 1982).

The oblique subduction along the Pacific margin, increased the northward displacement of basement blocks such as the Santa Marta and the Guajira blocks, producing a complex system of strike-slip faults in the late Cenozoic (Fig. 10).

The interaction of the N-S trending Mérida Andes and the orthogonal trend of the nappes produced the stress partitioning manifest by the E-W right-lateral displacements of the Oca and El Pilar faults, and the left-lateral offsets of the Bucaramanga and Santa Marta faults.

Most of these faults, and mainly the Boconó Fault, display important neotectonic activity (Schubert and Vivas, 1993).

Accretion of western Colombia

Most authors have recognized two major accretionary episodes in the Cretaceous and another one during the Neogene (Restrepo and Toussaint, 1988; Mégard, 1987) that are recorded in the ophiolite sequences and in blue-schist exposed on the western side of the Central Cordillera, in the Western Cordillera and in the Serranía de Baudó (Fig. 11).

The first tectonic unit corresponds to several dismembered ophiolite complexes that are emplaced with an eastern vergence on the western slope of Cordillera Central (Bourgois *et al.*, 1985). There are exposed in Jambaló (Fig. 11), where these authors described high-pressure metamorphic rocks with a probable Paleozoic protolith, but with a whole-rock age of 125±15 Ma (Orrego *et al.*, 1980), which is interpreted as the Early Cretaceous age of obduction. Several other bodies emplaced roughly along the Romeral Fault have similar ages (Mégard, 1987).

The second tectonic unit corresponds to the Dagua Terrane exposed in the Western Cordillera, and in some tectonic klippes of the Central Cordillera. The thrust sheets are 1.5 to 5 km thick and are folded by later deformation and also affected by subvertical strike-slip faults (Bourgois *et al.*, 1987). These rocks known as the Diabase Group are oceanic low-K tholeiite, with ages based on paleontological and geochronological data, ranging from Albian to Maastrichtian (Barrero, 1979). Most authors favour an oceanic flood basalt origin for the Diabase Group of the Dagua Terrane (Millward *et al.*, 1984, Lapierre *et al.*, 1999). The cross-cutting relationships between the Antioquía Batholith (59-57 Ma) and the ophiolite in the Yarumal Complex constrain the emplacement of this second tectonic unit between 80 and 60 Ma (Restrepo and Toussaint, 1988; Toussaint and Restrepo, 1994).

The third tectonic unit consists of the Choco Terrane, that includes the oceanic terranes W of the Atrato Suture (Fig. 11), and mainly exposed in the Serranía de Baudó (Dengo, 1983). This terrane was formed by island-arc tholeiitic assemblages, emplaced in an oceanic plateau at approximately 78 - 72 Ma (Kerr *et al.*, 1997a). This oceanic

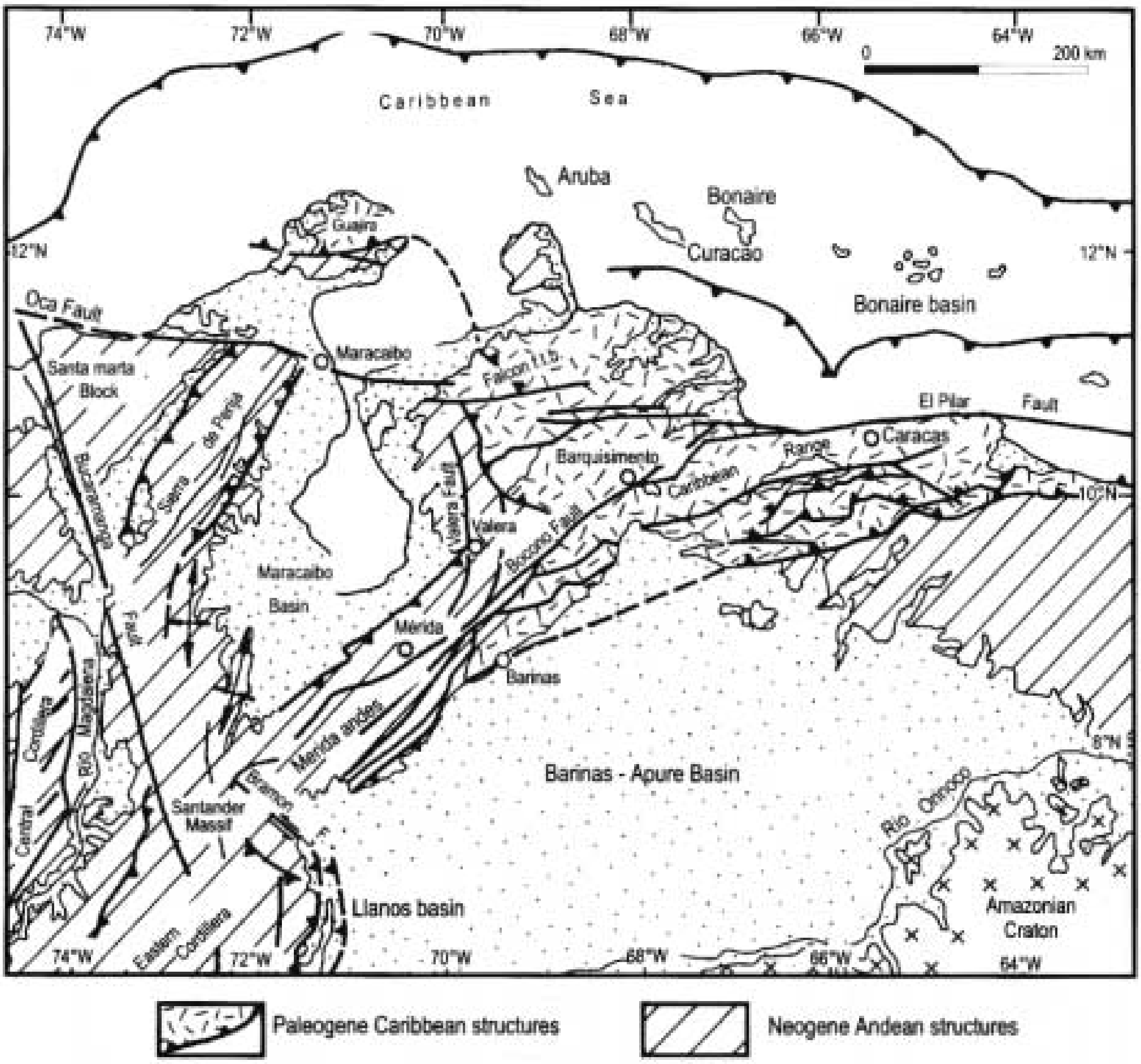


FIGURE 10: Present tectonic setting of the Caribbean Andes. Note the right-lateral offset produced in the nappes front by the Boconó strike-slip fault (modified after Colletta et al., 1997).

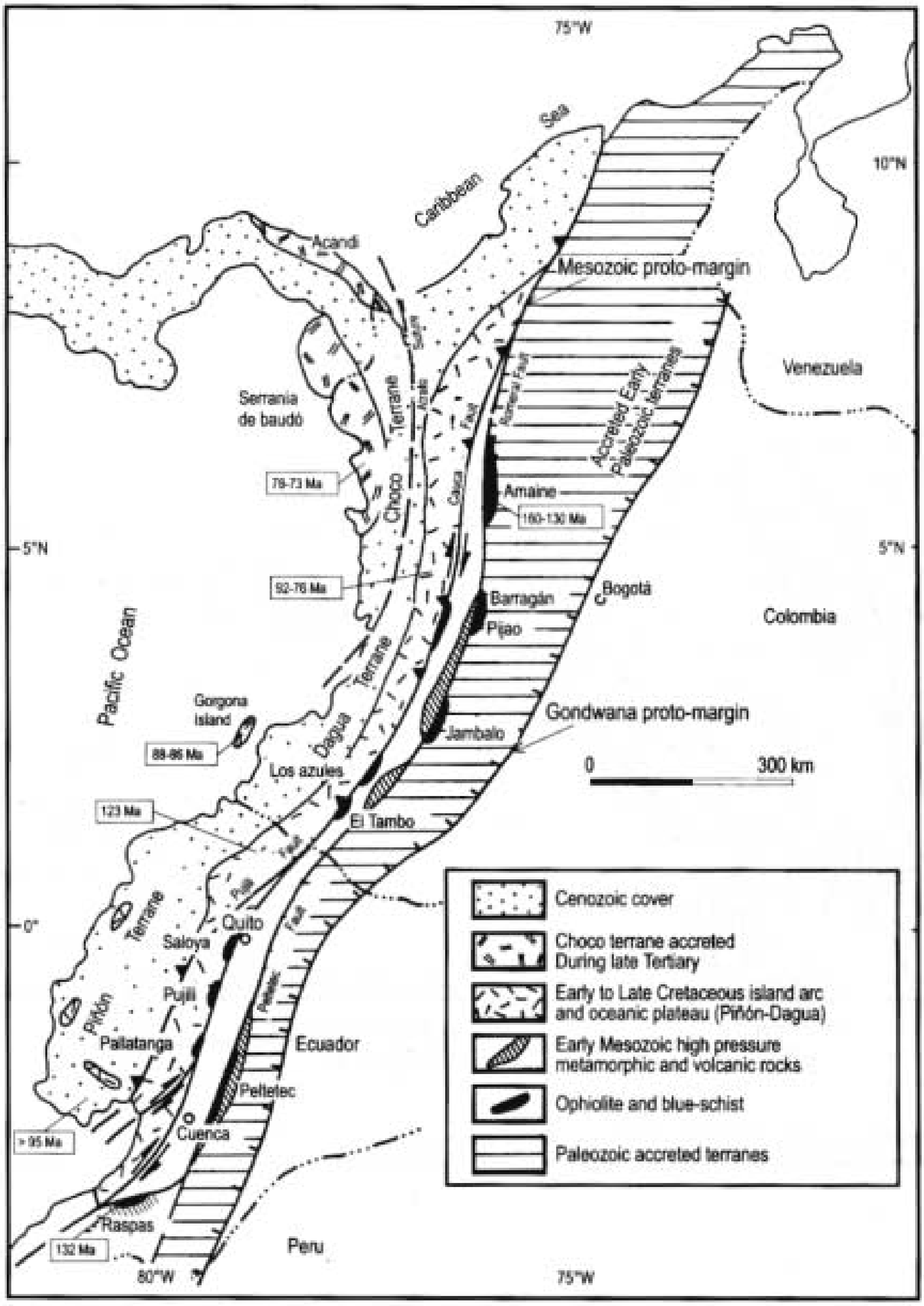


FIGURE 11: Accreted oceanic terranes of the Northern Andes with indication of main ophiolites and high pressure metamorphic rocks exposed along the Cauca Valley of Colombia and the Interandean Depression of Ecuador (modified after McCourt et al.*, 1984; Millward et al., 1984; Van Thornout* et al.*, 1992). Radiometric ages correspond to the oceanic basement (modified after Reynaud* et al.*, 1999).*

basement, covered by pelagic to hemipelagic sequences of Late Cretaceous to Miocene age, was accreted to continental South America during a short episode of obduction between 12.9 to 11.8 Ma (Duque-Caro, 1990). According to this author the planktonic assemblages found in the Paleocene cover of this terrane have Central American affinities, which corroborate a source in the Caribbean Plate of this terrane (Meschede and Frisch, 1998).

Based on the existing data it is interpreted that an island arc terrane, active during the Jurassic to Early Cretaceous time, was accreted to South America, as indicated by the calc-alkaline magmatic arcs developed along the Central Cordillera (Colletta *et al.*, 1990). Remnants of this oceanic terrane are now preserved on the western slope of Central Cordillera W of the Early Mesozoic margin, and are bounded by the Romeral Suture (Fig. 12).

A most prominent and best-preserved obduction corresponds to the Calima Orogeny (Barrero, 1979) that produced an important Alpine deformation during Late Cretaceous-Paleocene times. This deformation was transmitted through the Central Cordillera to the Magdalena Valley and coincides with the tectonic inversion of the western flank of Eastern Cordillera (Etayo-Serna, 1994; Restrepo- Pace, 1999). This deformation abruptly terminated with marine deposition in the retro-arc basins (Cooper *et al.*, 1995).

The Paleogene arc at this time was situated on the western slope of Western Cordillera (Toussaint and Restrepo, 1982). After a Paleogene period of oblique subduction the third obduction took place in the middle Miocene, with the emplacement of the Choco Terrane, that produced reactivation and folding of the older structures and intense deformation in the foothills of the Eastern Cordillera. The basin axes steadily migrated from the Maastrichtian in the Upper Magdalena Valley on the western margin of Eastern Cordillera to the Subandean foothills in the early to late Oligocene (Villamil, 1999). In middle Miocene times, deformation extended to the Llanos Basin developing a complex structure where thin-skinned deformation was followed by thick-skinned basement faults, as seen in the interaction of the Cusiana and Yopal fault systems (Cooper *et al.*, 1995).

After the emplacement of these oceanic terranes, Andean-type subduction in the present trench began in the late Miocene. The magmatic arc at first situated in the Western Cordillera migrated during Pliocene times to the Central Cordillera (Toussaint and Restrepo, 1982). However, probably due to the shallowing of the Wadati-Benioff Zone to the N, active volcanism in the late Cenozoic arc ends in Cerro Bravo, at 5°N (Méndez-Fajuri, 1989), developing the Bucaramanga flat subduction segment (Pennington, 1981).

Accretion of western Ecuador

Western Ecuador also recorded a complex history of accretion, preserved in the Western Cordillera and the adjacent coastal plains, W of the Pujili Fault, as well as in the Eastern Cordillera (or Cordillera Real) E of the Peltetec Fault (Fig. 11).

The most complex structures are in the Cordillera Real because Mesozoic and Paleogene deformation as well as late Miocene strike-slip obliterated the original relationships. Aspden and Litherland (1992) have reconstructed the tectonic settings of a series of metamorphic units and undeformed volcanic rocks of Triassic and Jurassic ages. These settings encompass ocean floor and island-arc assemblages for the Alao Terrane, and Jurassic arc rocks for the Misahualli andesite and dacite, as well as calc-alkaline granitoid rocks for the Abitagua Stock. A simplified tectonic evolution is shown in Figure 13.

The Peltetec Fault on the western flank of Cordillera Real is associated with a melange with island-arc signature, where the sheared metagabbro, metabasalt and foliated serpentinite has been interpreted as a subduction complex of Jurassic - Early Cretaceous age (Litherland *et al.*, 1994). These rocks are correlated with the ophiolite sequences exposed along the Romeral Suture in Colombia, as well as the Raspas metamorphic complex (132 Ma), and associated ophiolitic rocks exposed on the northern flank of the Amotape-Tahuin terranes (Feininger, 1982; Reynaud *et al.*, 1999).

The Western Cordillera oceanic rocks of the Piñón Terrane are interpreted as having developed in an oceanic plateau of 123 Ma (Sm/Nd age, Lapierre *et al.*, 1999). An island arc system was developed in this Early Cretaceous oceanic basement, known as the Macuchi Arc of Late Cretaceous - Paleocene age in northern Ecuador (Henderson, 1979; Van Thournout *et al.*, 1992). The Piñón Terrane had in the Manabi area along the coast in southern Ecuador, the San Lorenzo Arc (77-60 Ma), and to the E in the Guayaquil area the Cayo Arc (92-80 Ma), two oceanic island-arc systems of Late Cretaceous age (Jaillard *et al.*, 1995). These authors favour the development of a marginal basin between these two arcs, being the Cayo volcanics a remnant arc during latest Cretaceous.

The collision of the Piñón Terrane against the continental margin was diachronic. It started in the southern part by Maastrichtian times, and it was completely amalgamated during early Eocene time (Reynaud *et al.*, 1999). The andesite, dacite and breccia of the Tandapi Arc indicate a continental calc-alkaline setting (Cosma *et al.*, 1998). The Baudó oceanic rocks of the Choco Terrane are not present in the coastal plains of Ecuador.

East of the accreted terranes a magmatic arc developed in the Cordillera Real during Jurassic and Cretaceous times. The Celica Arc was developed during Late Cretaceous in central and southern Ecuador on the continental margin (Lebrat *et al.*, 1985). The metamorphic facies of the Celica Volcanics indicate its development in a highly attenuated crust during an extensional regime (Aguirre, 1992). Paleocene compression generated a thickened crust and the marginal basin inversion (Jaillard *et al.*, 1995), and since that time, the development of normal Andean-type volcanism along the margin.

During the Eocene, the arc developed on the accreted oceanic terranes, but after important late Eocene deformation (Jaillard *et al.*, 1995), the arc migrated to the Interandean Valley in Miocene times. The opening of the Guayaquil Gulf during Miocene times due to important strike-slip in the Pujili-Cauca and Peltetec-Romeral fault systems, generated a series of complex pull-apart basins in the coastal areas, and intermontane basins in the Andean areas (Jaillard *et al.*, 1999). Important deformation and basin

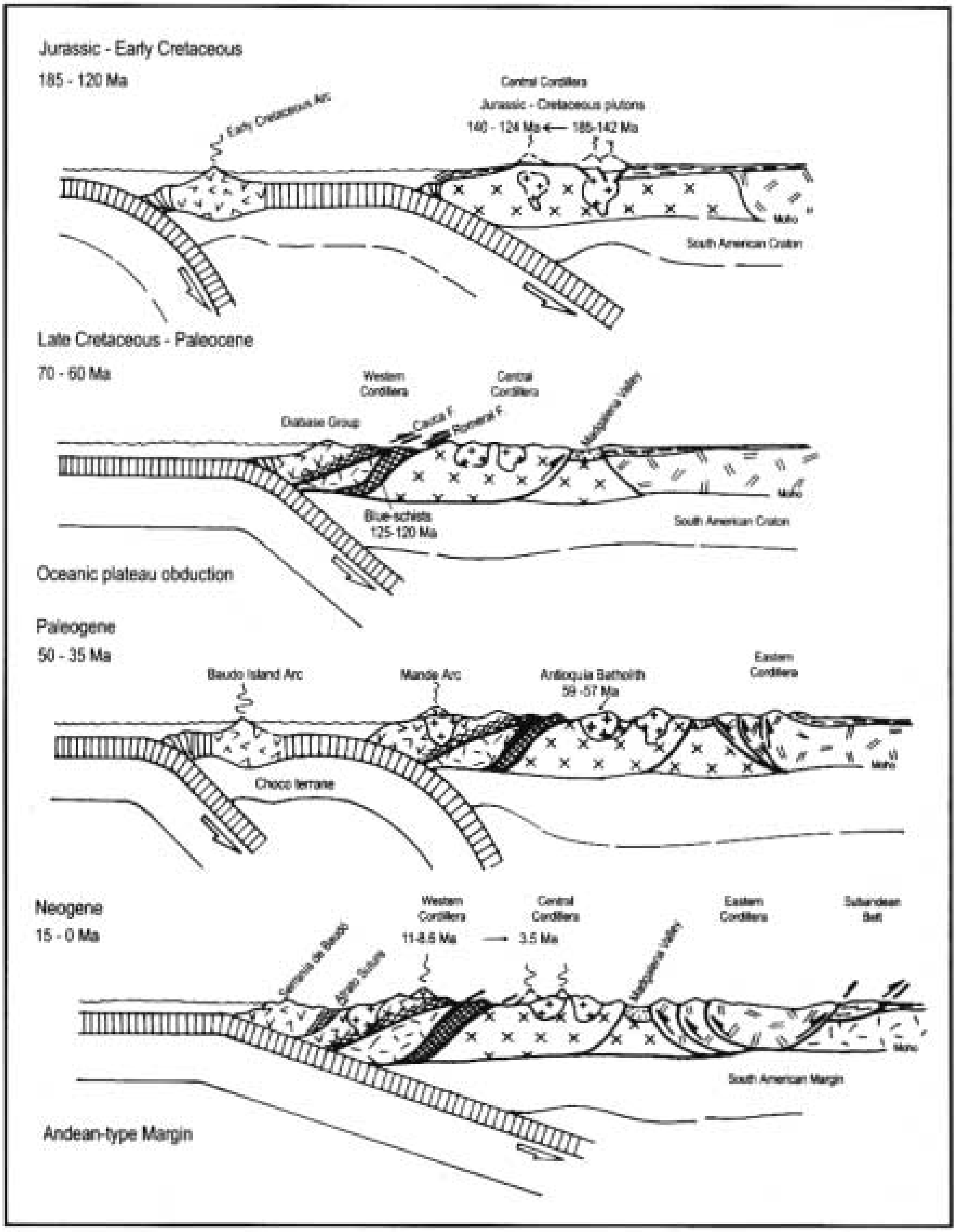


FIGURE 12: Schematic evolution of western Colombia showing the accretion of different terranes. Ophiolites related to the Romeral Fault represent the suture of the oceanic terranes with the Early Mesozoic continental margin. The obduction of an oceanic plateau emplaced the Diabase Group of the Dagua Terrane (modified after Barrero, 1979; Aspden and McCourt, 1986; Bourgois et al., 1987; Colletta et al., 1990; Kerr et al., 1997a,b).

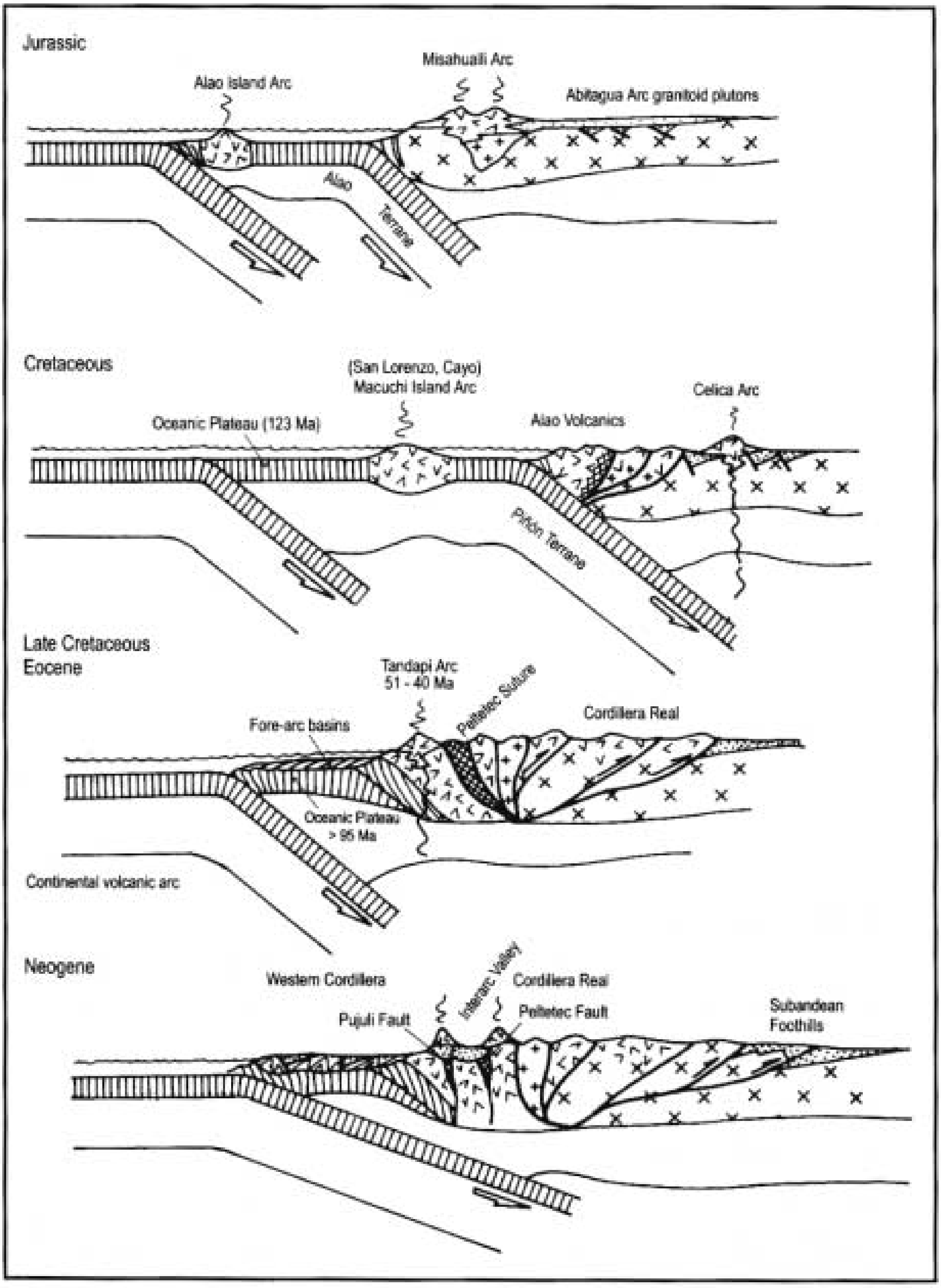


FIGURE 13: Generalized tectonic evolution of the accreted terranes of western Ecuador (modified after Aspden and Litherland, 1992; Van Thournout et al., 1992; Aguirre, 1992; Cosma et al., 1998; Reynaud et al., 1999).

inversion took place in middle to late Miocene times related to the onset of the Carnegie Aseismic Ridge (Daly, 1989). As a consequence of this, an uplift-rate of 0.7 mm/y occurred in the last 9 Ma, accounting for more than 6 km of uplift (Steinmann *et al.*, 1999).

Intra-arc and arc magmatism

The Andes between of the Guayaquil (3°S) and Penas (46°30'S) gulfs have a long history of Andean-type subduction, without accretion of oceanic terranes. This segment of the Andes known as the Central Andes after Gansser (1973) has a series of varying tectonic settings through time, prior to the present Andean-type subduction. The Early Mesozoic was a time of extension along most of the Andean margin. The first stages were related to active rifting during the Pangea break-up, and soon after that, in the latest Triassic or Early Jurassic times, to a peculiar type of subduction that developed a poorly evolved magmatism with intra and retro-arc extension.

This Early Mesozoic subduction was related to a negative trench rollback velocity as described by Uyeda (1983) and Daly (1989), as a consequence of the relative motions between the Farallon and South America plates. The change in true polar wandering path of South America at about 115 Ma, as calculated from the hot-spot reference frame (Somoza, 1995), modified the trench rollback to a positive velocity. As a result of that the South America Plate overrode the previous trench line beginning an important period of compression (Ramos, 1989a).

The timing and intensity of the extension was mainly controlled by the obliqueness of the convergence vector with the continental margin trend as inferred in northern Chile by Scheuber *et al.* (1994).

Northern Central Andes

This changing Early Mesozoic scenario has been recognized in central Peru (Fig. 14), where a marginal basin formed along the continental margin during Early Cretaceous subduction (Atherton *et al.*, 1983). This marginal basin, here interpreted as an intra-arc basin, was responsible of the Copara-Casma volcanism developed during Aptian-Albian times (Soler and Bonhomme, 1990). The basin was filled by several kilometres of pillow lava, sheet lava, hyaloclastite sediments, tuff and minor chert, siliceous and calcareous ooze. There are tholeiitic basalt flows and andesite along the axis with more acid high-K rocks toward the E (Atherton and Webb, 1989). Geochemical affinities indicate a destructive margin, with some rocks showing some MORB-type tendencies. This is in agreement with basin deepening to the W, and a general trend of deeper facies to the N, where the basin continues into the Celica Arc Basin (Jaillard *et al.*, 1999). The Casma Group Basin has been considered an aborted ensialic basin developed on highly attenuated continental crust (Aguirre and Offler, 1985) compatible with an extensional intra-arc setting (*sensu* Dickinson, 1974).

Localized Middle Albian Mochica compressional deformation was coeval with early batholith emplacement (± 105 Ma; Soler and Bonhomme, 1990). Softening of the lower crust and weakening of the lithosphere along the continental margin during the extension of the Casma Group volcanism favoured the compressional deformation. This phase was also associated with an increase in convergence rates, and was responsible of basin closure, compressional deformation, tectonic inversion and emplacement of the Coastal Batholith from Late Albian to Paleocene times (102 to 59 Ma). Several episodes of emplacement have been recognized, as well as a variety of facies from typical I-type medium-K to high-K granitoid rocks, described in detail by Pitcher *et al.* (1985). Oblique convergence was dominant during Coastal Batholith emplacement, with a low orthogonal convergence rate (Soler and Bonhomme, 1990). Compressional deformation of the Peruvian phase propagated eastwards during the Late Cretaceous, changing the volcanism from generally subaqueous to subaerial. The Late Cretaceous deformation was also well developed in the Western Cordillera of southern Peru, where the allochthonous thrust sheet of the Arequipa-Paracas Arc overrode the Early Mesozoic platform facies (Vicente, 1990).

Subsequent middle to late Eocene (47-32 Ma) compression, produced the fold and thrust belt of the Western Cordillera, E of the Cordillera Blanca Lineament (Benavides-Cáceres, 1999). A series of Paleogene pulses between the middle Eocene and the early Oligocene are known as the Incaic Phase, which produced the main orogenic uplift of the Peruvian Andes (Vicente *et al.*, 1979). This coincides with a period of rapid orthogonal convergence rates, and expansion of the magmatism as shown by Pilger (1984) and Soler and Bonhomme (1990). The expansion and migration of the magmatism was related to a shallowing of the geometry of the Wadati-Benioff Zone at that time (Fig. 15).

Diachronic Miocene compression, generally attributed to some of the Quechua phases, uplifted and deformed the Eastern Cordillera and the Subandean foothills. The emplacement of the Cordillera Blanca Batholith (13-3 Ma) in the highest sector of Western Cordillera was controlled by an important crustal discontinuity (Atherton and Sanderson, 1987). The Peruvian flat-slab segment was developed in Pliocene times, and a magmatic lull extended from the Ecuador boundary to the latitude of Arequipa, with the only exception of some minor Pliocene peralkaline stocks in the Subandean Basin (Sébrier and Soler, 1991).

Southern Peru was characterized by some sort of delamination as proposed by Sandeman *et al.* (1995), after the Paleogene shallowing of the subduction zone, associated with the Incaic deformation. A thermal effect with no associated magmatism in Paleogene times was recognized along 450 km of the Eastern Cordillera parallel to the Altiplano (Farrar *et al.*, 1988). This area, known as the Zongo-San Gabán Zone, has produced a thermal event detected by Ar^{39}/Ar^{40} at 38 Ma, associated with uplift, erosion and southwestward vergent thrusting on the western flank of the Eastern Cordillera. This deformation was prior to the northeastward vergent thrusting along the eastern foothills. Sandeman *et al.* (1995) favoured the formation of a slab-window linked to the detachment of the subducting plate. An alternative model could be the transition from flat to normal subduction that resulted in interaction of hot

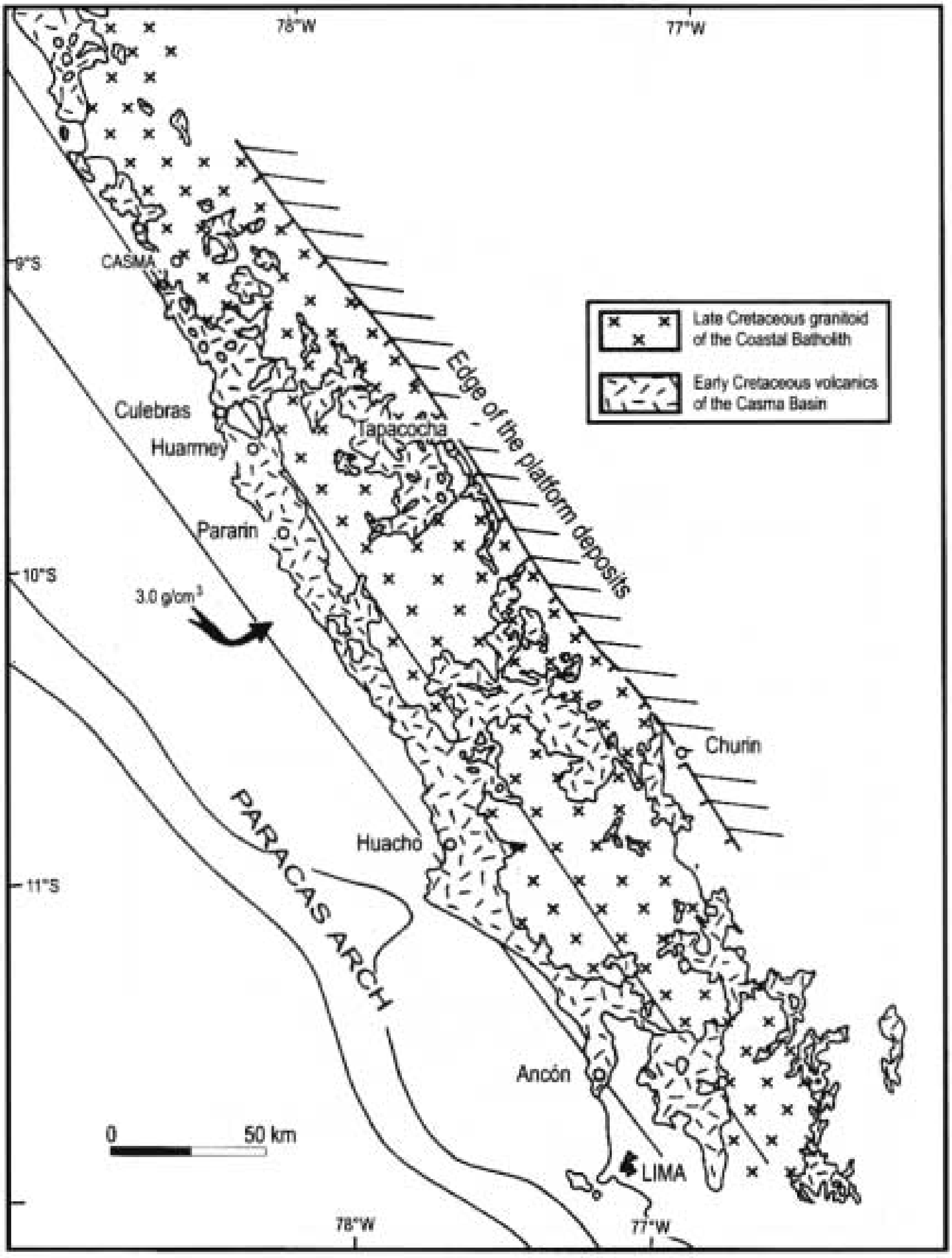


FIGURE 14: Surface exposures of the Casma Group volcanics, and offshore possible extension E of the 3.0 g/cm³. Basin is bounded by the Paraca Arch, possible offshore extension of the Arequipa Massif. The Coastal Batholith was emplaced along the Casma Group Basin (modified after Atherton and Webb, 1989; Soler and Bonhomme, 1990; Benavides-Cáceres, 1999).

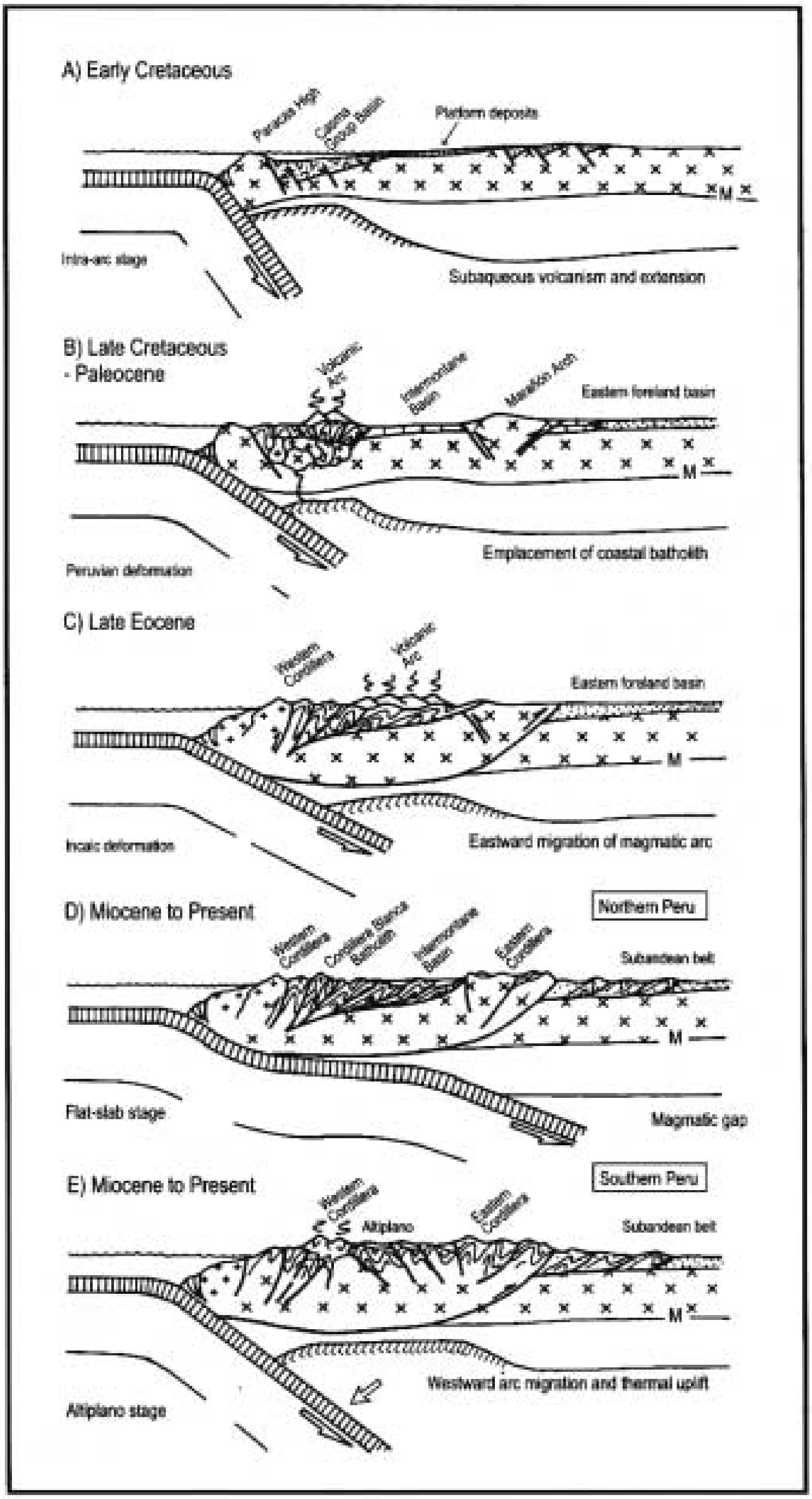


FIGURE 15: a) to c) Tectonic evolution of the Peruvian margin from an intra-arc setting and generalized rifting in the Early Cretaceous to a compressive regime in the Cenozoic; d) Late Cenozoic in Northern and Central Peru; e) Late Cenozoic in Southern Peru (modified after Soler and Bonhomme, 1990; Sébrier and Soler, 1991; Sandeman et al., 1995).

asthenosphere and hydrated mantle. This would cause mantle melts and hot fluids that may explain the thermal effect as being proposed farther S by Kay *et al.* (1999).

Late Cenozoic in the southern Peru was characterized by important arc magmatism along the Western Cordillera, with some shoshonitic volcanic rocks occurring in the Eastern Cordillera. The northern flank of the Altiplano was affected by important normal faulting in late Cenozoic times. This normal faulting was explained by the high topography produced after uplift of the Andean Cordillera and the Altiplano, and the reorganization of the stress that resulted in N-S rending extension (Sebriér *et al.*, 1985).

Central Andes

Early Mesozoic evolution of this segment differs from the previous segment. Although there was a generalized extension, during Jurassic and Early Cretaceous times between 21°S and 27°S, a deep retro-arc basin developed behind the arc (Mpodozis and Ramos, 1990). Volcanic activity was concentrated in the La Negra Arc, a poorly evolved tholeiitic series of basalt flows, basandesite, and mafic dykes, developed along the continental margin during an extensional regime. The western part of La Negra Arc was tectonically eroded (Rutland, 1971), due to the high relief horst-and-graben structure of the outer slope of the oceanic subducting plate at these latitudes (Von Huene *et al.*, 1999).

The highly oblique convergence during Early Mesozoic times (Scheuber *et al.*, 1994), was also responsible for the development of the Atacama strike-slip system (Fig. 16). This trench-parallel fault system, which affected the Coastal Cordillera, controlled the ascent and emplacement of the arc granitoid of the Coastal Batholith during the Jurassic and earliest Cretaceous (Naranjo *et al.*, 1984). The negative rollback velocity of the trench expanded the previous Triassic normal faults, until about 132-125 Ma, when a left-lateral strike-slip system generated important transtension and enhanced the emplacement of granitoid plutons (Grocott *et al.*, 1994).

The transtensional regime ended at about 106 Ma, when the Coastal Cordillera Arc was abandoned and the arc migrated toward the Central Valley (Mpodozis and Ramos, 1990). This 50 km shifting of the magmatic arc coincides with major compressional reactivation of the strike-slip system, where a series of compressional duplexes deformed the Coastal Cordillera granitoid rocks. Rotation associated with this strike-slip displacements was confirmed by paleomagnetic studies (Taylor *et al.*, 1998).

Late Cretaceous compressional deformation, approximately coeval with other Cretaceous phases described in the Peruvian segment, thrust and uplifted the proto-Cordillera de Domeyko (Mpodozis and Ramos, 1990), and generated the Purilactis-Salar de Atacama foreland basin (Muñoz *et al.*, 1997).

The early Paleogene was characterized by an intense explosive volcanism (Fig. 17), with huge calderas and associated intrusives, developed in an extensional regime (Cornejo *et al.*, 1994). During late Eocene and early Oligocene times a lull in arc activity, was associated with the Domeyko Fault System, another strike-slip system developed parallel to the Atacama, but further to the E in the present Cordillera de Domeyko (Tomlinson and Blanco, 1997). Magmatic activity reduced to some acid porphyries was emplaced along the Domeyko Fault System, where giant porphyry copper ore deposits are found (Zentili and Maksaev, 1995).

Deformation and magmatism shifted to the Altiplano (N of 22°S) in Bolivia and to the Puna (S of 22°S) in Argentina, initially linked to the tectonic erosion of the fore arc (Stern, 1991), and subsequently associated with shallowing of the subduction zone (Kay *et al.*, 1999).

Paleogene magmatic rocks are scarce in the Altiplano, where a few volcanic fields (*c.* 34-23 Ma) are known with a retro-arc setting (Avila Salinas, 1991). Rocks of this age are rare in the Argentine Puna, and are associated with acid porphyries such as Taca Taca, in the western sector.

Most of the Altiplano and Puna was affected by active extensional faulting until Albian times with the development of the Salta Rift, and equivalent systems in the Altiplano, as well as in the Subandean region (Sanjines-Saucedo, 1982; Salfity, 1982, 1994). Thermal subsidence in these basins persisted until early Paleogene times, although some authors interpreted part of this Paleogene subsidence as a distal response of the flexural loading produced by the Peruvian phase farther W (Sempere, 1995).

Shifting of the deformation to the Altiplano and Puna produced during late Oligocene-early Miocene (Sempere *et al.*, 1990) the tectonic inversion of the Cretaceous N-S trending grabens, and the uplift of the Eastern Cordillera. As a result, the Neogene flexural loading produced two distinctive settings, the Altiplano intermontane basins and the foreland Subandean basins. This tectonic subsidence was enhanced in the Altiplano by western vergent-thrust systems in Eastern Cordillera (Hérail *et al.*, 1993). Calc-alkaline volcanic rocks expanded between 17 to 12 Ma to the E in the Altiplano and Puna, associated with a shallowing of the subduction zone (Kay *et al.*, 1999).

After an important tectonic shortening in late Miocene, where deformation produced important uplift of the Altiplano-Puna and Eastern Cordillera, deformation shifted to the Subandean System, where thin-skinned thrust belt starts developing (Baby *et al.*, 1995).

By the late Miocene to Pliocene, voluminous ignimbrite sheets, with ages ranging from 11 to 3 Ma, were erupted from huge calderas. Centers with ages from 11 to 6.5 Ma are found across the Altiplano-Puna and into the Eastern Cordillera, whereas younger centers are restricted to Western Cordillera and the western flank of the Altiplano (Kay *et al.*, 1999). This change in the character of the volcanism, and the migration toward the trench was interpreted as evidence of steepening of the subduction zone, that concentrated during late Cenozoic times the magmatic activity in Western Cordillera (Coira *et al.*, 1993). The different geological processes associated with the steepening of a subduction zone have been analyzed by James and Sacks (1999), who noted the important influx of asthenosphere material from depth into the growing mantle wedge. As a result of this there occurred widespread melting beneath the older hydrated arc.

Tectonic evolution of the region continued through foreland propagating thrusts in the foothills of the Subandean System and adjacent Chaco plains, and out-of-sequence thrust reactivation in the Altiplano-Puna and Eastern Cordillera.

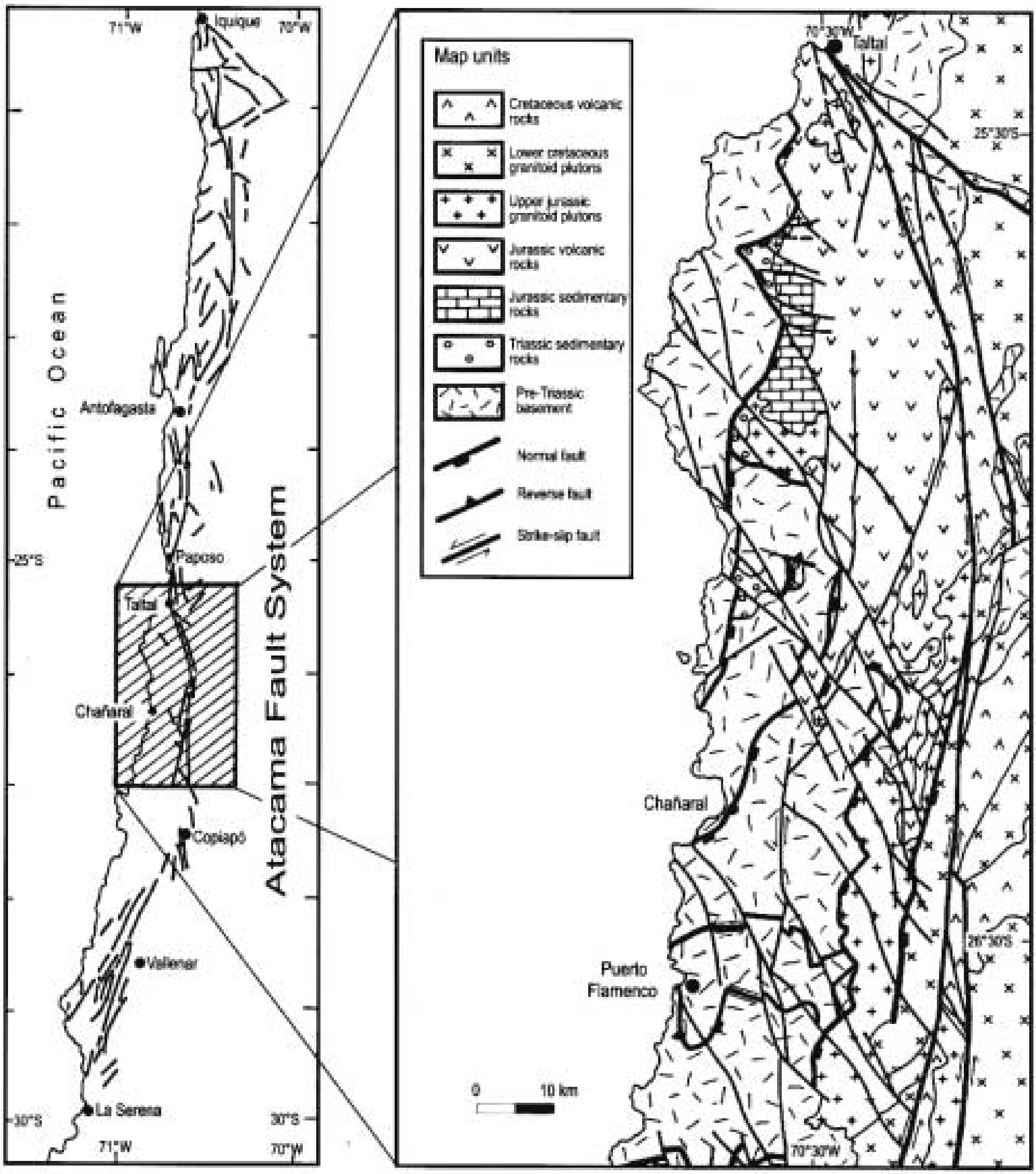


FIGURE 16: a) The Atacama strike-slip fault system. b) Detailed structural scheme between Taltal and Puerto Flamenco with location of early normal faults, and subsequent strike-slip faults with the different granitoid plutons emplaced during left lateral transtension (modified after Naranjo et al., 1984; Taylor et al., 1998).

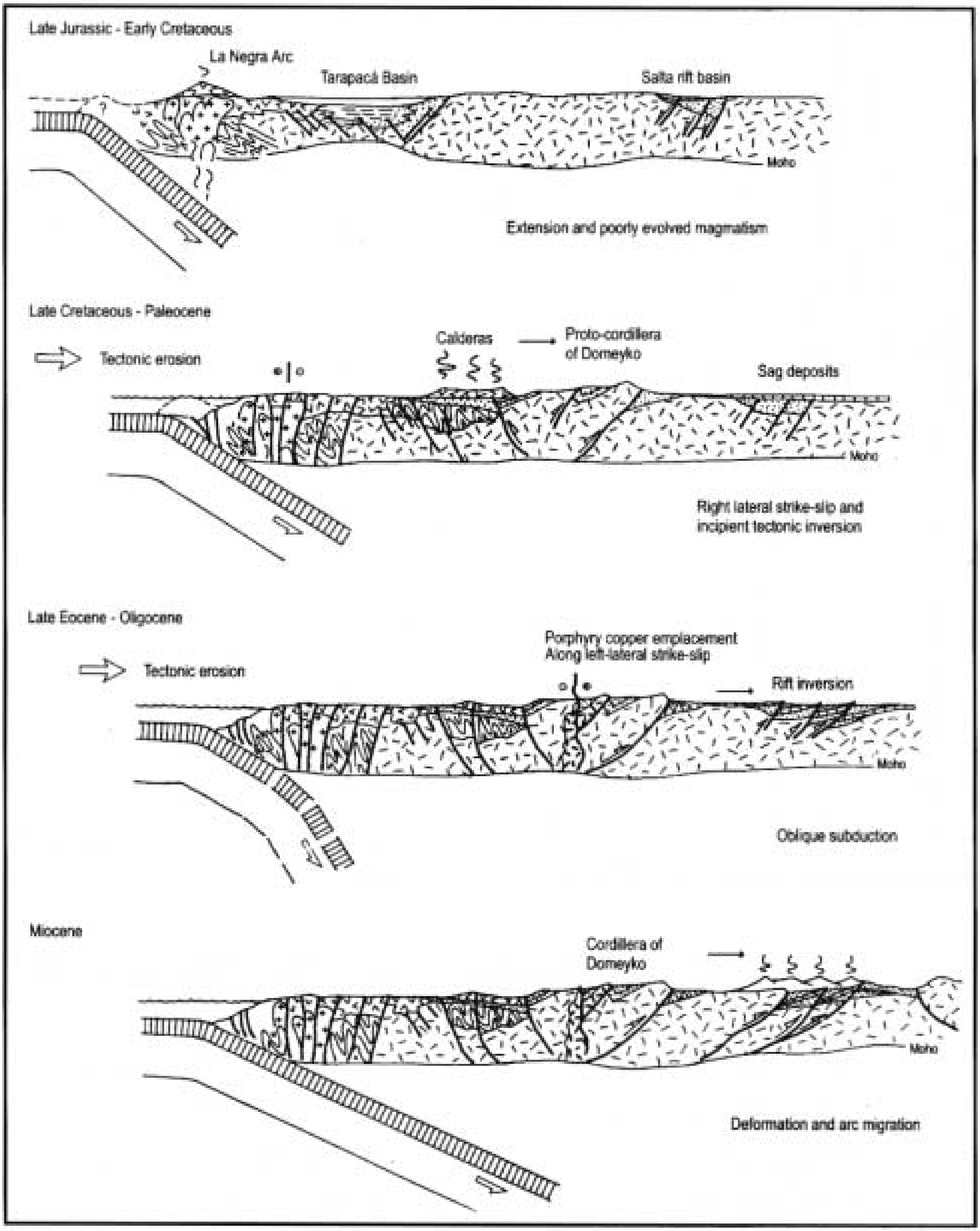


FIGURE 17: Tectonic evolution of the Central Andes since Mesozoic times (modified after Mpodozis and Ramos, 1990).

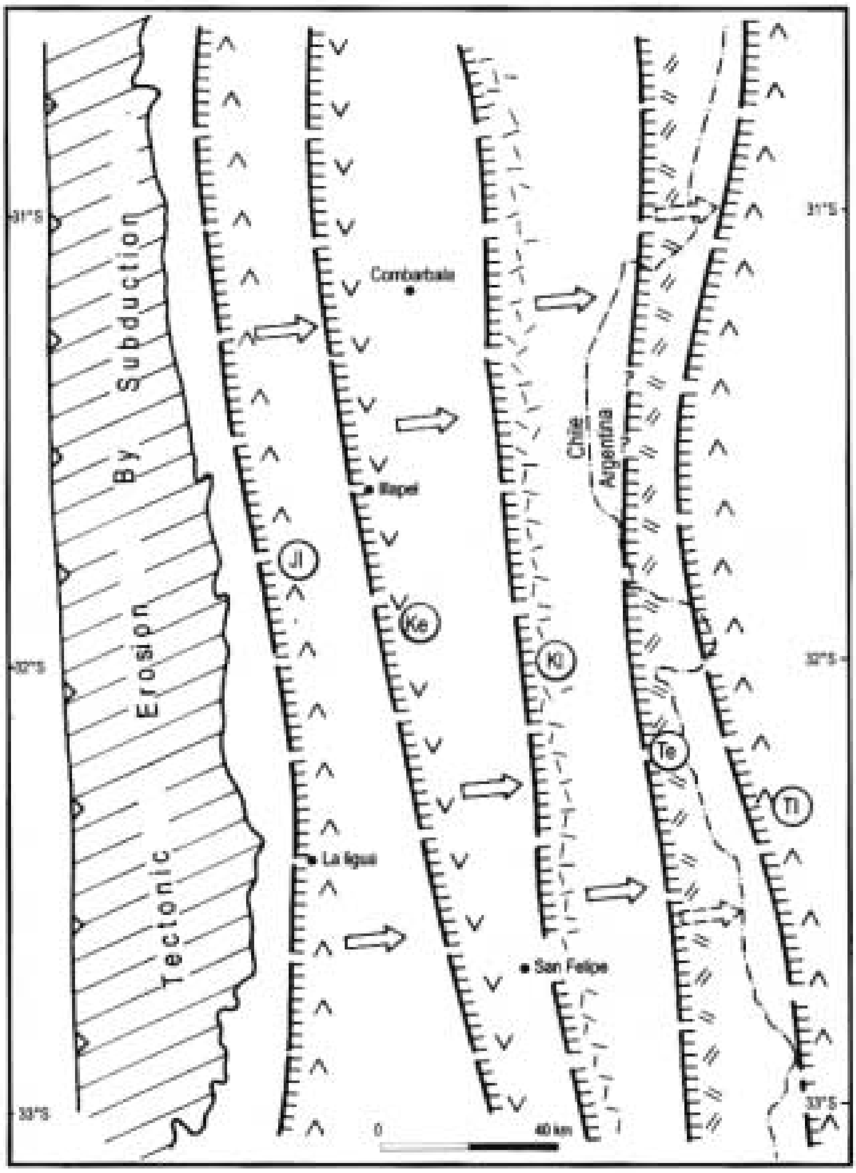


FIGURE 18: Eastward migration of the volcanic front related to tectonic erosion of the continental margin up to Paleogene times (modified after Ramos, 1988b). Miocene volcanic front migration and magmatic arc expansion were controlled by changes in the Wadati-Benioff geometry (modified after Jordan et al., 1983).

As a result, one of the most complete section of the Andes is observed in the Central Andes of Bolivia, with an active orogenic front propagated as far as 800 km from the trench.

Southern Central Andes

The central region of Chile and Argentina underwent an important extension during early Mesozoic times. This generalized extension was interpreted as the result of continental spreading of an ensialic aborted marginal basin where several kilometres of Middle Cretaceous volcanic rocks bearing an important burial metamorphism were deposited (Aguirre *et al.*, 1989). This tectonic scenario has also been explained as an intra-arc setting by Charrier (1984) and Ramos (1985), where an inner and outer arc with different petrographic characteristics were developed.

The tectonic setting of this central Chile basin is similar to the Casma Group basin described in the Peruvian segment, although the sedimentary facies associated with the intra-arc basin do not record deep or basinal facies as in the Peruvian Basin (Atherton and Webb, 1989). Most of the carbonate platform and associated clastic sediments indicate shallow water deposition, and interfingering with subaerial volcanism (Charrier, 1984). These intra-arc basins were coeval with the development of retro-arc basins in the eastern side of the orogen. Several cycles of carbonate and clastic sediments were deposited, controlled by tectono-eustatic sea-level changes (Uliana and Legarreta, 1993; Legarreta and Uliana, 1996).

Extension was driven by low-angle normal faults associated with subhorizontal detachments in the Copiapó region of central Chile (27°S, Mpodozis and Allmendinger, 1993). The retro-arc basins on the eastern side of the cordillera show older extensional structures that are characterized by high-angle faults.

The plate-kinematic changes observed along the entire South American Pacific margin are also noticeable in central Chile. By the end of the Early Cretaceous, a major plate reorganization took place, and as a response the extensional regime of the marine intra-arc and retro-arc basins ended (Mpodozis and Ramos, 1990). The new stress field was responsible of the sea withdrawal, and the inception of continental retro-arc basins. The volcanic front migrated to the E (Figs. 18 and 19) and a series of volcanic and

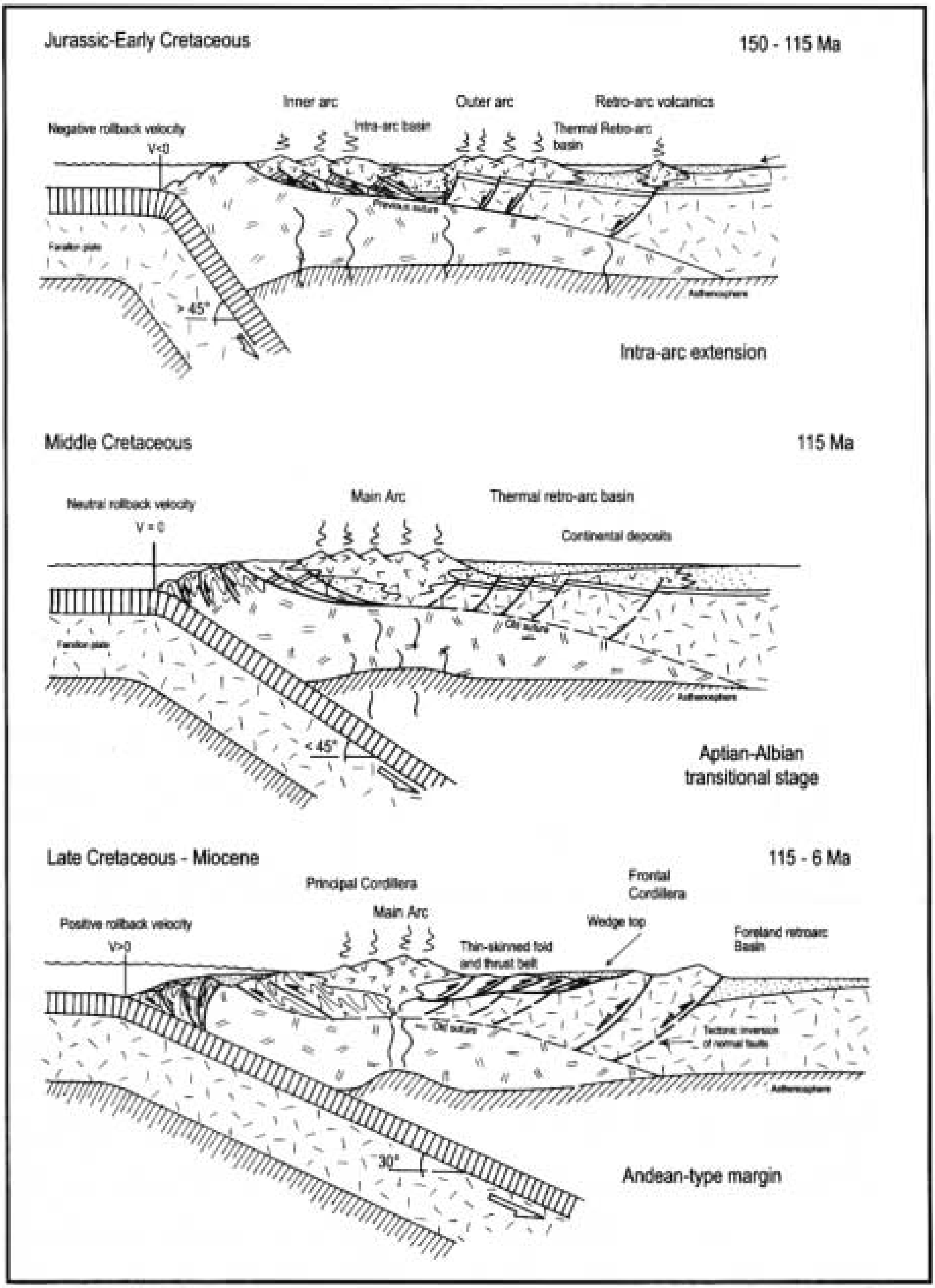


FIGURE 19: Different tectonic regimes in the southern Central Andes, showing the change between an extensional and compressional tectonic regime at about 115 Ma (modified after Ramos, 1988b; Mpodozis and Ramos, 1990; Somoza, 1995).

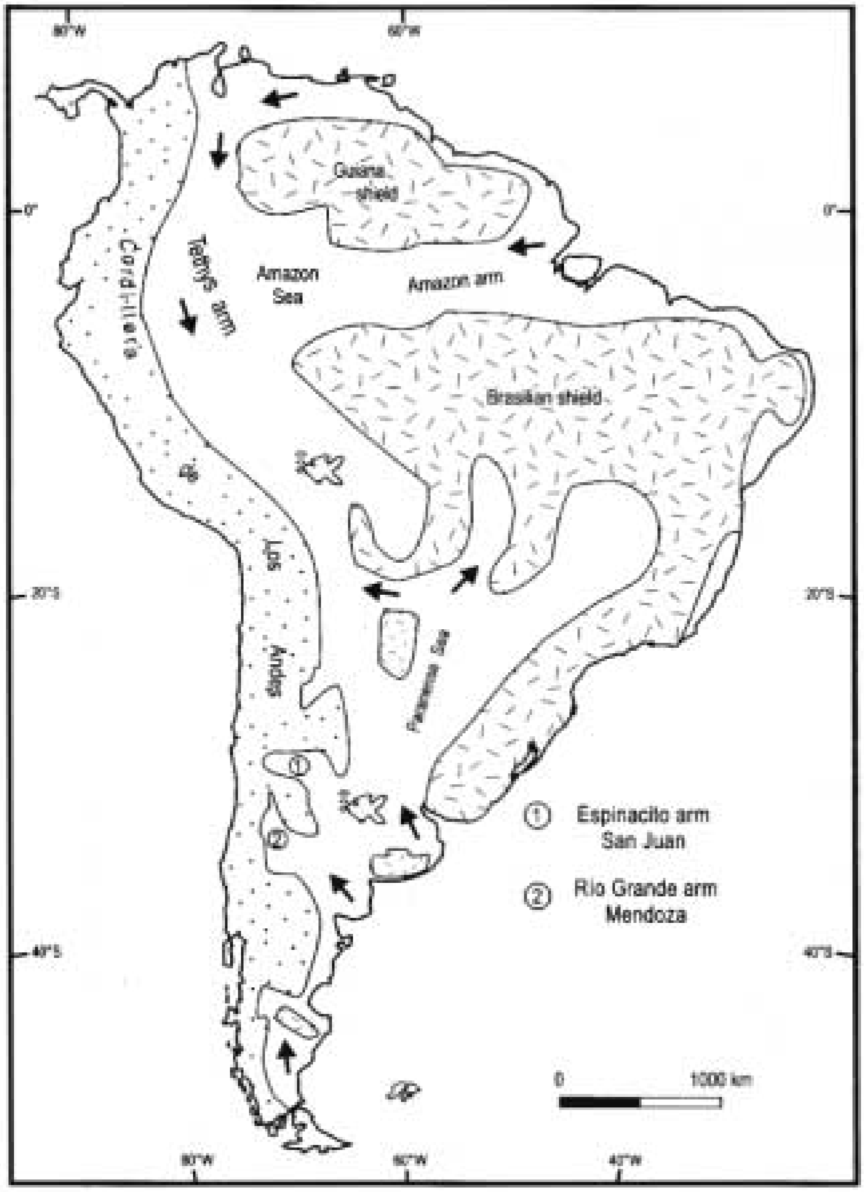


FIGURE 20: Miocene marine transgression into the Andean foothills as a result of rapid tectonic loading of the Andes. Records of these marine deposits are known in central and northern Argentina, subandean region of Bolivia and Peru, and the foothills of Eastern Cordillera of Ecuador and Colombia (modified after Ramos and Alonso, 1995; Pérez et al., 1996).

volcaniclastic rocks were accumulated in a set of eastwards-migrating depocenters. This steady migration since Jurassic times was attributed to tectonic erosion of the continental margin (Ramos, 1988b; Stern, 1991).

A rapid migration of the volcanic front by early to middle Miocene times was related to shallowing of the subduction zone (Jordan *et al.*, 1983). The eastwards shifting of the volcanism was also related to deformation that started *c.* 22 to 20 Ma in the westernmost Chilean séctor and prograded at *c.* 16 - 15 Ma on the Argentine slope of the cordillera.

Maximum flexural subsidence between 15 and 13 Ma was responsible of the marine transgression of the Paranense Sea, which diachronically invaded most of the Andean foothills between these latitudes and N of Colombia up to the Maracaibo area (Fig. 20). Testimonies of this Miocene transgression are preserved as isolated patches at 3200 m a.s.l. in several localities of the Principal Cordillera of central Argentina and in the Puna foothills of northern Argentina, where middle late Miocene foraminifera have been found and the age was also constrained by radiometric dating and magnetostratigraphy (Ramos and Alonso, 1995; Pérez *et al.*, 1996). These marine deposits continue farther N with the middle to late Miocene green shale beds of Santa Cruz de la Sierra in Bolivia where foraminifera have been found (Yecua Formation of Marshall *et al.*, 1993). These marine deposits have been recognized in the foreland basin of Peru (15 Ma Pebas Formation, Hoorn, 1993); in southern Ecuador, with possible Pacific connections through the Guayaquil Seaway in the middle Miocene (Steinmann *et al.*, 1999); and the middle Miocene foreland basin of Colombia (16 to 10.5 Ma León Formation in Cooper *et al.*, 1995; Villamil, 1999). The late middle Miocene sporadic marine transgression in this highly subsiding foreland basin can be traced up to the Maracaibo area (Hoorn *et al.*, 1995). This marine transgression of the Paranense Sea could be correlated with the Amazon Sea of Räsänen *et al.* (1995), if the peak of flooding is correlated with the middle Miocene highstand of 12 Ma (Haq *et al.*, 1987), as discussed by Webb (1995). This connection can explain the Caribbean affinities of the middle to late Miocene foraminifera fauna described by Boltovskoy (1991) in the Argentine offshore platform, which was noticeable absent on the Brazilian platform at that time.

The Andes of central Chile also record at these latitudes an important Miocene deformation favoured by the thermal weakening of the continental crust, that inverted the previous Oligocene intra-arc basin (Godoy *et al.*, 1999).

The shallowing of the subduction zone in the Pampean flat-slab, ends the last stages of volcanism at about 7 - 4 Ma in the Principal Cordillera, and resulted in subduction related magmatism as far as 750 km away from the trench at 1.9 Ma in the eastern Sierras Pampeanas (Ramos *et al.*, 1991; Kay *et al.*, 1991). The migration of the magmatism was accompanied by diachronic deformation and uplift: First, the Frontal Cordillera as a single mountain block (9 - 8 Ma); and subsequently, the thin-skinned Precordillera fold and thrust belt and the basement blocks of Sierras Pampeanas, during Pliocene to Quaternary times (Ramos *et al.*, 1996a, b). Present tectonic activity is shown in the highly seismic area of the Precordillera thrust front and in the western Sierras Pampeanas.

The Neogene was a time of important subsidence in the foothills of the central Argentina foreland basins (Jordan, 1995). During the early Miocene, it was a single extended foreland basin that was cannibalized during middle Miocene times. Subsequent deformation in the late Miocene, and out-of-sequence thrusting at about 7 Ma, originated and reactivated a series of intermontane basins such as the Uspallata and Iglesias piggy-back basins between the Principal and Frontal Cordilleras (Beer *et al.*, 1990; Cortés, 1993). The uplift of Precordillera was not coeval along strike, as demonstrated by magnetostratigraphic analyses in different sectors of the Precordillera (Jordan *et al.*, 1988, 1997). The uplift of western Sierras Pampeanas at about 3 Ma originated the Bermejo broken foreland basin (Fernández and Jordan, 1996; Zapata and Allmendinger, 1996) and the Los Colorados Basin farther N, that respectively accumulated more than 9000 m and 10 000 m of clastic deposits (Fernández and Jordan, 1996; Ramos, 1970).

The eastern Chaco plains of Bolivia and Argentina and the Pampas of central Argentina were affected by widespread dynamic subsidence responsible of few hundred metres of latest Cenozoic sedimentation as far as 1000 km of the Andes thrust front.

Southern Andes

The Patagonian Andes also show the generalized extension during early Mesozoic times with the development of foreland embayments such as the Neuquén and Río Mayo basins. Along the axis of the cordillera an intra-arc setting is observed at the latitude of Lago Fontana (45°S), where an inner and partially subaqueous arc interfingered with Neocomian and older marine deposits, in a complex graben system. The early Mesozoic extension encompassed most of the foreland region, almost to the Atlantic coast (Uliana and Biddle, 1988).

Middle Late Jurassic arc magmatism consists of andesite and dacite along the main Patagonia Cordillera, that after a pulse of lull magmatism during Neocomian times resumed in Aptian to Albian times (Ramos and Aguirre-Urreta, 1994). At this time vast volcanism of dacitic to rhyolitic composition was dominant in the extra-Andean region.

Mild inversion tectonics took place during Late Cretaceous times and partially uplifts the Patagonian Cordillera and the adjacent Precordillera. Oblique subduction in the northern sectors was responsible of the strike-slip and pull-apart basin formation in Paleogene times.

The Ñirihuao Basin between 41°S and 43°S is a retro-arc basin developed during late Eocene times, filled with continental deposits and sporadic marine transgression from the Pacific side (Ramos, 1982). This basin was reactivated as a pull-apart basin formed by transtension in Oligocene times (Dalla Salda and Franzese, 1987). The basin was subsequently inverted during Miocene times, developing a narrow fold and thrust belt (Giacosa and Heredia, 1999).

The Liquiñe-Ofqui Fault (Hervé, 1994), parallel to the Pacific margin along more than 750 km in the present fore-arc region of the Patagonian Andes, is an active fault that shows right-lateral displacement, at least since the Miocene (Cembrano *et al.*, 1996). The wrench fault trace is approximately aligned with the chain of active volcanoes from the Hudson (46°S) to the Lonquimay Volcano (39°S). Strike-slip decreases from S to N due to transfer of displacements to synthetic systems. The Cenozoic stress is partitioned in the fore arc, favoured by the thermally weakened crust of the continental margin (Cembrano *et al.*, 2000), and is not transmitted to the retro-arc (Dewey and Lamb, 1992). The decoupling of the deformation along the active arc by oblique subduction, as suggested farther N by Scheuber *et al.* (1994), left the region E of the arc with almost no deformation during Late Cenozoic times.

The situation dramatically changes S of the present Chile triple junction at about 46°30'S (Fig. 21). The effects of the ridge collision, ophiolitic emplacement, and the consequent triple junction migration show a connection between ridge-trench collision, the cessation of arc volcanism with deformation in the foreland fold and thrust belt and retro arc basaltic plateau volcanism in late Miocene to present times (Stern *et al.*, 1976; Ramos and Kay, 1992; Gorring *et al.*, 1997). Prior to the subduction of the ridge was emplaced the exceptional Cerro Pampa adakite due to partial melt of hot and young oceanic lithosphere (Kay *et al.*, 1993b). Oceanic plate reconstruction (Cande and Leslie, 1986, Tebbens *et al.*, 1997) suggests that Eocene retro-arc plateau magmatism may also be associated with ridge-trench interactions.

The geological and magmatic history of the eastern foothills at these latitudes shows a striking coincidence between shortening and uplift of the southern Patagonian Andes (Fig. 22), the arc gap and the distribution and chemistry of the volcanic arc rocks to the N and S, and the timing and chemistry of alkali plateau basaltic activity (Gorring *et al.*, 1997).

The southern Patagonian Andes S of 46°30'S due to these ridge collisions, have a 2 km higher topography, an important orogenic shortening generated by a fold and thrust belt and extensive foreland basin subsidence as seen in the Austral or Magallanes Basin (Ramos, 1989b; Biddle *et al.*, 1986). This sector is also characterized by the late orogenic emplacement of Miocene stocks of granitic composition such as San Valentín, San Lorenzo, Fitz Roy and Torres del Paine (Ramos *et al.*, 1982).

The distribution and timing of Eocene arc and plateau magmatism also appears to correlate with a ridge-trench

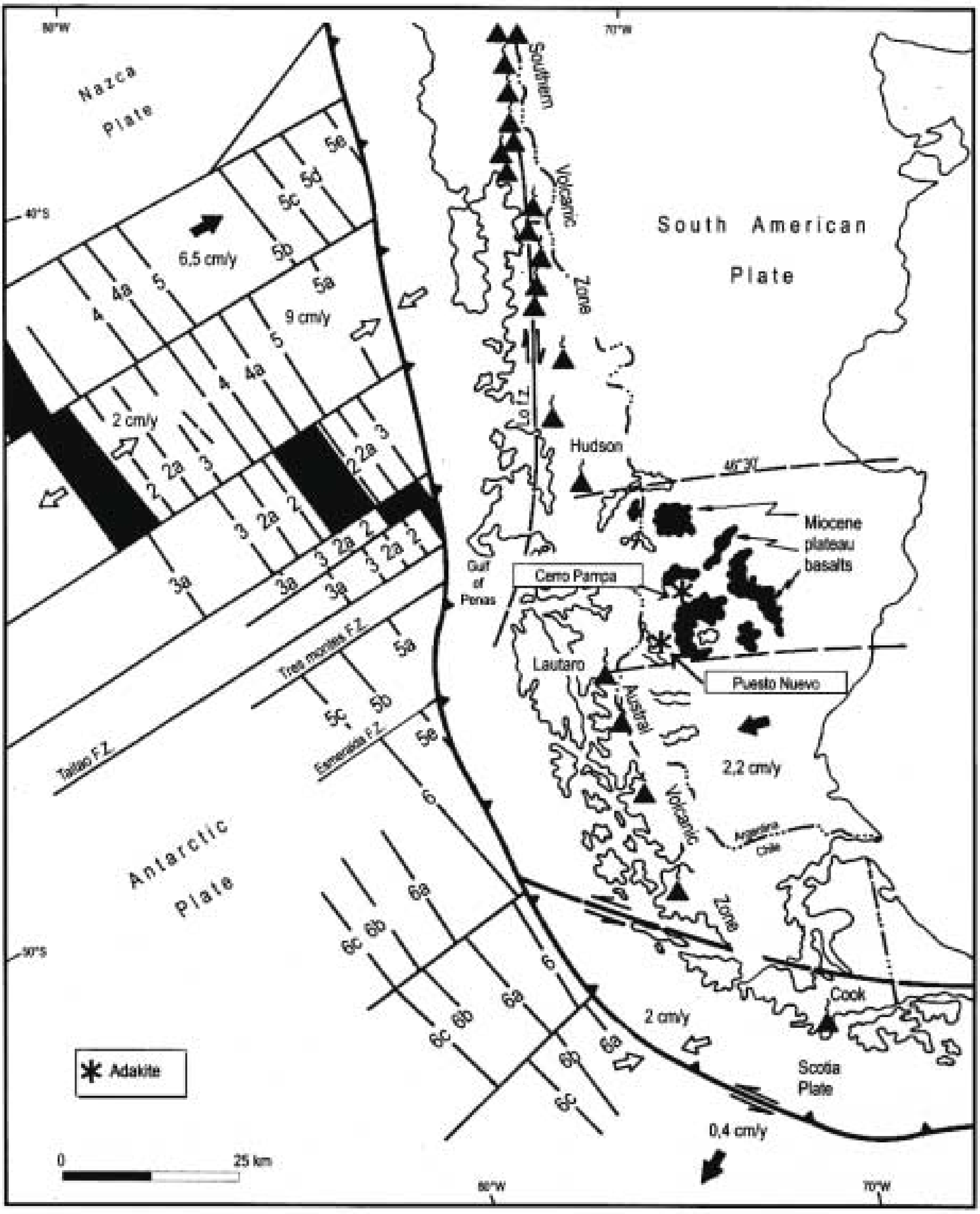


FIGURE 21: Plate tectonic framework of the Patagonian Andes N and S of the triple junction with location of the Southern and Austral volcanic zones (modified after Ramos and Kay, 1992; Gorring et al., 1997). LO F.Z.: Liquiñe-Ofqui Fault Zone.

collision (Fig. 23). In this case the ridge that separated the Farallon and Aluk plates collided at a much lower angle to the coast. The effects of the ridge collision appear only S of 43°S where extensive alkali plateau basalt was developed and the volcanic arc was absent (Ramos and Kay, 1992). Comparison of the magmatic rocks erupted in the retro-arc behind three different collided Chile ridge segments suggests that the volume of magmatism and the percentage of melting increases with the length of the ridge segment and the size of the slab window (Gorring *et al.*, 1997).

Back-arc basin formation and tectonic inversion in the Fueguian Andes

The generalized Early Mesozoic extension observed through all the Andean chain was maximum in the Fueguian Andes where the Rocas Verdes marginal basin was formed (Fig. 24) during latest Jurassic-Early Cretaceous times (Dalziel *et al.*, 1974). To the S of 52°S, along the Patagonian channels, the ophiolite rocks of the Cordillera de Sarmiento are exposed. The Sarmiento Ophiolite as well as the Tortuga Complex in the Fueguian Archipielago indicate the longitudinal development of oceanic crust along the axis of the back-arc basin (Stern *et al.*, 1976b; Dalziel, 1981).

This extension was also detected in recent seismic lines as basement-level normal faults (Klepeis and Austin, 1997). The regional extension produced the attenuated crust and the rhyolitic rocks and subordinated clastics of middle to Late Jurassic age that filled the half-graben system across the extra-Andean region (Suárez, 1976; Uliana and Biddle, 1988). Thermal subsidence produced the extensive Springhill clastic platform that expanded during the early Cretaceous throughout the entire extra-Andean region.

Closure of the marginal basin in middle Cretaceous times was followed by intense deformation and thrusting with northerly vergence (Mpodozis and Ramos, 1990).

Rocks containing upper amphibolite facies assemblages reflecting Mesozoic-Cenozoic metamorphism are exposed in the Cordillera Darwin (approximately. 55°S). The exhumation of these unique high-pressure rocks has been interpreted as having been produced by extension in a metamorphic core complex (Dalziel and Brown, 1989) after the important crustal thickening of the Late Cretaceous; low-angle extensional crustal shearing bounding the high grade metamorphic complex along the Beagle Channel, would be responsible of the tectonic denudation in latest Cretaceous-early Paleocene times. This hypothesis was challenged by Klepleis (1994) who interpreted the Cordillera Darwin as a 8 kbar basement uplift produced by thick-skinned thrusting that followed the thin-skinned deformation of the Magallanes Basin (Fig. 25). This late thrusting was produced out-of-sequence between 65-40 Ma, compatible with cooling ages and deformation history of Cordillera Darwin (Nelson, 1982; Kohn *et al.*, 1995). More recent seismic data permit the calibration of the final uplift of Cordillera Darwin during the shift away from compression towards transtension during late Tertiary crustal relaxation (Klepeis and Austin, 1997).

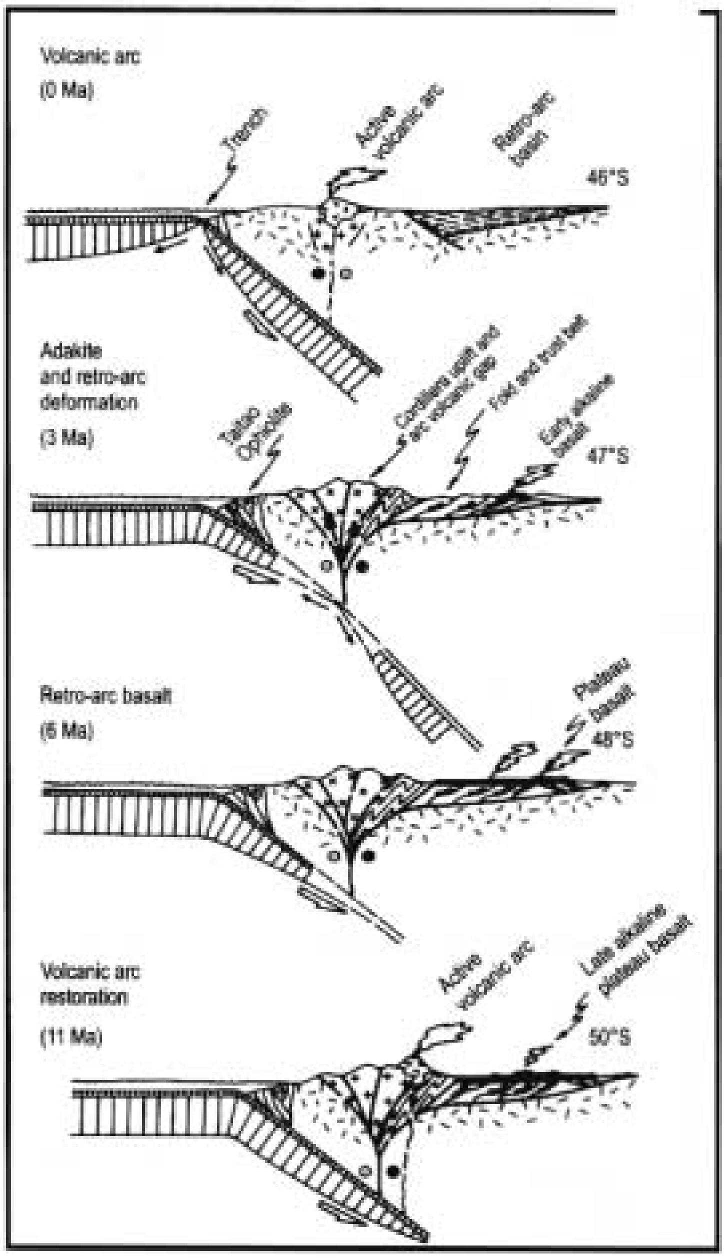


FIGURE 22: Tectonic evolution of the southern Patagonian Andes during ridge collision of a seismic ridge (modified after Stern et al., 1976a; Ramos and Kay, 1992).

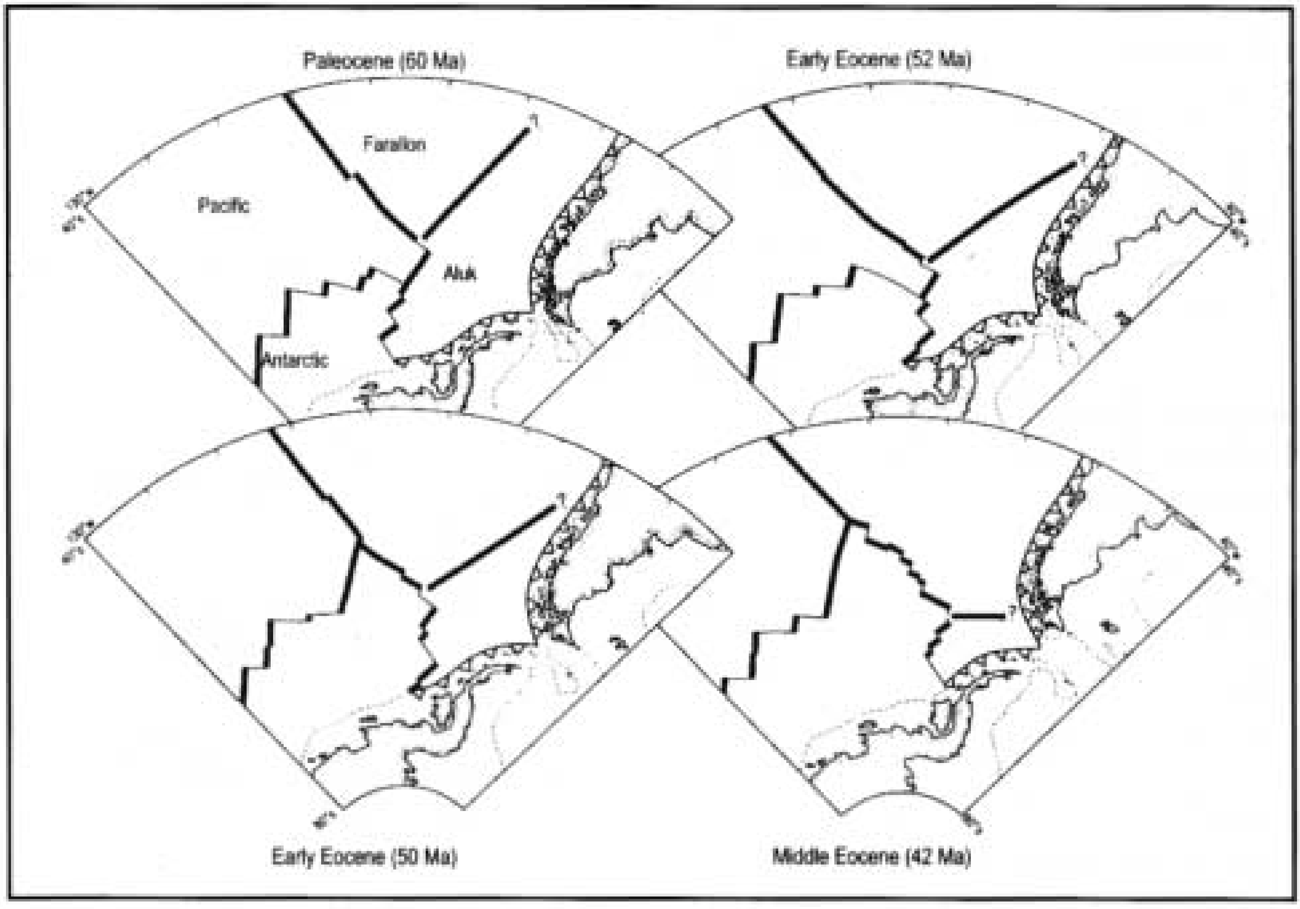


FIGURE 23: Ridge collisions along the Pacific margin based on the Paleogene and Neogene plate reconstruction of Cande and Leslie (1986).

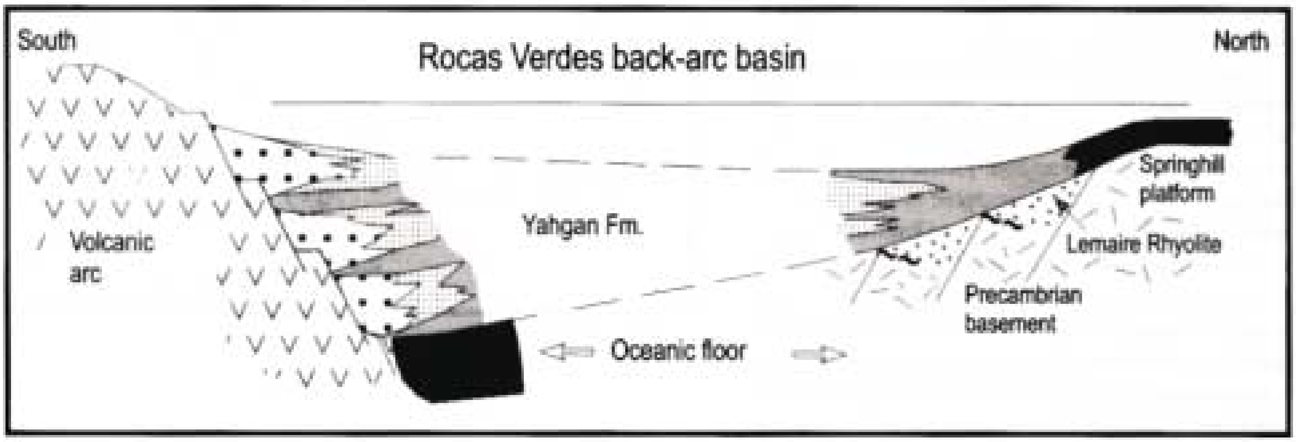


FIGURE 24: Rocas Verdes back-arc basin during Cretaceous times (modified after Dalziel et al., 1974; Olivero, 1998).

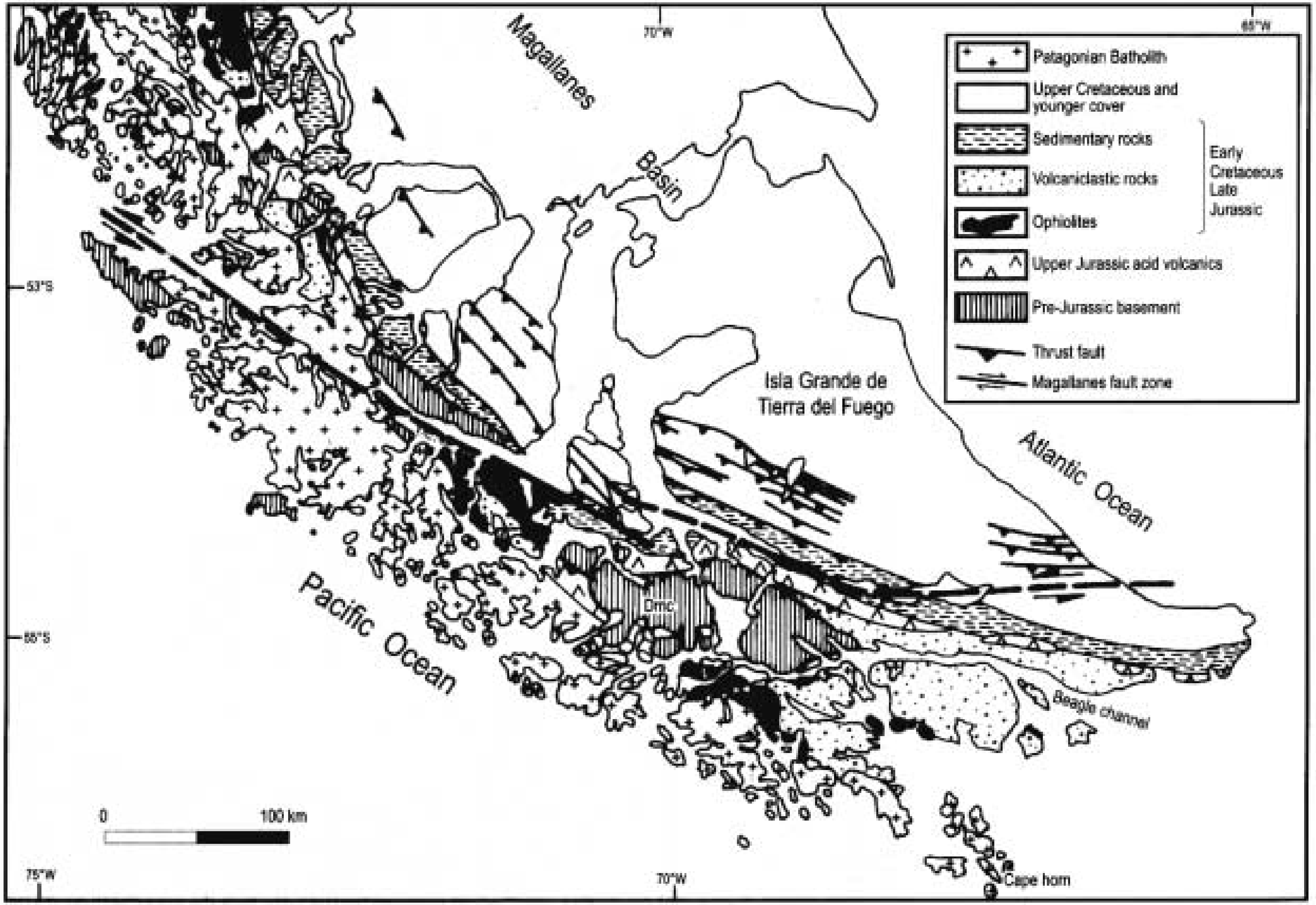


FIGURE 25: Tierra del Fuego tectonic setting with major strike-slip faults and fold and thrust belts with location of the ophiolitic assemblages of the Rocas Verdes Basin. DMC corresponds to Cordillera Darwin Metamorphic Complex (modified after Klepeis and Austin, 1997).

Important shortening in the E-W trending segment of the Fueguian Cordillera occurred in middle to late Eocene times, as demonstrated by the growth strata and limb rotation along the Atlantic coast of the Fueguian thin-skinned fold and thrust belt (Ghiglione *et al.*, 1999; Olivero and Malumián, 1999). Miocene deformation produced the present uplift of the Patagonia fold and thrust belt (Biddle *et al.*, 1986; Ramos, 1989b).

Neogene transtensional and transpressional deformation are still active along the present plate boundary between South America and Scotia plates. The Patagonia orocline appears to be the product of broad interplate shearing accommodated by strike-slip faulting, block rotation, and contraction (Cunningham, 1993).

Present plate tectonic setting of the Andes

As result of the complex and distinctive geological history of the different segments, and the oceanic plate interaction along the margin, the present tectonic setting of the Andes shows a great variety of plate tectonic conditions.

Seismicity and subduction geometry

The feature that most strongly characterizes the subduction geometry beneath the Andean Cordillera is the along-strike variation in dip of the subducted Nazca Plate from subhorizontal flat-slab segments to normal subduction (Barazangi and Isacks, 1976; Pennington, 1981; Cahill and Isacks, 1992). The seismic energy released above flat-slab segments is on average 3 or 5 times higher than in the adjacent steeper areas (Gutscher and Malavieille, 1999). Based on the interplate seismicity, several segments of flat-slab subduction have been recognized along the Andes (Fig. 26).

The Northern Andes N of 5°N correspond to the Bucaramanga segment, where flat-slab subduction has been recognized along the Colombian margin (Pennington, 1981). In this area important intraplate seismicity is recorded in the upper plate, and crustal thickening is combined with significant strike-slip motion. The large Bucaramanga earthquakes characterize the notable intraplate activity of this segment (Kellogg and Bonini, 1982) as well as the lack of active volcanism (Hall and Wood, 1985).

The other flat-slab segment corresponds to a 1500 km long sector of Peruvian Andes and is preserved between the gulfs of Guayaquil and Arequipa (5°-14°S). Along this segment of the Central Andes occurs the Cordillera Blanca tectonic uplift, one of the highest regions in the Andes. Important intraplate shallow seismicity is detected in the Eastern Cordillera and the Subandean Zone as reported by Suárez *et al.* (1983) and Dorbath *et al.* (1991), that accounts for present 4 mm/y shortening in central Peru. This segment is also characterized by the lack of volcanism and by a subhorizontal oceanic slab, dipping about 5° to the E and NE for several hundred kilometres (Barazangi and Isacks, 1976). Near 14°S there is an abrupt transition to a more steeply inclined zone (Cahill and Isacks, 1992). This flat-slab segment has been explained by the combined buoyancy of the Inca Plateau and the Nazca Aseismic Ridge subducted beneath the continental margin of Peru (Gutscher *et al.*, 1999a).

The third and southern subhorizontal subduction zone corresponds to the Pampean flat-slab segment, and is developed between 27°S and 33°S. The geometry of this segment is well established through a local seismic network (Smalley and Isacks, 1990) that defined a 300 km horizontal segment, with resumption of eastward descent below 125 km farther E (Cahill and Isacks, 1992). This segment focuses a high intraplate seismicity and is characterized by the lack of volcanism (Fig. 27) and the foreland uplift of Sierras Pampeanas (Jordan *et al.*, 1983). As a result, it concentrates the highest mountains along the Main Andes, such as the Aconcagua Massif (Ramos *et al.*, 1996b, Cristallini and Ramos, 2000).

Between the Bucaramanga and the Peruvian flat-slab segments there is a normal subduction sector where the slab dips 35° in the Cauca and Ecuador segments (Pennington, 1981; Gutscher *et al.*, 1999b). Farther S, between Arequipa and northern Argentina there is the most important normal subduction sector, where the slab inclines 30° to the E. This sector of southern Peru-Bolivia-northern Argentina, has an abrupt dip change between 14° and 16°S in the N, and a smooth transition between 24° and 27.5°S in the S. This central sector encompasses the best developed mountain chains, and is the paradigm of Andean-type orogen (Allmendinger *et al.*, 1997). To the S of 33°S the Andes have small variations in the Wadati-Benioff Zones, and a decreasing seismic activity. The normal subduction extends from 33°S to 46°30'S up to the Chile triple junction between the Nazca, South American, and Antarctic plates.

The shallow seismicity located within the Andes decreases drastically to the S of the Northern Andes, where shallow epicenters are restricted to two seismic belts, one in the fore arc and the other in the Subandean Belt (Ego *et al.*, 1996). These authors related the shallow seismicity and the absence of important seismicity in the Subandean Belt to the obliquity of the subduction. Sectors with more orthogonal subduction developed an active Subandean Belt, with little seismicity beneath the main Andes (Suárez *et al.*, 1983).

The Wadati-Benioff geometry dips at 30° to the E between 33°S and 36°S, and is deepening up to about 40° farther S (López *et al.*, 1997).

To the S of the triple junction seismicity resumes, as well as the convergent rate, that drops from 9 to 2.2 cm/y (Corvalán, 1981). The almost orthogonal subduction gives place to a more oblique subduction around 50°S with dominant strike-slip focal mechanisms, to end in a strike-slip transform boundary with the Scotia Plate (Dalziel, 1986).

Volcanic arc and related magmatism

The subduction geometry defines four distinctive zones of active volcanism: the Northern Volcanic Zone (5°N to 2°S) developed along the Cauca and Ecuador segments; the Central Volcanic Zone (16°S to 26°S) between southern Peru and Northern Chile; the Southern Volcanic Zone (34°S to 46°30'S); and the Austral Volcanic Zone, S of 47°S (Thorpe and Francis, 1979; Thorpe, 1984; Stern and Kilian, 1996). Each volcanic zone has its own peculiarities.

The Northern Volcanic Zone comprises a series of active volcanoes developed in the Western and Central Cordilleras

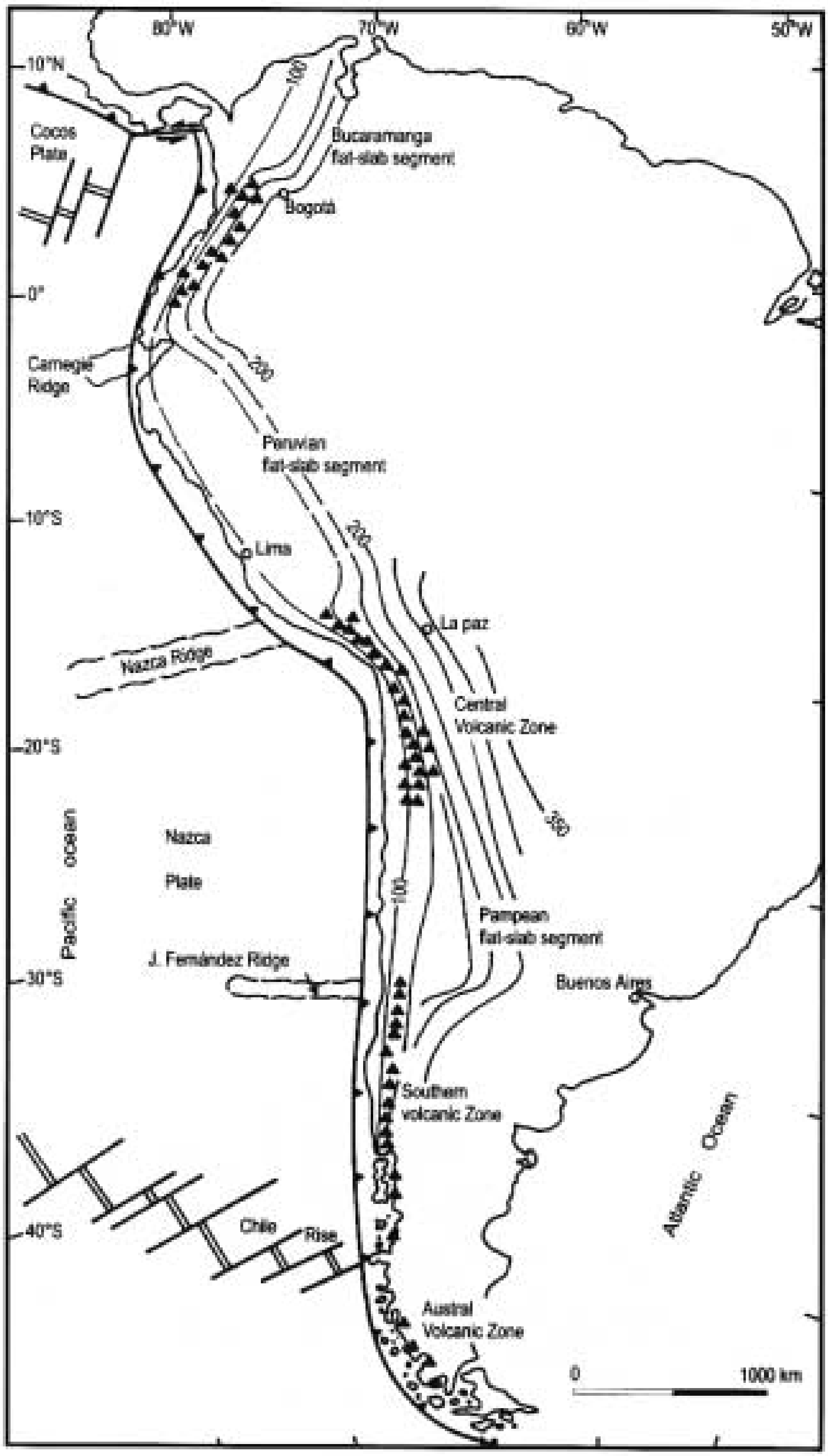


FIGURE 26: Geometry of the subduction zone along the Andes with indication of the major segments (modified after Pennington, 1981; Cahill and Isacks, 1992; López et al., 1997; Gutscher et al., 1999a).

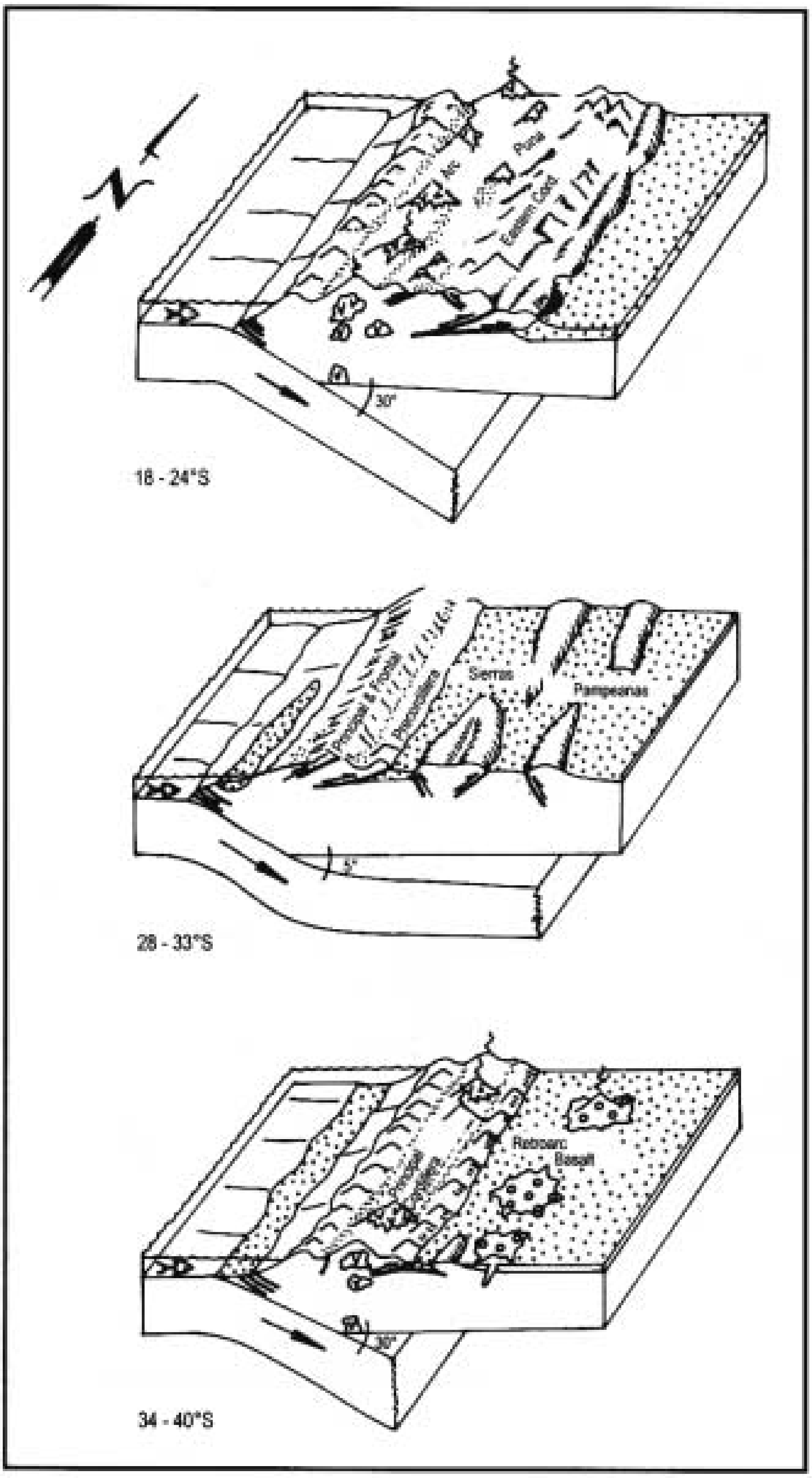


FIGURE 27: Present segmentation of southern Central Andes showing the main morphostructural units of the normal subduction and Pampean flat-slab segments (modified after Jordan et al., 1983).

of Colombia to the S of 5°N, such as the Nevados de Ruiz, Galeras, and Cerro Bravo (Méndez Fajury, 1989). This zone continues on both sides of the Interandean Valley of Ecuador where there occur several volcanoes such as the Mojanda, Chimborazo, and Pichincha persist up to 2°S (Hall and Beate, 1991; Robin *et al.*, 1997). Some other volcanoes are developed in the eastern foothills of the Andes, such as the Reventador and Sumaco (Hall and Calle, 1982).

The volcanic rocks of these volcanoes range from basaltic andesite to andesite, and are more primitive than the products of the Central Volcanic Zone (Thorpe, 1984). They are situated in the vicinity of the suture between the Piñón-Dagua oceanic terrane and other Jurassic terranes accreted during Early and Late Cretaceous to the Paleozoic proto-margin of Gondwana. Their volcanic rocks are consistent with derivation from fractional crystallization of basaltic magma produced from partial melting of the asthenosphere wedge containing components from oceanic lithosphere. However, their low to moderate Sr^{87}/Sr^{86} ratios (0.7036 to 0.7046), as well as the higher Pb and O-isotope ratios may indicate some assimilation of younger continental crust (Harmon *et al.*, 1984). Lava flows, small pyroclastic flows, and large lahar deposits, associated with ashfall deposits widely cover the Interandean Valley of Ecuador (Hall and Calle, 1982).

The Central Volcanic Zone is widely developed between Arequipa and northern Chile, along the Western Cordillera, which is bounding the Altiplano-Puna high plateau. Hundreds of volcanoes are widely spread along this region. These volcanoes are characterized by their eruption in a thick crust, in places over 70 km thick, and record high degrees of differentiation. Stravolcanoes constructed largely from andesitic and dacitic lava are dominant in additon to which there occur significant volumes of dacitic ignimbrites of latest Cenozoic age (Davison *et al.*, 1993). A Sr^{87}/Sr^{86} ratio varies from 0.7056 to 0.7149 (Harmon *et al.*, 1984). There is also a striking correlation between the Pb-isotopes of the volcanic rocks and those of the underlying basement (Wörner *et al.*, 1994). These facts, together with the chemical and isotope composition, led Hildreth and Morbath (1988) and Davidson *et al.* (1993), to assume that extensive modification of the mantle-derived magma took place as they ascended through an exceptionally thick crust. The amount of contamination has been evaluated in the order of 35% to 70% (Hawkesworth and Clarke, 1994). On the other hand, subduction of oceanic and terrigenous sediments into the asthenosphere wedge, as well as subduction erosion, could have contribute to the differentiation of the mantle-derived magma (Stern, 1991). Subduction geometry changed during Late Cenozoic times as recorded by the magmatism and by an extensive episode of rhyolitic and dacitic calderas and ignimbritic flows formed during an episode of steepening of the Wadati-Benioff Zone (Coira *et al.*, 1993; Kay *et al.*, 1999). This extensive silicic volcanic province produced in the late Miocene, known as the Altiplano-Puna volcanic complex, covers more than 50 000 km^2 and is one of the largest ignimbrite concentration in the world (De Silva, 1989). Teleseismic broadband records of earthquakes in this province at Bolivia have identified an active crustal magma flat body at 19 km depth, very extensive and 750-810 m thick (Chmielowski *et al.*, 1999). The southern part of this Central Volcanic Zone records a crustal delamination associated with mafic magma and extension (Kay *et al.*, 1994).

The Southern Volcanic Zone is developed between 33°30'S and 46°30'S, and corresponds with the southern part of the Central Andes. It comprises late Cenozoic and active volcanoes such as Tupungato, San José, Lonquimay, and Hudson, mainly developed in the Chilean slope of the main Andean Cordillera. The northern sector of this 1000 km long volcanic chain has more crustal influence and is characterized by andesite and dacite (López Escobar *et al.*, 1995). South of 37°S the volcanic province consists of rocks of basalt to rhyolite composition, but with a predominance of basalt and basaltic andesite with low Sr^{87}/Sr^{86} ratios (0.7037 to 0.7044). This volcanic province is heavily controlled by the onset of important strike-slip faults as the Iquiñe-Ofqui (Hervé, 1994). The increased angle of subduction S of 35°S from 30° to near 40° beneath the volcanic zone as well as the migration toward the trench recorded since Pliocene times, may account for a minimum coupling between the Nazca and the South American plates, and the dominant poorly differentiated volcanism (Stern, 1990).

The Austral Volcanic Zone has been recently defined by Stern and Kilian (1996). This volcanic zone consists of a few volcanoes developed in the Southern or Patagonian Andes S of the volcanic gap (Fig. 21) associated with the ridge subduction (Stern *et al.*, 1976b). Adakitic volcanic rocks of low Sr^{86}/Sr^{87} ratio, formed by components of the asthenosphere wedge, plus a partial melting of the subducted slab constitute the poorly evolved lava of the Lautaro, Aguilera, Diablo, Burney and Cook volcanoes (Stern and Kilian, 1996).

The areas of slab-flat subduction record the shifting, expansion, and the cessation of the volcanic arc through time, with striking compositional changes, declining volumes of volcanic rocks, and unusual petrological characteristics (Kay *et al.*, 1991; Ramos *et al.*, 1991; Kay and Abbruzzi, 1996).

The normal subduction southern segment of the Central Andes records an important retro-arc basaltic magmatism of alkaline composition. It is associated with trenchward migration of the volcanic front (Muñoz and Stern, 1988); a decreasing age of the oceanic crust being subducted (Ramos and Barbieri, 1989) and the presence of transient hot spots (Kay *et al.*, 1993a). Farther S, along the Southern Andes, asthenosphere windows formed after by ridge subduction controls the near trench magmatism, adakite and retro-arc plateau basalt (Ramos and Kay, 1992; Kay *et al.*, 1993b).

Crustal thickening and orogenic shortening

The convergence vector and subduction rate between South America and the Farallon Plate during the Paleogene (Pilger, 1984) favoured a more orthogonal convergence on the Peruvian margin and in the Fueguian Andes. The continental margins in these sectors have northwestern trends, and therefore the effects of the high convergence rates by the end of the Eocene were more severe. These two segments show an important compression during the Incaic deformation (Vicente, 1972; Galeazzi, 1996). On the other hand, along the Chilean margin stress partitioning of this oblique convergence reactivated strike-slip displacements (Mpodozis *et al.*, 1994) along the Domeyko Fault System and the western fissure (Tomlinson *et al.*, 1994).

The break-up of the Farallon Plate into the Cocos and Nazca plates, which occurred at about 25 Ma, seems to mark the beginning of a period of higher and more orthogonal convergence rates in most of the Central and Southern Andes (Pardo Casas and Molnar, 1987). This age coincides with the initiation of widespread Miocene magmatism and is a milestone in the geodynamic evolution of the area. The convergence vector and subduction rate were more important along the Chilean margin, during the middle and late Miocene (26 - 8 Ma), time of onset of major Miocene deformation generally assigned to the Quechua Orogeny (Mpodozis and Ramos, 1990).

The Northern Andes underwent an important stress partitioning during the Paleogene and the Neogene, due to the northeastern trend of the continental margin. As a result of that the oblique component of plate motion is taken up by dextral slip displacements in a series of crustal discontinuities (Fig. 11), such as the Pujili-Cauca and Peltetec-Romeral faults (Dewey and Lamb, 1992; Mégard, 1987).

As a result of several pulses of compression, generally grouped in the Incaic and Quechua deformations, important crustal thickening took place along the Andes. Maximum crustal thickening is observed along the central sector of the Central Andes, where orogenic shortening was the largest (Allmendinger *et al.*, 1997). New crustal balances made across Northern Chile and Bolivia, combined crustal shortening constrained by refraction seismic and partial seismic reflection surveys, and orogenic shortening derived from structural cross-sections (Fig. 28) (Schmitz, 1994; Kley *et al.*, 1999). These balances indicate up to 320 km of total shortening during the Cenozoic. This area also has the thickest crust of the Andes beneath the Western Cordillera at about 20°S. Broadband seismic analyses indicate a crust some 70-74 km thick (Beck *et al.*, 1996).

There is a well-defined gradient to the N and S of this shortening, coherent with fore-arc rotation constrained by paleomagnetism (Beck, 1998). The crustal thickening and orogenic shortening gradients are well established in the southern segment of the Central Andes (22°S to 46°30'S). There is a continuous decline in the thickening of Andean roots with orogenic shortenings from 160-40 km at 30°S-32°S (Introcaso *et al.*, 1992; Ramos *et al.*, 1996b) up to 44 to 20 km at 37°S-39°S (Martínez *et al.*, 1997).

This decrease in crustal thickness and orogenic shortening is also observed from the Altiplano to the Peruvian Andes (Cabassi *et al.*, 1999). Both gradients, from the Central Andes to the N and S, can be correlated with a decrease in the age of oceanic crust being subducted. There is a continuous trend of younger oceanic crust along the trench as the Chile Ridge gets closer to the present triple junction (Tebbens *et al.*, 1997).

Orogenic shortening and tectonic styles

The estimates of crustal thickening and the consequent orogenic shortening impose important constraints to the tectonic style of the Andean Cordillera. Studies in the last 20 years have shown contrasting attempts to understand the mode of structural deformation along the different foreland thrust belts of the Andes. Models with high angle basement thrusts (Mégard, 1987; Zeil, 1979) alternate in similar regions with thin-skinned deformation (Vicente, 1972; Ramos, 1988b; Allmendinger *et al.*, 1990). Tectonic inversion of previous normal faults in recent years has become one of the significant mechanism of thrusting (Daly, 1989; Grier and Dalmeyer, 1990; Manceda and Figueroa, 1995; Ramos *et al.*, 1996b; Cristallini *et al.*, 1997; Colletta *et al.*, 1997; Kley *et al.*, 1999). Those areas where seismic control is suitable either show: (1) unequivocal evidence of thin-skinned thrusting as the Subandean fold and thrust belt of Bolivia and Northern Argentina (Baby *et al.*, 1992; Mosquera, 1999); or (2) salt detachment of the Santiago fold and thrust belt of Peru (Alemán and Marksteiner, 1997); or (3) tectonic inversion with subordinate thin-skinned tectonics as in the Eastern Cordillera of Colombia (Cooper *et al.*, 1995). However, in areas where the seismic control is inappropriate, crustal balance can give a further constraint for the amount of shortening. For example, in the Subandean Neuquén Basin (37-39°S), several authors have proposed different tectonic styles from thin to thick-skinned deformation varying from hundred to ten of kilometres. As indicated by geophysical studies the crustal shortening does not exceed 44 km (37°S) to 20 km (39°S). These data constrain the structural style to thick-skinned systems such as proposed by Kozlowski *et al.* (1993) or Zapata *et al.* (1999), with less than 40 km shortening.

In the Northern Andes, where stress partitioning superimposed important strike slip-displacements, crustal thickening of Cenozoic deformation is more complex to evaluate. Shortening in the most active Neogene region of Eastern Cordillera of Colombia shows a minimum of 68 km (Cooper *et al.*, 1995) of the same order than the Mérida Andes with 50 to 60 km (Colletta *et al.*, 1997).

In spite of the overwhelming evidence of the importance of tectonic inversion of normal faults in the Andes, there is good testimony in some areas for thin-skinned deformation (Ramos *et al.*, 1996b). Stress partitioning under oblique subduction as proposed by Dewey and Lamb (1992) produced notable strike-slip displacements as detected in the Northern Andes (Campbell, 1968; Dengo and Covey, 1993), or in the Southern Andes (Hervé, 1994; Diraison *et al.*, 1998).

Neotectonics and orogenic shortening

As an active orogenic belt, the Andes display significant neotectonics along the thrust front of the Central Andes and strike-slip displacements in the Northern Andean block and in the Fueguian Andes (Fig. 29).

Active faults with notable strike-slip displacements have been documented in the Boconó Fault of the Mérida Andes by Schubert (1982) and Schubert and Vivas (1993). Similar settings have been described in the Eastern Cordillera of Colombia with active faulting in the Garzón Massif and in the Guaicáramo Fault (Van der Wiel, 1991). Farther S, active faulting is recorded in the Ecuadorian Andes in the Interandean Valley, associated with Quaternary volcanism (Hungerbuehler *et al.*, 1996).

Inversion of focal mechanisms in the Northern Andes, together with neotectonic analyses, have shown differences

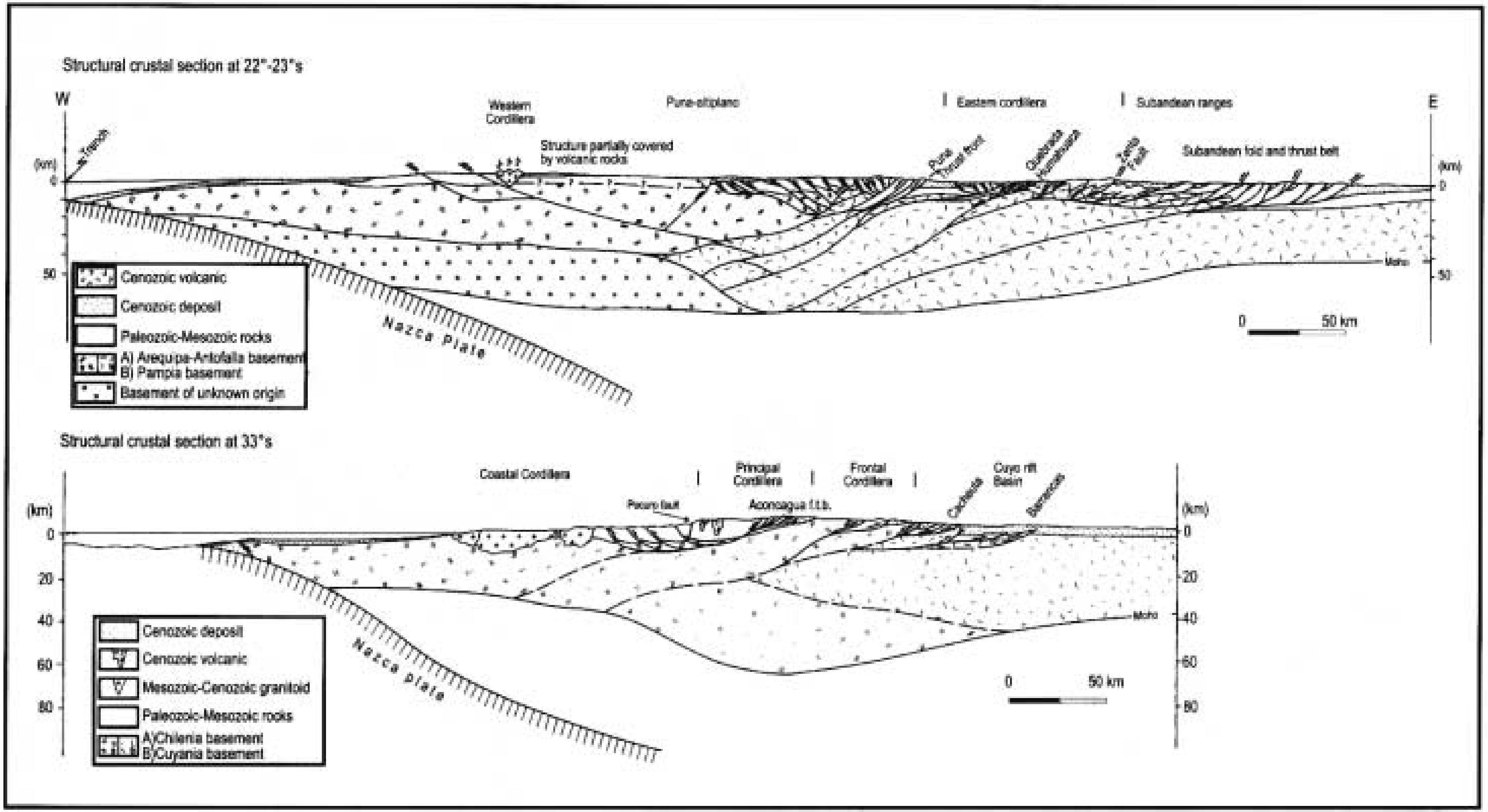


FIGURE 28: Crustal sections of the Andes showing the decrease in crustal thickness from the Altiplano region to the Southern Andes (22°S to 41°S)(modified after Schmitz, 1994; Introcaso et al., 1992; Ramos et al., 1996b).

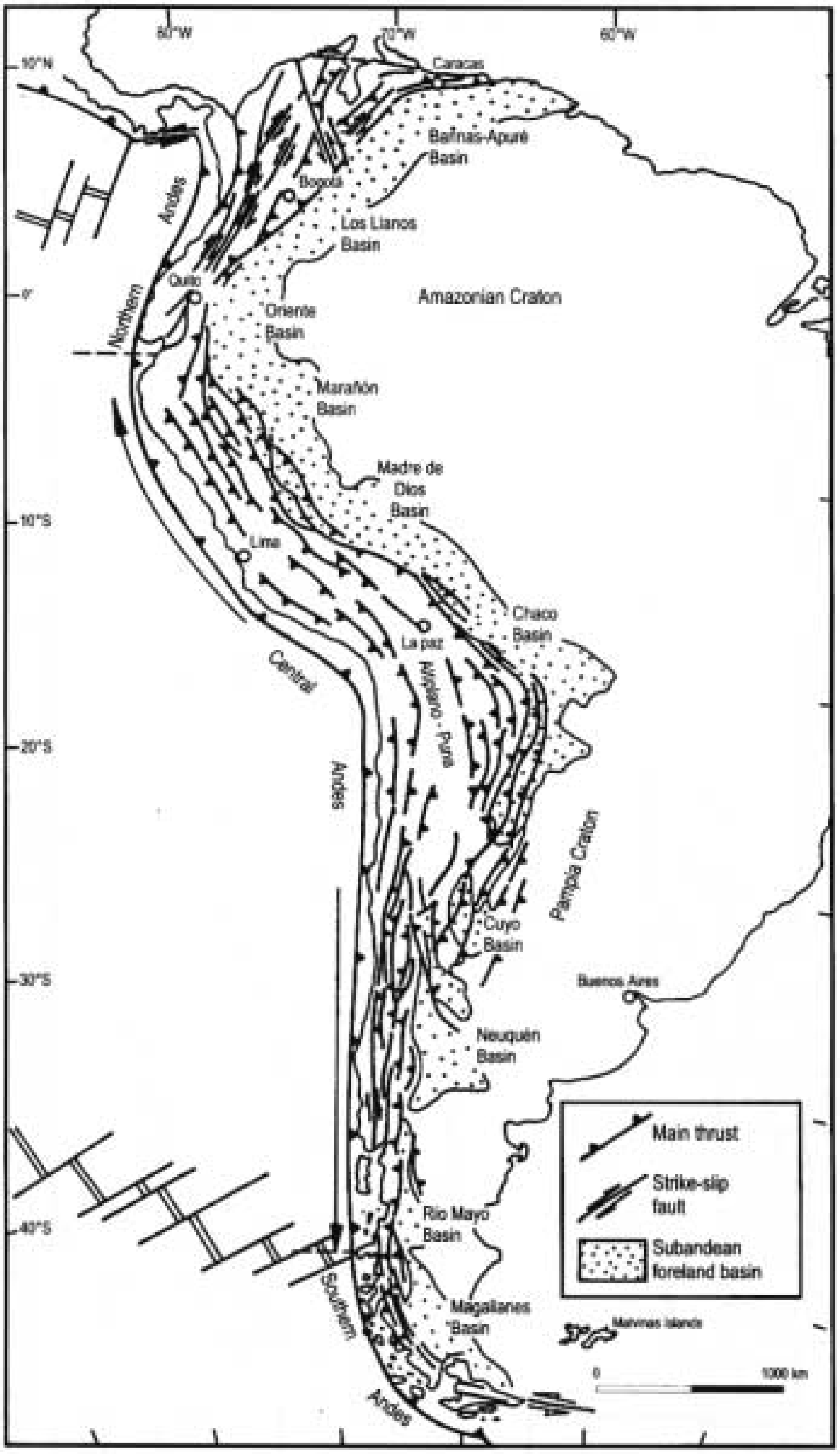
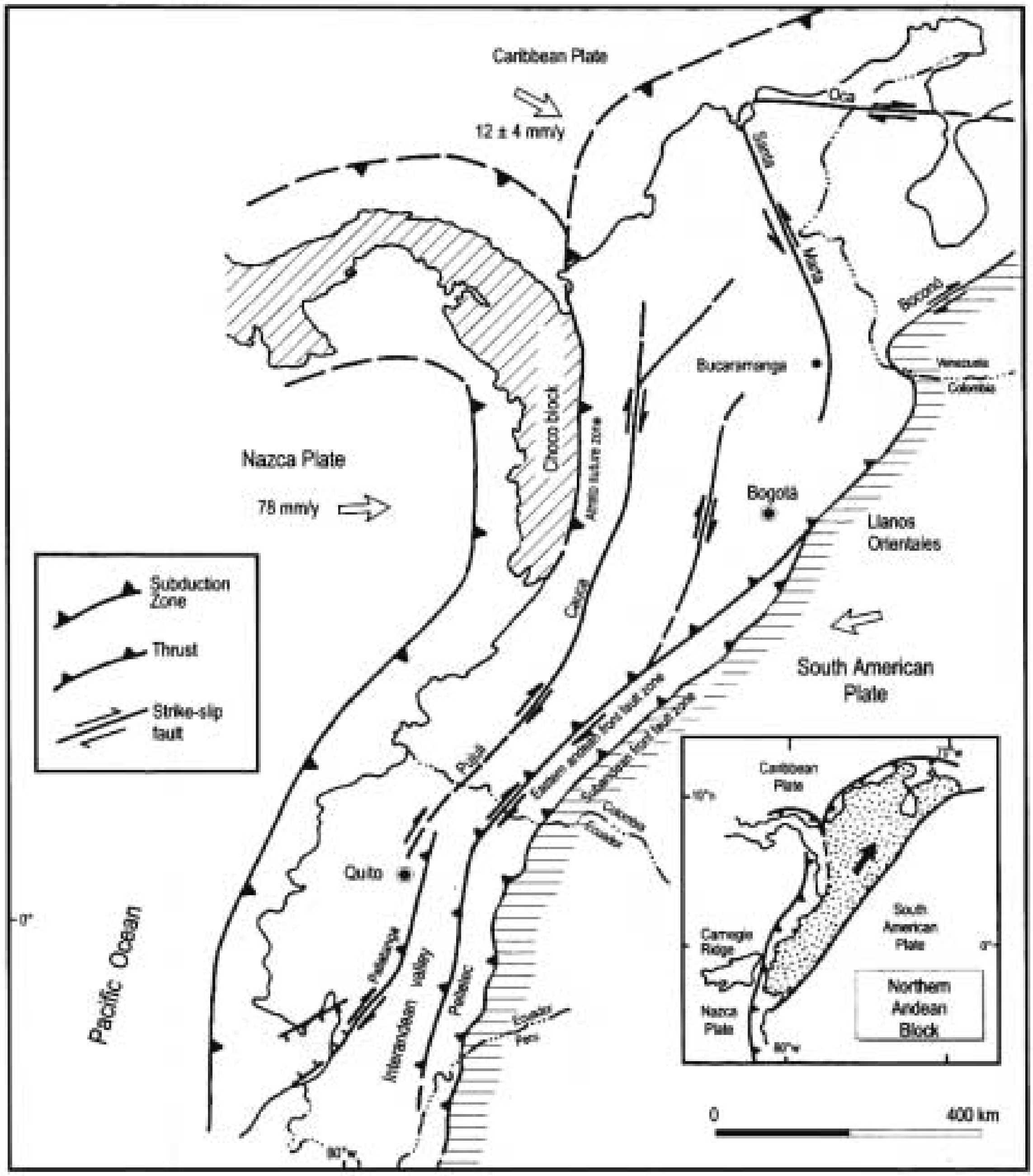


FIGURE 29: Major structural features of the Andes with major thrust fronts. Note the strike-slip faulting of the Northern Andes, and the dominant wrenching and dismembering of the Fueguian Andes (modified after Ramos, 1999).

FIGURE 30: Major active fault in the Northern Andean Block (modified after Kellogg and Bonini, 1985; Ego et al., 1996). Inset a) shows the North Andean block or microplate.



in the state of stress N and S of 5°N (Ego *et al.*, 1996). The northern sector seems to be controlled by the interaction between the Caribbean and South American plates (Fig. 30), coherent with dextral motion of the Northern Andean Block relative to stable South America as proposed by Kellogg and Bonini (1985).

This relative dextral motion of the Northern Andean Block was intensified in the last 9 Ma by the collision of the Carnegie Aseismic Ridge (Daly, 1989). As a result, a vertical uplift of 6 km took place since the late Miocene (Steinmann *et al.*, 1999), as well as important dextral strike-slip along the Peltetec-Romeral and Pujili-Cauca fault systems. Topographic evidence along the coast indicates that collision with the trench took place at least 8 Ma ago (Gutscher *et al.*, 1999b).

Along the Peruvian Andes contrasting active tectonics took place in the flat-slab segment with contraction and reverse faults (Schwartz, 1988; Sébrier *et al.*, 1988), in comparison with normal active faults at the northern end of the Altiplano (Sébrier *et al.*, 1985). While continuous foreland blind thrusting is seen in the Chaco plains E of the Subandean Belt (Mujica and Zorzín, 1996), at the southern end of the Altiplano-Puna high plateau active faulting is dominated by normal faulting (Allmendinger *et al.*, 1997). This normal faulting was explained by collapse originated in body forces by Froidevaux and Isacks (1984) or by stress partitioning of oblique strike-slip faults by Dewey and Lamb (1992).

Farther S, active faulting is occurring in the Pampean flat-slab segment (27°S-33°S) where important shortening has been described in the Precordillera and Sierras Pampeanas (Bastías *et al.*, 1990; Costa, 1992). This activity declines to the S, where in the southern normal subduction sector of the Central Andes (34°S-38°S), is less conspicuous, until it disappears between 39°S and 46°30'S. Active faulting is again important at the latitude of the Chile triple junction,

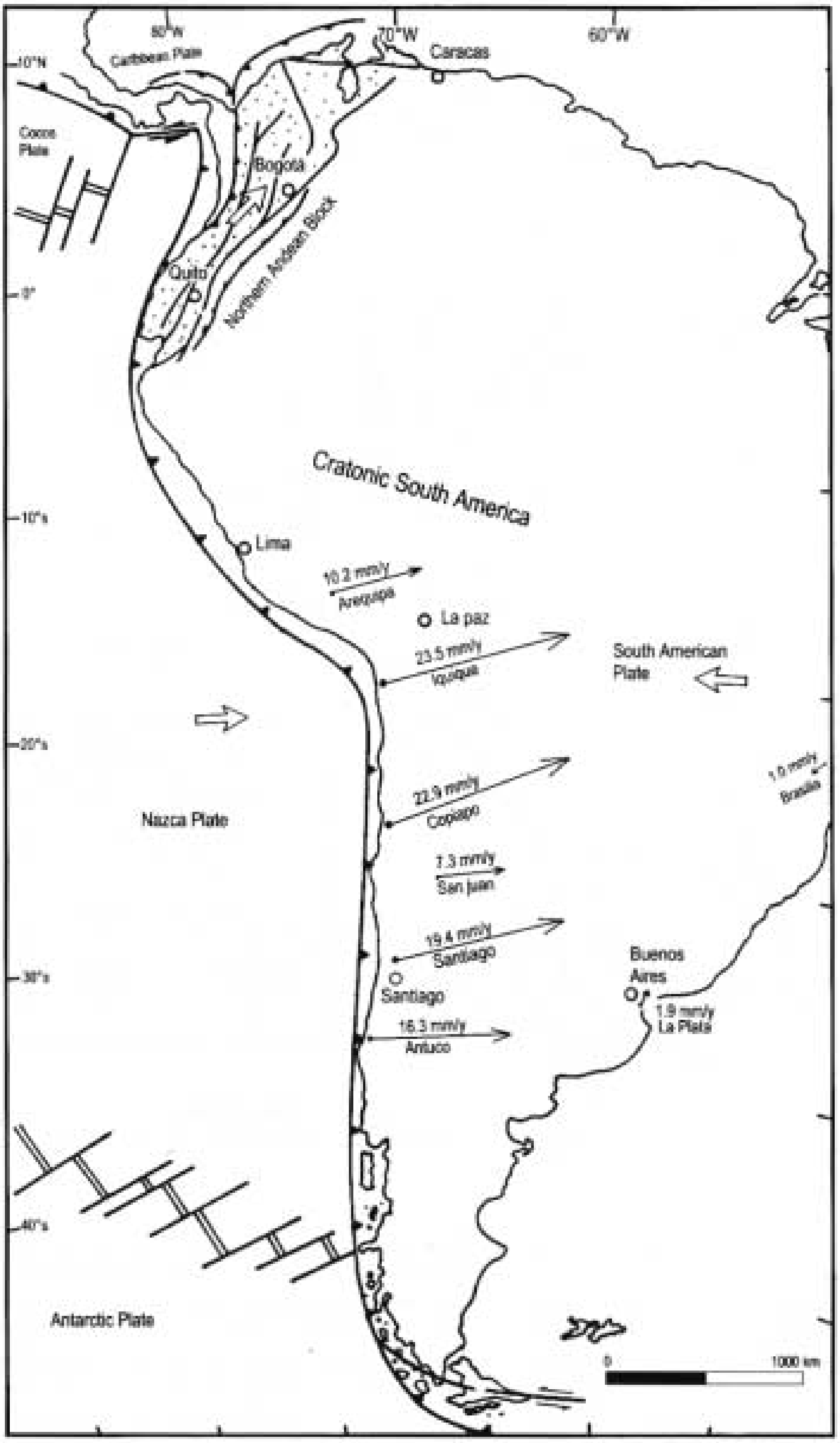


FIGURE 31: Space-based geodetic measurements along the Pacific border of the southern Central Andes in respect to stable South America. The differences between Santiago and the San Juan stations may account for the present shortening of the Andes, and between San Juan and La Plata for the Sierras Pampeanas shortening (modified after Kendrick et al., 1999).

where ridge collision is associated with Quaternary faults (Ramos, 1989b). Farther S, active strike-slip faulting controls the onset of intraplate basaltic flows along the northern part of the Magallanes Strait.

New space geodetic data record rates and direction of motion across the Andes, mainly between the continental margin affected by the Nazca Plate convergence and stable South America (Norabuena *et al.*, 1998). Recent data presented by Kendrick *et al.* (1999) and illustrated here in Figure 31 show significant shortening rates between the western slope of the Andes and the average position of stations situated in cratonic stable areas of Argentina and Brazil. The total relative motion is the result of several components such as (1) transient elastic deformation on the lock portion of plate interface, that can be released during large thrust earthquakes; and (2) permanent deformation through crustal shortening and mountain uplift. In the assessment of this permanent deformation it is important to weight the Andean shortening. The amount of shortening between Santiago de Chile (19.4 mm/y), San Juan (7.3 mm/y) and La Plata (1.9 mm/y) permits the evaluation of the active shortening within the Main Andes-Precordillera and Sierras Pampeanas in the Pampean flat-slab of Argentina and Chile. These figures indicate a shortening between both slopes of the Andes of 12 mm/y, and within Sierras Pampeanas of 5.4 mm/y. If these figures are compared with those derived from crustal balance of Andean roots (7.65 mm/y) or from structural cross sections (5.25 mm/y) in the Main Andes at these latitudes, the G.P.S. results are higher. This fact may indicate either a concentration of elastic deformation along the continental margin or an increase in recent years of the average Neogene shortening. More suitable figures are obtained when the structural shortening computed for Sierras Pampeanas, mainly the Pie de Palo area (5 mm/y in the last 3 Ma), one of the most active areas, is compared with the G.P.S. data (5.4 mm/y). The similar values may indicate that elastic deformation accumulated in this region is minimum, and if it existed, it was released by the large Pie de Palo earthquake in 1977 (Smalley *et al.*, 1993).

Although these values are still preliminary, they illustrate that space-based geodesy is opening a new era for studies of plate convergence along the Andes.

Concluding remarks

The previous analyses of the formation of the Andes and the description of their present tectonic setting show some contrasting characteristics among the different segments.

Early in the evolution of the proto-Andean margin of western Gondwana during Early Paleozoic times a great variety of basement blocks are observed, presently hidden under a thick Andean cover. Some of them are para-autochthonous basement blocks derived from Gondwana, and therefore considered by different authors as peri-Gondwanan terranes. On the other hand, some of the blocks have a distinct basement in comparison with western Gondwana, as indicated by isotope analyses, paleontological evidence, paleomagnetic data, and paleoclimatic conditions. These exotic blocks, such as the Cuyania Terrane, have well defined Laurentian affinities, and therefore are interpreted as allochthonous terranes accreted to the proto-margin of Gondwana during the Ocloyic and Chanic orogenies. Prior to the final amalgamation of these peri-Gondwanan and Laurentian blocks, a proto-Andean margin stage has been recognized. The Sierras Pampeanas basement records the best evidence of an Early Paleozoic magmatic arc, with time constraints that show a parallel evolution to the sedimentary record in the approaching Cuyania Terrane.

An intriguing characteristic of these basement terranes is the dominant Grenville age signature that almost all of them share. This is being interpreted as a confirmation of the common source in the Rodinia Supercontinent, and their participation in the Grenville Orogeny that may have amalgamated this supercontinent in middle Proterozoic times.

The nature of the different Paleozoic orogenies varies from N to S. The Northern Andes are the result of terrane accretion during Early Paleozoic times, that ended with a continent-to-continent collision between Laurentia and Gondwana to form the Alleghanides in Late Paleozoic times. The Southern Andes resulted from the collage of peri-Gondwanan and exotic Laurentian terranes amalgamated in Early Paleozoic times, without continent-to-continent collision. There is good evidence to show that the successive proto-margins were always facing to an open ocean, after docking a series of terranes. The Gondwanides mountain chain was the result of a Late Paleozoic orogeny that occurred after the final amalgamation of these terranes to the southwestern margin of Gondwana in Devonian times.

There is partial evidence that a magmatic arc occurred along the present continental margin during Late Paleozoic times. However, by the end of the Permian subduction there was a lull, and a well-developed generalized extension that predates the final break-up of the Pangea Supercontinent. The distribution of the rift systems and the associated magmatism shows that the extension was mainly concentrated in the hanging-wall of the suture between large cratonic basement terranes.

The opening of the North and South Atlantic seas induced another period of renewed subduction, but in the early stages linked to extension. Negative trench rollback velocities are in agreement with the northeastern true polar wandering of South America still connected with Africa. A reorganization of the stress state in the Andes produced a shift to compression that occurred between 115 and 105 Ma, and which is generally explained as the result of a change in the absolute motion of South America. The drift stage was completely obtained after the final break-up of South America and Africa at about 80 Ma. Oblique convergence vectors, regulated by the local orientation of the continental margin controlled strike-slip faulting in the fore arc.

Accretion of oceanic terranes in Colombia and Venezuela, most of them associated with oceanic plateaux derived from the Caribbean Plate, was the result of the interaction of this plate with South America. Island arc terranes have been accreted in Ecuador. These accretions took place in three different stages: in Early Cretaceous, Late Cretaceous to Paleogene, and in middle Miocene times. Obduction of oceanic basement, associated metamorphism and deformation characterized these stages.

The Incaic late Eocene-early Oligocene orogeny was important in some segments of the Andes, such as the Peruvian and Fueguian margins, with more orthogonal orientation to the slip vector of the oceanic plates.

The break-up of the Farallon oceanic plate in the late Oligocene-early Miocene, and the formation of the Cocos and Nazca plates was another major reorganization in the tectonics of the Andes, that led to the present plate tectonic setting of the different Andean active margin segments.

Although in most cases the Andean uplift was controlled by these oceanic factors, it is possible that uplift may also have resulted from the tectonic inversion of a complex extensional setting not directly related to subduction and arc magmatism as in the case of the Mérida Andes. When analyzed through time the stress regime in normal Andean-type segments shows:

1) The magmatic arc was either stable or oscillatory within few kilometres. Deformation in these cases shows minor shortening.

2) The magmatic arc and deformation migrated tens of kilometres following tectonic erosion of the continental margin.

3) The magmatic arc and deformation steadily migrated to the foreland in function of the changes in the Wadati-Benioff geometry.

4) The magmatic arc expanded and migrated to the foreland until cessation of magmatism and the development of a flat-slab setting. This produced strong deformation with foreland basement block-faulting in mature settings.

5) The magmatic arc is associated with near-trench magmatism, followed by strong deformation, and the development of retro-arc basaltic plateaux. Arc magmatism resumed after few millions years with an adakitic character.

6) The magmatic arc is retreated to the trench development in an extensional regime during the steepening of the angle of subduction, locally accompanied by the development of huge calderas with acid rocks, or extensive basaltic flows.

These variations are closely related to the interaction with the adjacent oceanic plates, the age of the oceanic crust being subducted, the collision of aseismic and seismic ridges, as well as the convergence vectors. Basement fabrics and earlier local geological history impose additional constraints on the resulting deformation.

There is general agreement that shallowing of the subduction zone is related to an increase in deformation and shifting of the magmatic arc. However, the effects of the steepening of the subduction zone are not well established. The trenchward migration of the magmatic arc is one of the first assumed pieces of evidence for this steepening. In some cases this is closely linked with extension and basaltic magmatism on a vast scale in the retro-arc settings. In other cases, there is no evidence for extension and we have generalized ignimbritic flare-up in huge calderas.

The space-based geodetic observations are permitting the evaluation of instantaneous deformation and its comparison with long-term deformation rates. This is opening an exciting field in the understanding of the relationships between the Andean uplift and its interaction with adjacent oceanic plates.

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REFERENCES

Aceñolaza, F.G., 1992. El Sistema Ordovícico de Latinoamérica. In: Gutiérrez Marco, J.C., Saavedra, J., Rábano, I. (eds.) *Paleozoico inferior de Ibero-América*, Universidad de Extremadura, Mérida, 85-118.

Aguirre, L., 1992. Metamorphic pattern of the Cretaceous Celica Formation, SW Ecuador, and its geodynamic implications. *Tectonophysics*, 205, 223-237.

Aguirre, L., Offler, R., 1985. Burial metamorphism in the western Peruvian trough: its relation to Andean magmatism and tectonics. In: Pitcher, W.S., Atherton, M.P., Cobbing, E.J., Beckinsale, R.D. (eds.) *Magmatism at the Plate edge: the Peruvian Andes*. Blackie Halstead Press, London, 59-71.

Aguirre, L., Levi, B., Nyström, J.O., 1989. The link between metamorphism, volcanism and geotectonic setting during the evolution of the Andes. In: Daly, J.S., Cliff, R.A., Yardley, B.W.D. (eds.) *Evolution of metamorphic belts*. Geological Society of London Special Publication, 43, 223-232.

Alemán, A., Marksteiner, R.M., 1997. Petroleum Systems and Structural Styles in the Santiago Fold and Thrust Belt: A salt Related Orogenic Belt. *VI° Simposio Bolivariano, Bogotá, Petroleum Exploration in the Subandean Basins*, 2, 171-186.

Aleman, A., Ramos, V.A., 2000. The Northern Andes. In: Milani, E.J., Thomaz Filho, A. (eds.) *Tectonic evolution of South America*, 31° International Geological Congress (Río de Janeiro), this volume.

Allmendinger, R., Jordan, T., Palma, M., Ramos, V.A., 1982. Perfil estructural en la Puna Catamarqueña (25-27°S), Argentina. *V° Congreso Latinoamericano de Geología (Buenos Aires), Actas*, 1, 499-518.

Allmendinger, R.W., Figueroa, D., Snyder, D., Beer, J., Mpodozis, C., Isacks, B.L., 1990. Foreland shortening and crustal balancing in the Andes at 30° S latitude. *Tectonics*, 9(4), 789-809.

Allmendinger, R.W., Jordan, T.E, Kay, S.M., Isacks, B.L., 1997. The evolution of the Altiplano-Puna Plateau of the Central Andes. *Annual Reviews of Earth and Planetary Sciences*, 25, 139-174.

Alvarez-Marrón, J., McClay, K.R., Harambour, S., Rojas, L., Skarmeta, J., 1993. Geometry and evolution of the frontal part of the Magallanes foreland thrust and fold belt (Vicuña area), Tierra del Fuego, southern Chile. *American Association of Petroleum Geologists Bulletin*, 77(11), 1904-1921.

Aspden, J.A., Litherland, M., 1992. The geology and Mesozoic collisional history of the Cordillera Real, Ecuador. *Tectonophysics*, 205, 187-204.

Aspden, J.A., McCourt, W.J., 1986. Mesozoic oceanic terrane in the Central Andes of Colombia. *Geology*, 14, 415-418.

Astini, R.A., Thomas, W.A. 1999. Origin and evolution of the Precordillera terrane of western Argentina: A drifted Laurentian orphan. In: Ramos, V.A., Keppie, D. (eds) *Laurentia-Gondwana connections before Pangea*. Geological Society of America Special Paper, 336, 1-20.

Atherton, M.P., Sanderson, L.M., 1987. The Cordillera Blanca Batholith: a study of granite intrusions and the relation of crustal thickening to peraluminosity. *Geologische Rundschau*, 76, 213-232.

Atherton, M.P., Webb, S., 1989. Volcanic facies, structure and

geochemistry of the marginal basin rocks of Central Perú. *Journal of South American Earth Sciences*, 2, 241-261.
Atherton, M.P., Pitcher, W.S., Warden, V., 1983. The Mesozoic Marginal basin of Central Perú. *Nature*, 305, 303-305.
Auboin, J.A., Borrello, A.V., Cecione, G., Charrier, R., Chotin, P., Frutos, J., Thiele, R., Vicente, J.C., 1973. Esquisse paleogeographique et structurale des Andes Meridionales. *Revue de Géographie Physique et de Geologie Dynamique*, XV(1-2), 11-71.
Avila Salinas, W., 1991. Petrologic and Cenozoic evolution of the Cenozoic volcanism in the Bolivian Western Andes. In: Harmon, R.S., Rapela, C.W. (eds.) *Andean Magmatism and its Tectonic Setting*, Geological Society of America Special Paper, 265, 245-257.
Baby, P., Heráil, G., Salinas, R., Sempere, T., 1992. Geometric and cinematic evolution of passive roof duplexes deduced from cross section balancing: example from the foreland thrust system of the southern Bolivian Subandean zone. *Tectonics*, 11(3), 523-536.
Baby, P., Moretti, I., Guillier, B., Limachi, R. Méndez, E. Oller, J., Specht, M., 1995. Petroleum system of the northern and central Bolivian Subandean zone. In: Tankard, A.J., Suárez Soruco, R., Welsink, H.J., (eds.) *Petroleum Basins of South America*. American Association of Petroleum Geologists Memoir, 62, 445-458.
Bahlburg, H., 1990. The Ordovician basin in the Puna of NW Argentina and N Chile: geodynamic evolution from back-arc to foreland basin. *Geotektonische Forschungen*, 75, 1-107.
Bahlburg, H., Hervé, F., 1997. Geodynamic evolution and tectonostratigraphic terranes of Northwestern Argentina and Northern Chile. *Geological Society of America Bulletin*, 109(7), 869-884.
Barazangi, M., Isacks, B., 1976. Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America. *Geology*, 4, 686-692.
Barrero, D., 1979. Geology of the Western Cordillera, west of Buga and Roldanillo, Colombia. *Publicaciones Geológicas Especiales del Ingeominas*, 4, 1-75.
Bastías, H.E., Uliarte, E., Paredes, J.D., Sánches, A., Bastías, J.A., Ruzycki, L., Perucca, P., 1990. Neotectónica de la provincia de San Juan. In: Bordonaro, O. (ed.) *Geología y Recursos Naturales de la provincia de San Juan*. XI° Congreso Geológico Argentino, San Juan, Relatorio, 228-244.
Beck, M.E. Jr., 1998. On the mechanism of crustal block rotations in the Central Andes. *Tectonophysics*, 299, 75-92.
Beck, S.L., Zandt, G., Myers, S.C., Wallace, T.C., Silver, P.G., Drake, L., 1996. Crustal-thickness variations in the central Andes. *Geology*, 24(5), 407-410.
Beer, J.A., Allmendinger, R.W., Figueroa, D.E., Jordan, T., 1990. Seismic stratigraphy of a Neogene Piggyback basin, Argentina. *American Association of Petroleum Geologist Bulletin*, 74(8), 1183-1202.
Bellizzia, A., Pimentel, N. 1994. Terreno Mérida: un cinturón alóctono Herciniano en la Cordillera de Los Andes de Venezuela. *V Simposio Bolivariano Exploración Petrolera en las Cuencas Subandinas, Memoria*, 271-299.
Benavides-Cáceres, V., 1999. Orogenic evoluion of the Peruvian Andes: the Andean cycle. In: Skinner, B. (ed.) *Geology and Mineral Deposits of Central Andes*. Society of Economic Geology Special Publication, 7, 61-108.
Biddle, K.T., Uliana, M.A., Mitchum, R.M., Fitzgerald, M.G., Wright, R.C., 1986. The stratigraphic and structural evolution of the central eastern Magallanes basin, southern South America. In: Allen, P., Homewood, P. (eds.) *Foreland Basins*. International Association of Sedimentologists Special Publication, 8, 41-61.
Boltovskoy, E., 1991. Ihering's hypothesis in the light of foraminiferological data. *Lethaia*, 24, 191-198.
Bordonaro, O., 1992. El Cámbrico de Sudamérica. In: Gutiérrez Marco, J.C., Saavedra, J., Rábano, I. (eds.) *Paleozoico inferior de Ibero-América*. Universidad de Extremadura, Mérida, 69-84.
Bosch, M., Rodríguez, I., 1992. North Venezuelan collisional crustal block: the boundary between the Caribbean and Southamerican plates. *Journal of South American Earth Sciences*, 6(3), 133-144.
Bourgois, J., Toussaint, J. F., Gonzalez, H., Orrego, A., Azema, J., Calle, B., Desmet, A., Murcia, L. A., Pablo, A. P., Parra, E., Tournon, J., 1985. Les ophiolites des Andes de Colombia: évolution structurale et signification géodynamique. In: Mascle, A. (ed.) *Géodynamique des Caraibes*, 475-493, Technip, Paris.
Bourgois, J., Toussaint, J. F., Gonzalez, H., Azema, J., Calle, B., Desmet, A., Murcia, L. A., Acevedo, A. P., Parra, E., Tournon, J. 1987. Geological history of the Cretaceous ophiolitic complexes of northwestern South America (Colombian Andes). *Tectonophysics*, 143, 307-327.
Brito Neves, B.B., Cordani, U.G., 1991. Tectonic evolution of South America during the Late Proterozoic. *Precambrian Research*, 53, 23-40.
Brito Neves, B.B., Campos Neto, M.C., Fuck, R.A., 1999. From Rodinia to Western Gondwana: an approach to the Brasiliano-Pan African Cycle and orogenic collage. *Episodes*, 22(3), 155-166.
Burke, K., 1988. Tectonic evolution of the Caribbean. *Annual Reviews Earth Planetary Sciences*, 16, 201-230.
Cabassi, I., Cabassi, R., Introcaso, A., 1999. Los Andes peruanos y argentino-chilenos: un estudio cortical comparativo. *XIV° Congreso Geológico Argentino, Salta, Actas*, I, 295-297.
Cahill, T., Isacks, B.L., 1992. Seismicity and the shape of the subducted Nazca plate. *Journal of Geophysical Research*, 97, 17503-17529.
Campbell, C.J., 1968. The Santa Marta wrench fault of Colombia and its regional setting. *4th Caribbean Geology Conference, Transactions*, 247-261.
Cande, S.C., Leslie, R.B., 1986. Late Cenozoic Tectonic of the Southern Chile Trench. *Journal of Geophysical Research*, 91 B1, 471-496.
Cembrano, J., Hervé, F., and Lavenu, A., 1996. The Liquiñe Ofqui fault zone: a long-lived intra-arc fault system in southern Chile. *Tectonophysics*, 259, 55-66.
Cembrano, J., Schermer, E., Lavenu, A. Sanhueza, A., 2000. Contrasting nature of deformation along an intra-arc shear zone, the Liquiñe-Ofqui fault zone, southern Chilean Andes. *Tectonophysics*, 319(2), 129-149.
Charrier, R., 1979. El Triásico de Chile y regiones adyacentes de Argentina: una reconstrucción paleogeográfica y paleoclimática. *Comunicaciones*, Santiago de Chile, 26, 1-37.
Charrier, R., 1984. Areas subsidentes en el borde occidental de la cuenca de tras-arco jurásico-cretácica, Cordillera Principal Chilena entre 34° y 34°30'S. *IX° Congreso Geológico Argentino (S.C. Bariloche), Actas*, 2, 107-124.
Chmielowski, J., Zandt, G., Haberland, C., 1999. The Central ANdean Altiplano-Puna magma body. *Geophysical Research Letters*, 26(6), 783-786.
Coira, B.L., Davidson, J.D., Mpodozis, C., Ramos, V.A., 1982. Tectonic and magmatic evolution of the Andes of Northern Argentina and Chile. *Earth Science Review*, 18(3-4), 303-332.
Coira, B.L., Mahlburg Kay, S., Viramonte, J., 1993. Upper Cenozoic magmatic evolution of the Argentine Puna: A model for changing subduction geometry. *International Geology Review*, 35(8), 677-720.
Coira, B.L., Mahlburg Kay, S., Pérez, B., Woll, B., Hanning, M., Flores, P., 1999. Magmatic sources and tectonic setting of Gondwana margin Ordovician magmas, northern Puna of Argentina and

Chile. In: Ramos, V.A., Keppie, D. (eds.) *Laurentia Gondwana Connections before Pangea.* Geological Society of America Special Paper 336, 145-170.

Colletta, B., Hebrard, F., Letouzey, J., Werner, P., Rudkiewicz, J-L., 1990. Tectonic style and crustal structure of the Eastern Cordillera (Colombia) from a balanced cross-section. In: Letouzey, J. (ed.). *Petroleum and Tectonics in Mobile Belts.* Editions Technip, 81-100.

Colletta, B., Roure, F., de Toni, B., Loureiro, D., Passalacqua, H., Gou, Y., 1997. Tectonic inheritance, crustal architecture and contrasting structural styles in the Venezuela Andes. *Tectonics,* 16(5), 777-794.

Comínguez, A., Ramos, V.A., 1995. Geometry and seismic expression of the Cretaceous Salta rift system, northwestern Argentina. In: Tankard, A.J., Suárez Soruco, R., Welsink, H.J., (eds.) *Petroleum Basins of South America.* American Association of Petroleum Geologists Memoir, 62, 325-340.

Cooper, M.A., F.T. Addison, R. Alvarez, M. Coral, R.H. Graham, A.B. Hayward, S. Howe, J. Martínez, J. Naar, R. Peñas, A.J. Pulham, A. Taborda, 1995. Basin development and tectonic history of the Llanos basin, Eastern Cordillera, and Middle Magdalena valley, Colombia. *American Association of Petroleum Geologists Bulletin,* 79(10), 1421-1443.

Cornejo, P.C., Mpodozis, C., Kay, S.M., Tomlinson, A.J., 1994. Volcanismo bimodal potásico en regimen extensional del Cretácico superior-Eoceno en la región de El Salvador (26°-27°S), Chile. *7° Congreso Geológico Chileno (Concepción), Actas,* II, 1306-1310.

Cortés, J.M., 1993. El frente de corrimiento de la Cordillera Frontal y el extremo sur del valle de Uspallata, Mendoza. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas,* III, 168-178.

Corvalán, J., 1981. Paleotectonic Map of the Circumpacific region Southeast Quadrant. *Circumpacific Council for Energy and Mineral Resources, A.A.P.G.,* scale 1:10.000.000.

Cosma, L., Lapierre, H., Jaillard, E., Laubacher, G., Bosch, D., Desmet, A., Mamberti, M., Gabriele, P., 1998. Petrographie and geochimie de la Cordillere Occidentale du Nord de l'Equateur (0°30'): implications tectoniques. *Bulletin Societé Geologique de France,* 169, 739-751.

Costa, C.H., 1992. Neotectónica del sur de la Sierra de San Luis. Universidad Nacional de San Luis, Ph. D. Thesis (unpublished) 390 pp.

Cunningham, W.D., 1993. Strike-slip faults in the southernmost Andes and the development of the Patagonian orocline. *Tectonics,* 12(1), 169-186.

Dalla Salda, L., Francese, J., 1987. Las megaestructuras del Macizo y la Cordillera Norpatagónica argentina y la génesis de las cuencas volcano-sedimentarias terciarias. *Revista Geológica de Chile,* 31, 3-13.

Cristallini, E.O., Ramos, V.A., 2000. Thick-skinned and thin-skinned thrusting in La Ramada Fold and Thrust belt: crustal evolution of the High Andes of San Juan, Argentina (32°SL). *Tectonophysics,* 317, 205-235.

Cristallini, E., Comínguez, A.H., Ramos, V.A., 1997. Deep structure of the Metan - Guachipas region: Tectonic inversion in northwestern Argentina. *Journal of South American Earth Sciences,* 10(5-6), 403-421.

Daly, M., 1989. Correlations between Nazca/Farallon plate kinematics and forearc evolution in Ecuador. *Tectonics,* 8(4), 769-790.

Dalziel, I.W.D., 1981.- Back-arc extension in the southern Andes: a review and critical reapraisal. *Philosophycal Transactions Royal Society of London,* A 300, 319-335.

Dalziel, I.W.D., 1986. Collision and Cordilleran orogenesis: an Andean perspective. In: Coward, M.P., Riers, A.C. (eds.) *Collision Tectonics,* Geological Society Special Publication, 19, 389-404.

Dalziel, I.W.D., 1991. Pacific margins of Laurentia and East-Antarctica-Australia as a conjugate rift pair: evidence and implications for an Eocambrian supercontinent. *Geology,* 19(6), 598-601.

Dalziel, I.W.D., 1994. Precambrian Scotland as a Laurentia-Gondwana link: the origin and significance of cratonic promontories. *Geology,* 22(7), 589-592.

Dalziel, I.W.D., 1986. Collision and Cordilleran orogenesis: an Andean perspective. In: Coward, M.P., Riers, A.C. (eds.) *Collision Tectonics.* Geological Society Special Publication, 19, 389-404.

Dalziel, I.W.D., Brown, R.L., 1989. Tectonic denudation of the Darwin metamorphic core complex in the Andes of Tierra del Fuego, southernmost Chile: implications for Cordilleran orogenesis. *Geology,* 17(8), 699-703.

Dalziel, I.W.D., Forsythe, R.D., 1985. Andean evolution and the terrane concept. Tectonostratigraphic Terranes of the Circum-Pacific Region. *Earth Sciences Series,* 1, 565-581.

Dalziel, I.W.D, de Wit, M.F., Palmer, K.F., 1974. Fossil marginal basin in the southern Andes. *Nature,* 250(5464), 291-294.

Davidson, J.P., Harmon, R.S., Wörner, G., 1993. The source of Central Andean magmas: some considerations. In: Harmon, R.S., Rapela, C.W. (eds.) *Andean Magmatism and its tectonic setting,* Geological Society of America Special Paper, 265, 233-243.

Dengo, G., 1983. Mid-America: Tectonic setting for the Pacific margin from southern Mexico to northern Colombia. *Centro de Estudios Geológicos de América Central,* Guatemala, 1-56.

Dengo, C.A., Covey, M.C., 1993. Structure of the Eastern Cordillera of Colombia: Implications for traps and regional tectonics. *American Association of Petroleum Geologists Bulletin,* 77, 1315-1337.

De Silva, S.L., 1989. Altiplano-Puna volcanic complex of the Central Andes. *Geology,* 17, 1102-1106.

Dewey, J.F., Bird, J. 1970. Mountain belts and the new global tectonics. *Journal of Geophysical Research,* 75, 2625-2647.

Dewey, J.F., Lamb, S. H., 1992. Active tectonics of the Andes. *Tectonophysics,* 205, 79-95.

Díaz Martínez, E., 1996. Síntesis estratigráfica y geodinámica del Carbonífero de Bolivia. *XII° Congreso Geológico de Bolivia (Tarija), Memorias,* 1, 355-367.

Dickinson, W. 1974. Plate tectonics and sedimentation. In: Dickinson, W.R. (ed.) *Tectonics and sedimentation.* Society of Economic Paleontologists and Mineralogists Special Publication, 22, 1-27.

Diraison, M., Cobbold, P.R., Rossello, E.A., Amos, A.J., 1998. Neogene dextral transpression due to oblique convergence across the Andes of northwestern Patagonia, Argentina. *Journal of South America Earth Sciences,* 11(6), 519-532.

Dorbath, L., Dorbath, C., Jiménez, E., Rivera, L., 1991. Seismicity and tectonic deformation in the Eastern Cordillera and the sub-Andean zone of Central Perú. *Journal of South American Earth Sciences,* 4(1-2), 13-24.

Duque-Caro, H., 1990. The Choco block in the northwestern corner of South America: structural, tectonostratigraphic, and paleogeographic implications. *Journal of South American Earth Sciences,* 3(1), 71-84.

Du Toit, A.L., 1937. *Our wandering continents. An hypothesis of continental drifting.* Oliver & Boyd, 366 pp., London.

Ego, F., Sébrier, M., Lavenu, A., Yepes, H., Egues, A., 1996. Quaternary stress in the Northern Andes and the restraining bend model for the Ecuadorian Andes. *Tectonophysics,* 259(1-3), 101-116.

Engebretson, D.C., Cox, A., Gordon, G.R., 1985. Relative plate motions between oceanic and continental plates in the Pacific basin. *Geological Society of America Special Paper,* 206, 1-64.

Erlich, R.N., Macsotay, O., Nederbragt, A.J., Lorente, M.A., 1999. Palaeoecology, Palaeogeography and depositional environments of Upper Cretaceous rocks of western Venezuela. *Palaeogeography, Palaeoclimatology, Palaeoecology,* 153, 203-238.

Etayo-Serna, F., 1994. *Estudios geológicos del valle superior del Magdalena.* Universidad Nacional de Colombia, 1-266.

Etayo-Serna, F., Barrero, D., 1983. Mapa de terrenos geológicos de Colombia. *Instituto Nacional de Investigaciones Geológico-Mineras Publicaciones Especiales*, 14, 235 p.

Farrar, E., Clark, A.H., Kontak, D.J., Archibal, D.A., 1988. Zongo-San Gabán zone: Eocene foreland boundary of the Central Andean orogen, northwest Bolivia and southeast Peru. *Geology*, 16(1), 55-58.

Feininger, T., 1982. Glaucophane schists in the Andes of Jambaló, Colombia. *Canadian Mineralogist*, 20, 41-47.

Feininger, T., 1987. Allochthonous terranes in the Andes of Ecuador and northwestern Peru. *Canadian Journal of Earth Sciences*, 24, 266-278.

Fernández, A.E., Jordan, T.E., 1996. Analysis of controls on foreland basin stratigraphy using seismic and outcrop data: application to the Bermejo Basin, Central Andes, Argentina. *XIII° Congreso Geológico Argentino y III° Congreso de Exploración de Hidrocarburos (Buenos Aires), Actas*, 1, 373-384.

Fernández Seveso, F., Pérez, M.A., Brisson, I.E., Alvárez, L.A., 1993. Sequence stratigraphy and tectonic analysis of the Paganzo basin, Western Argentina. *XII° International Congress on the Carboniferous-Permian Systems (Buenos Aires), Comptes Rendus*, 2, 223-260.

Forero-Suarez, A., 1990. The basement of the Eastern Cordillera, Colombia: an allochthonous terrane in northwestern South America. *Journal of South American Earth Sciences*, 3, 141-151.

Froidevaux, C., Isacks, B., 1984. The mechanical state of the lithosphere in the Altiplano Puna segment of the Andes. *Earth and Science Planetary Letters*, 71, 305-314.

Galeazzi, J.S., 1996. Cuenca de Malvinas. In: Ramos, V.A., Turic, M.A. (eds.), *Geología y Recursos Naturales de la Plataforma Continental Argentina*. XIII° Congreso Geológico Argentino y III° Congreso de Exploración de Hidrocarburos, Mendoza, Relatorio 15, 273-309.

Gansser, A., 1973. Facts and theories on the Andes. *Journal geological Society London*, 129, 93-131.

Ghiglione, M., Olivero, E.B., Ramos, V.A. 1999. Estratos de crecimiento en el frente orogénico palógeno de la Isla Grande de Tierra del Fuego. *II° Congreso Latinoamericano de Sedimentología y VIII° Reunión Argentina de Sedimentología* (resúmenes), 81-82, Mar del Plata.

Giacosa, R., Heredia, N., 1999. La cuenca de antepaís terciaria asociada a la faja plegada y corrida de los Andes Patagónicos entre los 41° y 42°S, SO de Argentina. *Acta Geológica Hispana*, 32(1-2), 103-112.

Godoy, E., Yañez, G, Vera, E., 1999. Inversion of an Oligocene volcano-tectonic basin and uplifting of its superimposed Miocene magmatic arc in the Chilean Central Andes: first seismic and gravity evidences. *Tectonophysics*, 306(2), 217-336.

Gorring, M.L, Kay, S.M., Zeitler, P.K., Ramos, V.A., Rubiolo, D., Fernández, M.I., Panza, J.L., 1997. Neogene Patagonian plateau lavas: continental magmas associated with ridge collision at the Chile Triple Junction. *Tectonics*, 16(1), 1-17.

Grier, M.E., Dalmeyer, R.D., 1990. Age of the Payogastilla Group: implications for foreland basin development, NW Argentina. *Journal of South American Earth Sciences*, 3(4), 269-278.

Grocott, J., Brown, M., Dallmeyer, R.D., Taylor, G.K., Treloar, P.J., 1994. Mechanisms of continental growth in extensional arcs: an example from the Andean plate-boundary zone. *Geology*, 22(5), 391-394.

Groeber, P., 1938. *Mineralogía y Geología*. Espasa-Calpe Argentina, 1-492, Buenos Aires.

Gutscher, M.A., Malavieille, J., 1999. Style of upper plate deformation linked to interplate coupling over flat slab segments: evidence from the Andean Margin. In: McClay, K. (ed.) *Thrust Tectonics 99 (London), Abstracts*, 214-217.

Gutscher, M.A., Olivet, J.L., Aslanian, D., Eissen, J.P., Maury, R., 1999a. The "lost Inca Plateau": cause of flat subduction beneath Peru?. *Earth and Planetary Science Letters*, 171, 335-341.

Gutscher, R., Malavieille, J., Lallemend, S., Collot, J.Y., 1999b. Tectonic segmentation of the North Andean margin: impact of the Carnegie Ridge Collision. *Earth and Planetary Science Letters*, 168, 255-270.

Hall, M. L., Beate, B., 1991. El volcanismo plio-cuaternario en los Andes de Ecuador. *Boletín del Colegio de Geógrafos del Ecuador*, 4, 5-17.

Hall, M.L., Calle, 1982. Geochronological control for the main tectonic-magmatic events of Ecuador. *Earth Science Reviews*, 18, 215-239.

Hall, M.L., Wood, C.A., 1985. Volcano-tectonic segmentation of the Northern Andes. *Geology*, 13(2), 203-207.

Haq, B.U., Hardenbol, J., Vail, P.R., 1987. Chronology of fluctuating sea level since the Triassic. *Science*, 235, 1156-1166.

Harmon, R.S., Barreiro, B., Moorbath, S., Hoefs, J., Francis, P. W., Thorpe, R.S., Deruelle, B., McHugh, J., Viglino, J.A., 1984. Regional O-, Sr-, and Pb-isotope relationships in late Cenozoic calc-alkaline lavas of the Andean Cordillera. *Journal of the Geological Society*, 141(5), 803-822.

Harrington, H.J., Leanza, A.F., 1957. Ordovician trilobites of Argentina. *University of Kansas, Department of Geology, Special Publication*, 1, 1-276.

Hatcher, R.D., Thomas, W.A., Geiser, P.A., Snoke, A.W., Mosher, S., Wiltschko, D.V., 1989. Alleghanian orogen. In: Hatcher, R.D., Thomas, W.A., Viele, G.W. (eds.) *The Appalachian-Ouachita orogen in the United States*. The Geology of North America, F-2, 233-318.

Hawkesworth, C., Clarke, C., 1994. Partial melting in the Lower crust: new constraints on crustal contamination processes in the Central Andes. In: Reutter, K.J., Scheuber, E., Wigger, P.J. (eds.) *Tectonics of the Southern Central Andes, Structure and Evolution of an Active Continental Margin*. Berlin, Springer Verlag, 93-101.

Helwig, J., 1972. Stratigraphy, sedimentation, paleogeography and paleoclimates of Carboniferous ("Gondwana") and Permian of Bolivia. *American Association of Petroleum Geologists Bulletin*, 56, 108-133.

Henderson, W.G., 1979. Cretaceous to Eocene volcanic arc activity in the Andes of northern Ecuador. *Journal of the Geological Society London*, 136, 367-378.

Hérail, G., Soler, P., Bonhomme, M., Lizeca, J.L., 1993. Evolution géodynamique du contact Altiplano-Cordilliére Orientale au Nord d'Oruro (Bolivie): implications sur le déroulement de l'orogénése andine. *Compte Rendus Academie Sciences, Paris*, 317, 512-522.

Hervé, F., 1988. Late Paleozoic subduction and accretion in Southern Chile. *Episodes*, 11(3), 183-188.

Hervé, F., 1994. The Southern Andes between 39° and 44°S Latitude: the geological signature of a transpressive tectonic regime related to a magmatic arc. K.J. Reutter, E. Scheuber y P.J. Wigger *Tectonics of the Southern Central Andes, Structure and Evolution of an Active Continental Margin*, 249-261, Springer Verlag, Berlín.

Hervé, F., Nelson, E., Kawashita, K., Suárez, M., 1981. New isotopic ages and the timing of orogenic events in the Cordillera Darwin, southernmost Chilean Andes. *Earth Planetary Science Letters*, 55, 257-265.

Hildreth, W., Moorbath, S., 1988. Crustal contributions to arc magmatism in the Andes of Central Chile. *Contributions to Mineralogy and Petrology*, 98, 455-489.

Hoffman, P.F., 1991. Did the breakout of Laurentia turn Gondwanaland inside-out?. *Science*, 252, 1409-1412.

Hoorn, C., 1993. Marine incursions and the influence of Andean tectonics on the depositional history of northwestern Amazonia: results of a palynological study. *Palaeogeography,* *Palaeoclimatology, Palaeoecology*, 105, 267-309.

Hoorn, C., Guerrero, J., Sarmiento G.A., Lorente M.A., 1995. Andean tectonic as cause of changing drainage patterns in Miocene northern South America. *Geology*, 23(3), 237-240.

Hungerbuehler, D., Steinmann, M., Winkler, W., Seward, D., 1996. Neogene fission-track stratigraphy of the southern Ecuadorian basins: Implications for regional tectonic history. *3er International Symposium of Andean Geodynamics, Saint-Malo*, ORSTOM, 387-390.

Introcaso, A., Pacino, M.C., Fraga, H., 1992. Gravity, isostasy and Andean crustal shortening between latitudes 30 and 35° S. *Tectonophysics*, 205(1-3), 31-48.

Jaillard, E., Ordoñez, M., Benítez, S., Berrones, G., Jiménez, N., Montenegro, G., Zambrano, I., 1995. Basin development in an accretionary, oceanic floored forearc setting: southern Coastal Ecuador during Late Cretaceous to Late Eocene times. In: Tankard, A.J., Suárez S., R., Welsink, H.J. (eds.) *Petroleum Basins of South America*. American Association of Petroleum Geologists Memoir, 62, 615-631.

Jaillard, E., Laubacher, G., Bengtson, P., Dhondt, A.V., Bulot, L.G., 1999. Stratigraphy and evolution of the Cretaceous forearc Celica. Lancones basin of southern Ecuador. *Journal of South American Earth Sciences*, 12(1), 51-68.

Jaillard, E., Hérail, G., Díaz Martínez, E., Baby, P., 2000. Tectonic evolution of the Andes of Ecuador, Peru, Bolivia, and northernmost Chile (1°N - 23°S). In: Milani, E.J., Thomaz Filho, A. (eds.) *Tectonic evolution of South America*, 31° International Geological Congress (Río de Janeiro), this volume.

James, D.E., Sacks, I.S., 1999. Cenozoic formation of the Central Andes: a geophysical perspective. In: Skinner, B. (ed.) *Geology and Mineral Deposits of Central Andes*. Society of Economic Geology Special Publication, 7, 1-25.

Jordan, T.E., 1995. Retroarc foreland and related basins. In: Spera, C., Ingersoll, R. V. (eds.) *Tectonics of Sedimentary basins*. Blackwell Scientific, 331-362, Cambridge, Mass.

Jordan, T.E., B. Isacks, V.A. Ramos, R.W. Allmendinger, 1983. Mountain building in the Central Andes. *Episodes*, 1983(3), 20-26.

Jordan, T.E., P.B. Flemings, J.A. Beer, 1988. Dating thrust-fault activity by use of foreland-basin strata. In: Kleinspehn, K.L., Paola, C. (eds.) *New Perspectives in Basin Analysis*, 307-330, Springer-Verlag, New York.

Jordan, T., J.H. Reynolds, J.P. Erikson, 1997. Variability in age of initial shortening and uplift in the Central Andes, 16-33°30'S. In: W. Ruddiman (ed.) *Tectonic uplift and Climate change*, 41-61, Plenum Press, New York.

Kay, S.M., Abbruzzi, J.M., 1996. Magmatic evidence for Neogene lithospheric evolution of the Central Andean "flat-slab" between 30° and 32°S. *Tectonophysics*, 259, 15-28.

Kay, S. M., Coira, B., Viramonte, J., 1994. Young mafic backarc volcanic rocks as indicators of continental lithospheric delamination beneath the Argentina Puna Plateau, Central Andes. *Journal of Geophysical Research*, 99, 24323-24339.

Kay, S.M., Mpodozis, C., Coira, B., 1999. Neogene magmatism, tectonism, and mineral deposits of the Central Andes (22°S to 33°S). In: Skinner, B. (ed.) *Geology and Mineral Deposits of Central Andes*. Society of Economic Geology, Special Publication 7, 27-59.

Kay, S.M., Ramos, V.A., Márquez, M., 1993b. Evidence in Cerro Pampa volcanic rocks for slab-melting prior to ridge-trench collision in southern South America. *Journal of Geology*, 101, 703-714.

Kay, S.M., Ramos, V.A., Mpodozis, C., Sruoga, P., 1989. Late Paleozoic to Jurassic silicic magmatism at the Gondwanaland margin: analogy to the Middle Proterozoic in North America?. *Geology*, 17(4), 324-328.

Kay, S.M., Mpodozis, C., Ramos, V.A., Munizaga, F., 1991. Magma source variations for mid to late Tertiary volcanic rocks erupted over a shallowing subduction zone and through a thickening crust in the Main Andean Cordillera (28-33°S). In: Harmon, R.S., Rapela, C. (eds.) *Andean Magmatism and its Tectonic Setting*. Geological Society of America Special Paper, 265, 113-137.

Kay, S.M., Ardolino, A., Franchi, M., Ramos, V., 1993a. Origen de la Meseta de Somún Curá: distribución y geoquímica de sus rocas volcánicas máficas. *XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos (Mendoza), Actas*, 4, 236-248.

Keidel, J., 1916. La geología de las Sierras de la Provincia de Buenos Aires y sus relaciones con las montañas de Sudáfrica y Los Andes. *Ministerio de Agricultura de La Nación, Sección Geología, Mineralogía y Minería Anales*, XI(3), 1-78.

Keidel, J., 1917. Über das Alter, die Verbreitung und die gegen-seitigen *Beziehungen der verschiedenen tektonischen strukturen in den* argentinischen Gebirgen. *XII° Session du Congrés Géologique International (Prague) Compte Rendus*, 671-687.

Keidel, J., 1921. Sobre la distribución de los depósitos glaciares del Pérmico conocidos en la Argentina y su significación para la estratigrafía de la serie del Gondwana y la paleogeografía del Hemisferio Austral. *Academia Nacional de Ciencias Boletín*, 25, 239-368.

Kellogg, J.N., Bonini, W.E., 1982. Subduction of the Caribbean plate and basement uplifts in the overriding South American plate. *Tectonics*, 1, 251-276.

Kellogg, J.N., Bonini, W.E., 1985. Reply to C. Schubert. *Tectonics* 4(7), 785-790.

Kendrick, E.C., Bevis, M., Smalley Jr. R.F., Cifuentes, O., Galban, F., 1999. Current rates of convergence acrpss the Central Andes: estimates fron continuous GPS observations. *Geophysical Research Letters*, 26(5), 541-544.

Keppie, J.D., Ortega-Gutiérrez, F., 1999. Middle America Precambrian basement: a missing piece of the reconstructed 1-Ga orogen. In: Ramos, V.A., Keppie, D. (eds.) *Laurentia Gondwana Connections before Pangea*. Geological Society of America Special Paper, 336, 199-210.

Keppie, J.D., Ramos, V.A., 1999. Odyssey of terranes in the Iapetus and Rheic oceans during the Paleozoic. In: Ramos, V.A., Keppie, D. (eds.) *Laurentia Gondwana Connections before Pangea*. Geological Society of America Special Paper, 336, 267-276.

Kerr, A.C., Marriner, G.F., Tarney, J., Nivia, A., Saunders, A.D., Thirlwall, M.F., Sinton, C., 1997a. Cretaceous basaltic terranes in western Colombia: elemental, chronological and Sr Nd isotopic constraints on petrogenesis. *Journal of Petrology*, 38, 677-702.

Kerr, A.C., Tarney, J., Marriner, G.F., Nivia, A., Klaver, G., Saunders, A.D., 1997b. The geochemistry and tectonic setting of Late Cretaceous Caribbean and Colombian Volcanism. *Journal of South American Earth Sciences*, 307(1-2): 235-254.

Klepeis, K.A., 1994. Relationship between uplift of the metamorphic core of the southernmost Andes and shortening in the Magallanes foreland fold and thrust belt, Tiera del Fuego, Chile. *Tectonics*, 13(4), 882-904.

Klepeis, K.A., Austin, J.A., 1997. Contrasting styles of superposed deformation in the southernmost Andes. *Tectonics*, 16(5), 755-776.

Kley, J., Monaldi, C.R., Salfity, J.A., 1999. Along-strike segmentation of the Andean foreland: causes and consequences, *Tectonophysics*, 301, 75-94.

Kohn, M.J., Spear, F.S., Harrison, T.M., Dalziel, I.W.D., 1995. $^{40}Ar/^{39}Ar$ geochronology P-T-t paths from the Cordillera de Darwin metamorphic complex. *Journal of Metamorphic Petrology*, 13, 251-270.

Kokogián, D.A., Fernández Seveso, F., Mosquera, A. 1993. Las secuencias sedimentarias triásicas. In: Ramos, V.A. (ed.) *Geología y Recursos Naturales de Mendoza*. XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos, Relatorio, I(7), 65-78.

Kontak, D.J., Clark, A.H., Farrar, E., Strong, D.F., 1985. The rift associated

Permo-Triassic magmatism of the Eastern Cordillera: a precirsor of the Andean orogeny. In: Pitcher, W.S., Atherton, M.P., Cobbing, E.J., Beckinsale, R.D. (eds.) *Magmatism at the Plate edge: the Peruvian Andes.* Blackie Halstead Press, London, 36-44.

Kozlowski, E., Manceda, R., Ramos, V.A., 1993. Estructura. In: Ramos, V.A. (ed.), *Geología y Recursos Naturales de Mendoza*, XII° Congreso Geológico Argentino y II° Congreso de Exploración de Hidrocarburos, Mendoza, Relatorio 235-256.

Lapierre, H., Dupuis, V., Bosch, D., Polvé, M., Maury, R.C., Hernández, J., Monié P., Yéghicheyan, D., Jaillard, E., Tardy, M., Mercier de Lépinay, B., MambertiM., Desnet, A., Keller, F., Sénebier, F., 1999. Multiple plume events in the genesis of the peri-Caribbean Cretaceous Oceanic Plateau Province. *Journal of Geophysical Research* (in press).

Lavenu, A., Cembrano, J., Arancibia, G., Deruelle, B., López Escobar, L., Moreno, H., 1997. Neotectónica transpresiva dextral y volcanismo: Falla Iquiñe-Ofqui, Sur de Chile. *8° Congreso Geológico de Chile (Antofagasta), Actas*, 1, 129-133.

Lebrat, M., Mégard, F., Juteau, T., Calle, J. 1985. Pre-orogenic volcanic assemblages and structure in the Western Cordillera of Ecuador between 1°40'S and 2°20'S. *Geologische Rundschau*, 74, 343-351.

Legarreta, L., Uliana, M.A., 1996. The Jurassic succession in west-central Argentina: stratal patterns, sequences and paleogeographic evolution. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 120, 303-330.

Limarino, C.O., Fauqué, L., Net, L., 1999. Análisis tectosedimentario de las cuencas Paganzo y Río Blanco, Noroeste Argentino. *Primer Simposio Argentino del Paleozoico Superior. Ameghiniana*, 36(4-Suplemento), Resúmenes, 33R.

Litherland, M., Aspden, J.A., Jemielita, R.A., 1994. The metamorphic belts of Ecuador. *British Geological Survey Overseas Memoir*, 11, 147 p.

Llambías, E., Sato, A.M., 1995. El batolito de Colangüil: transición entre orogénesis y anorogénesis. *Asociación Geológica Argentina Revista*, 50(1-4), 111-131.

López, G., Hatzfeld, D., Madariaga, R., Barrientos, S., Campos, J., Lyon-Caen, H., Zollo, A., Giannacone, G., Kausel, E. 1997. Microsismicidad en la zona centro-sur de Chile. *7° Congreso Geológico de Chile (Antofagasta) Actas*, 3, 1771-1774.

López Escobar, L., Cembrano, J., Moreno, H., 1995. Geochemistry and tectonics of Chilean Southern Andes basaltic Quaternary volcanism (37°-46°S). *Revista Geológica de Chile*, 22, 219-234.

Lugo, J., Mann, P., 1995. Jurassic-Eocene tectonic evolution of Maracaibo basin, Venezuela. In: Tankard, A.J., Suárez S., R., Welsink, H.J. (eds.) *Petroleum Basins of South America.* American Association of Petroleum Geologists Memoir, 62, 699-725.

Macellari, C.E., 1988. Cretaceous paleogeography and depositional cycles of western South America. *Journal of South American Earth Sciences*, 1(4), 373-418.

Manceda, R., Figueroa, D., 1995. Inversion of the Mesozoic Neuquén rift in the Malargüe fold-thrust belt, Mendoza, Argentina. In: Tankard, A.J., Suárez, R. Welsink, H.J. (eds.) *Petroleum Basins of South America.* American Association of Petroleum Geologists Memoir, 62, 369-382.

Marechal, P., 1983. Les Temoins de Chaine Hercynienne dans le noyau ancien des Andes de Mérida (Vènèzuèla). Ph.D. Thesis, Université de Bretagne Occidentale, Brest, 176 p.

Marshall, L.G., Sempere, T., Gayet, M., 1993. The Petaca (Late Oligocene-Early Miocene) and Yecua (Late Miocene) formations of the Subandean-Chaco basin, Bolivia, and their tectonic significance. *Documents Laboratoire Géologique Lyon*, 125, 291-301.

Martínez, M.P., Gímenez, M.E., Introcaso, A., Robles, J.A., 1997. Excesos de espesores corticales y acortamientos andinos en tres secciones ubicadas en 36°, 37° y 39° de latitud sur. *8° Congreso Geológico Chileno (Antofagasta), Actas*, 1, 101-105.

Mathalone, J.M.P., Montoya, M., 1995. Petroleum geology of the Subandean basins of Peru. In: Tankard, A.J., Suárez, R., Welsink, H.J., (eds.) *Petroleum Basins of South America.* American Association of Petroleum Geologists Memoir, 62, 423-444.

McCourt, W.J., Feininger, T., 1984. High-pressure metamorphic rocks in the Central Cordillera of Colombia. *British Geological Survey Report Series*, 16(1), 28-35.

McCourt, W.J., Aspden, J.A., Brook, M., 1984. New geological and geochronological data for the Colombian Andes: Continental growth by multiple accretion. *Journal of the Geological Society of London*, 141, 831-845.

Mégard, F., 1987. Cordilleran Andes and Marginal Andes: a review of Andean geology north of the Arica elbow (18°S). In: Monger, J.M.H., Francheteau, J. (eds.) *Circum-Pacific orogenic belts and evolution of the Pacific Ocean basin.* International Lithosphere Program Contribution, Geodynamic Series, 18, 71-96.

Mégard, F., Dalmayrac, B., Laubacher, G., Marocco, R., Martínez, C., Paredes, J., Tomasi, P., 1971. Lachaine hercynienne au Perou et en Bolivie, premiers resultats. *Cahiers Orstom, Serie Geologie*, III(1), 5-44.

Méndez Fajury, R. A., 1989. Catálogo de volcanes activos de Colombia. *Boletín Geológico del Ingeominas*, 30(3), 1-75.

Meschede, M., Frisch, W., 1998. A plate tectonic model for the Mesozoic and early Cenozoic history of the Caribbean plate. *Tectonophysics*, 296(3-4), 269-292.

Millward, D., Marriner, G.F., Saunders, A.D., 1984. Cretaceous tholeiitic volcanic rocks from the Western Cordillera of Colombia. *Journal of the Geological Society of London*, 141, 847-860.

Mojica, J., Dorado, J., 1987. El Jurásico anterior a los movimientos intermálmicos en los Andes Colombianos. In: Volkheimer W. (ed.) *Bioestratigrafía de los sistemas regionales del Jurásico y Cretácico de América del Sur.* Comité Sudamericano del Jurásico y el Cretácico, 49-110, Mendoza.

Mojica, J., Villaroel, C., Cuerda, A., Alfaro, M., 1988. La fauna de graptolitos de la Formación Higado (Llanvirniano-?Landeiliano), Serranía de Las Minas, Valle Superior del Magdalena, Colombia. *V° Congreso Geológico Chileno (Antofagasta), Actas*, II, C189-C202.

Mojica, J., Kammer, A., Ujeta, G., 1996. El Jurásico del sector noroccidental de Suramérica y guía de la excursión al Valle Superior del Magdalena (Nov. 1-4/95), regiones de Payandé y Prado, Departamento del Tolima, Colombia. *Geología Colombiana*, 21, 3-40.

Mosquera, A., 1999. Evolución temporal de la deformación en las Sierras Subandinas Orientales. *IV° Congreso de Exploración y Desarrollo de Hidrocarburos (Mar del Plata), Actas*, II, 573-568.

Moya, M.C., 1988. Estratigrafía del Tremadociano en el tramo austral de la Cordillera Oriental Argentina. Universidad Nacional de Salta, Tesis Doctoral (unpublished), 223 pp.

Moya, M.C., Malanca, S., Hongn, F.D., Bahlburg, H., 1993. El Tremadoc temprano en la Puna Occidental Argentina. *XII° Congreso Geológico Argentino y II° Congreso Exploración de Hidrocarburos (Mendoza), Actas*, 2, 20-30.

Mpodozis, C., Allmendinger, R.W., 1993. Extensional tectonics, Cretaceous Andes, northern Chile (27°S). *Geological Society of America Bulletin*, 105, 1462-1477.

Mpodozis, C., Cornejo, P., 1997. El rift triásico-sinemuriano de Sierra Exploradora, Cordillera de Domeyko (25-26°S): asociación de facies y reconstrucción tectónica. *VIII° Congreso Geológico Chileno (Antofagasta), Actas*, I, 550-553.

Mpodozis, C., Forsythe, R., 1983. Stratigraphy and geochemistry of accreted fragments of the ancestral Pacific floor in Southern South America. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 41, 103-124.

Mpodozis, C., Kay, S.M. 1992. Late Paleozoic to Triassic evolution of the Gondwana margin: evidence from Chilean Frontal Cordilleran Batholiths (28°-31°S). *Geological Society of America Bulletin*, 104, 999-1014.

Mpodozis, C., Ramos, V.A. 1990. The Andes of Chile and Argentina. In: Ericksen, G.E., Cañas Pinochet, M.T., Reinemud, J.A. (eds.) *Geology of the Andes and its relation to Hydrocarbon and Mineral Resources*. Circumpacific Council for Energy and Mineral Resources, Earth Sciences Series, 11, 59-90.

Mpodozis, C., Tomlinson, A.J., Cornejo, P.C., 1994. Acerca del control estructural de intrusivos eocenos y pórfidos cupríferos en la región de Potrerillos-El Salvador. *7° Congreso Geológico Chileno (Concepción), Actas*, II, 1596-1600.

Mujica, H. J., Zorzín, V., 1996. Extensión oriental de la tectónica andina en la llanura salteña, Provincia de Salta. *XIII° Congreso Geológico Argentino y III° Congreso Exploración de Hidrocarburos, (Buenos Aires), Actas*, I, 385-398.

Muñoz, J., Stern, C., 1988. The Quaternary volcanic belt of the Southern continental margin of South America: transverse structural and petrochemical variations across the segment between 38°S and 39°S. *Journal of South American Earth Sciences*, 1(2), 147-162.

Muñoz, N., Charrier, R., Reutter, K-J. 1997. Evolución de la cuenca Salar de Atacama: inversión tectónica y relleno de una cuenca de antepaís de retroarco. *VIII° Congreso Geológico Chileno (Antofagasta), Actas*, I, 195-199.

Naranjo, J.A., Hervé, F., Prieto, X., Munizaga, F. 1984. Actividad cretácica de la falla Atacama al este de Chañaral: milonización y plutonismo. *Comunicaciones*, Santiago de Chile, 34, 37-45.

Nelson, E.P., 1982. Post-tectonic uplift of the Cordillera de Darwin orogenic core complex: evidence for fission track geochronology and closing temperature-time relationship. *Journal of the Geological Society of London*, 130, 755-762.

Norabuena, E., Leffler-Griffin, L., Mao, A., Dixon, T., Stein, S., Sacks, I. S., Ocola, L. Ellis, M., 1998. Space geodetic observations of Nazca-South America convergence across the Andes. *Science*, 279, 358-362.

Olivero, E.B., 1998. Mesozoic-Paleogene geology of the marginal-Austral basin of Tierra del Fuego. *Third Annual Conference, IGCP Project 381, Aouth Atlantic Mesozoic Correlations, Field Trip Guide*, 1-42.

Olivero, E.B., Malumián, N., 1999. Eocene stratigraphy of southeastern Tierra del Fuego island, Argentina. *American Association of Petroleum Geologists Bulletin*, 83(2), 295-313.

Omarini, R.H., Sureda, R.J., Götze, H.J., Seilacher, A., Plüger, F., 1999. Puncoviscana fold belt in Northwestern Argentina: testimony of Late Proterozoic Rodinia fragmentation and pre-Gondwana collisional episodes. *International Journal of Earth Sciences*, 88, 76-97.

Orrego, L.A., Cepeda, H.V., Rodríguez, S.G.I., 1980. Esquistos glaucofánicos en el área de Jambaló, Cauca, Colombia. *Geología Norandina*, 1, 5-10.

Palma, M.A., Parica, P., Ramos, V.A., 1987. El Granito Archibarca: su edad y significado tectónico. *Asociación Geológica Argentina Revista*, 41(3-4, 1986), 414-418.

Pankhurst, R., Rapela, C., Saavedra, J., Baldo, E., Dahlquist, J., Pascua, I., Fanning, C.M., 1998. The Famatinian magmatic arc in the Central Sierras Pampeanas: an Early to mid-Ordovician continental arc on the Gondwana margin. In: Pankhurst, R.J., Rapela, C.W., (eds.) *The Proto-Andean margin of Gondwana*. Geological Society of London Special Publication, 142, 343-367.

Parnaud, F., Gou, Y., Pascual, J.C., Capello, M.A., Truskowski, I., Passalacqua, H., 1995. Stratigraphic synthesis of western Venezuela. In: Tankard, A.J., Suárez S., R., Welsink, H.J. (eds.) *Petroleum Basins of South America*. American Association of Petroleum Geologists Memoir, 62, 681-698.

Pardo Casas, F., Molnar, P., 1987. Relative motion of the Nazca (Farallon) and South American Plates since Late Cretaceous time. *Tectonics*, 6(3), 233-248.

Pennington, W.D., 1981. Subduction of the Eastern Panama Basin and seismotectonics of Northwest South America. *Journal of Geophysical Research*, 86, 10753-10770.

Pérez, D.J., Ottone, G., Ramos, V.A., 1996. La ingresión marina miocena en la provincia de San Juan: sus implicancias paleogeográficas. *XIII° Congreso Geológico Argentino y III° Congreso Exploración de Hidrocarburos (Buenos Aires), Actas*, I, 385-398.

Pilger, R.H., 1984. Cenozoic plate kinematics, subduction and magmatism: South American Andes. *Journal Geological Society of London*, 141, 793-802.

Pindell, J.L., 1985. Alleghenian reconstruction and subsequent evolution of the Gulf of Mexico, Bahamas, and Proto-Caribbean. *Tectonics*, 4(1), 1-39.

Pindell, J.L., Barrett, S.F., 1990. Geological evolution of the Caribbean region: a plate-tectonic perspective. In: Dengo, G. and Case, J.E. (eds.). *The Caribbean Region*. The Geology of North America. Geological Society of America, H, Chapter 16, 405-432.

Pitcher, W.S., Atherton, M.P., Cobbing, E.J., Beckinsale, R.D., 1985. *Magmatism at a Plate Edge: The Peruvian Andes*. 1-289 p., Blackie Halstead Press, London.

Quenardelle, S., Ramos, V.A., 1999. The Ordovician western Sierras Pampeanas magmatic belt: record of Precordillera accretion in Argentina. In: Ramos, V.A., Keppie, D., (eds.) *Laurentia Gondwana Connections before Pangea*. Geological Society of America Special Paper, 336, 63-86.

Ramos, V.A., 1970. Estratigrafía y estructura de la Sierra de Los Colorados, provincia de La Rioja. *Asociación Geológica Argentina Revista*, XXV(3), 359-382.

Ramos, V.A., 1982. Las ingresiones pacíficas del Terciario en el Norte de la Patagonia. *III° Congreso Geológico Chileno (Concepción), Actas*, I(A), 262-288.

Ramos, V.A., 1983. Evolución tectónica y metalogénesis de la Cordillera Patagónica. *II° Congreso Nacional Geología Económica (San Juan), Actas*, 1, 108-124.

Ramos, V.A., 1985. El Mesozoico de la Alta Cordillera de Mendoza: reconstrucción tectónica de sus facies. Argentina. *IV° Congreso Geológico Chileno (Antofagasta), Actas*, I(2), 104-118.

Ramos, V.A., 1986. El diastrofismo oclóyico: un ejemplo de tectónica de colisión durante el Eopaleozoico en el noroeste Argentino. *Revista Instituto Ciencias Geológicas*, Jujuy, 6, 13-28.

Ramos, V.A., 1988a. Tectonics of the Late Proterozoic-Early Paleozoic: a collisional history of southern South America. *Episodes*, 11(3), 168-174.

Ramos, V.A., 1988b. The tectonics of the Central Andes: 30° to 33° S latitude. In: Clark, S., Burchfiel, D. (eds.) *Processes in Continental Lithospheric Deformation*. Geological Society of America Special Paper, 218, 31-54.

Ramos, V.A., 1989a. The birth of Southern South America. *American Scientist*, 77(5), 444-450.

Ramos, V.A., 1989b. Foothills structure in Northern Magallanes Basin, Argentina. *American Association Petroleum Geologists Bulletin*, 73(7), 887-903.

Ramos, V.A., 1999. Plate tectonic setting of the Andean Cordillera. *Episodes*, 22(3), 183-190.

Ramos, V.A., Aguirre-Urreta, M.B., 1994. Cretaceous evolution of the Magallanes basin. In: Salfity, J.A., (ed.) *Cretaceous Tectonics of the Andes*. Earth Evolution Series, 315-345, Fried. Vieweg & Sohn, Braunschweig/Wiesbaden.

Ramos, V.A., Aguirre-Urreta, M.B., 2000. Patagonia. In: Milani, E.J., Thomaz Filho, A. (eds.) *Tectonic evolution of South America*. 31° International Geological Congress (Rio de Janeiro), this volume.

Ramos V.A., Alonso, R.N., 1995. El mar paranense en la provincia de Jujuy. *Revista Instituto de Geología y Minería*, Jujuy, 10, 73-82.

Ramos, V.A., Barbieri, M., 1989. El volcanismo Cenozoico de Huantraico: edad y relacionesisotopicas iniciales, provincia del Neuquen.

Asociación Geológica Argentina Revista, XLIII(2), 210-223.

Ramos, V.A., Basei, M., 1997a. The basement of Chilenia: an exotic continental terrane to Gondwana during the Early Paleozoic. *Symposium on Terrane Dynamics' 97 (Christchurch)*, 140-143.

Ramos, V.A., Basei, M., 1997b. Gondwanan, Perigondwanan, and exotic terranes of southern South America. *South American Symposium on Isotope Geology (Sao Paulo), Abstracts*, 250-252.

Ramos, V.A., Kay, S.M., 1991. Triassic rifting and associated basalts in the Cuyo basin, central Argentina. In: Harmon, R.S., Rapela, C.W. (eds.) *Andean Magmatism and its Tectonic Setting*. Geological Society of America Special Paper, 265, 79-91.

Ramos, V.A., Kay, S.M., 1992. The Southern Patagonian plateau basalts: retroarc testimony of a ridge collision, Argentina. *Tectonophysics*, 205, 261-282.

Ramos, V.A., Munizaga, F., Kay, S.M., 1991. El magmatismo cenozoico a los 33°S de latitud: geocronología y relaciones tectónicas. *VI° Congreso Geológico Chileno (Viña del Mar), Actas*, 1, 892-896.

Ramos, V.A., Cegarra, M., Cristallini, E., 1996a. Cenozoic tectonics of the High Andes of west-central Argentina, (30°- 36°S latitude). *Tectonophysics*, 259, 185-200.

Ramos, V.A., Niemeyer, H., Skarmeta, J., Muñoz, J., 1982. The magmatic evolution of the Austral Patagonian Andes. *Earth Science Reviews*, 18(3-4), 411-443.

Ramos, V.A., Jordan, T.E., Allmendinger, R.W., Mpodozis, C., Kay, S., Cortés, J.M., Palma, M.A. 1986. Paleozoic terranes of the Central Argentine-Chilean Andes. *Tectonics*, 5(6), 855-880.

Ramos, V.A., Aguirre Urreta, M.B., Alvarez, P.P., Cegarra, M., Cristallini, E.O., Kay, S.M., Lo Forte, G.L., Pereyra, F., Pérez, D., 1996b. *Geología de la Región del Aconcagua, Provincias de San Juan y Mendoza*. Dirección Nacional del Servicio Geológico Anales, 24, 1-510.

Rapalini, A.E., Vilas, J.F., 1991. Preliminary paleomagnetic data from the Sierra Grande Formation: Tectonic consequences of the first mid-Paleozoic paleopoles from Patagonia. *Journal of South American Earth Sciences*, 4, 25-41.

Rapalini, A.E., Vizán, H., 1993. Evidence of intrapangea movements in Gondwanaland. *XII° International Congress on Carboniferous - Permian systems (Buenos Aires), Comptes Rendus*, 1, 405-434.

Rapalini, A.E., Astini, R.A., Conti, C.M., 1999. Paleomagnetic constraints on the evolution of Paleozoic suspect terranes from South America. In: Ramos, V.A., Keppie, D., (eds.) *Laurentia Gondwana Connections before Pangea*. Geological Society of America Special Paper, 336, 171-182.

Rapela, C.W., Pankhurst, R.J., Dahlquist, J., Fanning, C.M., 1999. U-Pb Shrimp ages of Famatinian granites: new constraints on the timing, origin and tectonic setting of I- and S-type magmas in an ensialic arc. *II° South America Symposium on Isotope Geology (Córdoba), Actas*, 264-267.

Räsänen, M.E., Linna, A.M., Santos, J.C.R., Negri, F., 1995. Late Miocene tidal deposits in the Amazonian foreland basin. *Science*, 269, 386-390.

Restrepo, J.J., Toussaint, J.F., 1973). Obducción cretácea en el occidente Colombiano. *Publicaciones Especiales de Geología, Universidad Nacional, Facultad de Minas*, Medellin, 3, 1-26.

Restrepo, J.J., Toussaint, J-F., 1988. Terranes and continental accretion in the Colombian Andes. *Episodes*, 11, 189-193.

Restrepo-Pace, P.A., 1992. Petrotectonic characterization of the Central Andean Terrane, Colombia. *Journal of South American Earth Sciences*, 5, 97-116.

Restrepo-Pace, P.A., 1999. Fold and thrust belt along the western flankof the Eastern Cordillera of Colobia: style, Kinematics and timing constraints derived from seismic data and detailed surface mapping. In: McClay, K. (ed.) *Thrust Tectonics*, 112-114.

Restrepo-Pace, P.A., Ruiz, J., Gehrels, G., Cosca, M., 1997. Geochronology and Nd isotopic data of Grenville-age rocks in the Colombian Andes: new constraints for Late Proterozoic-Early Paleozoic paleocontinental reconstructions of the Americas. *Earth and Planetary Science Letters*, 150, 427-441.

Reynaud, C., Jaillard, E., Lapiere, H., Mamberti, M., Mascle, G., 1999. Oceanic plateau and island arcs of southwestern Ecuador: their place in the geodynamic evolution of northwestern South America. *Tectonophysics*, 307, 235-254.

Robin, C., Hall, M., Jimenez, M., Monzier, M., Escobar, P., 1997, Mojanda volcanic examples (Ecuador): development of two adjacent contemporaneous volcanoes with contrasting eruptive styles and magmatic suites. *Journal of South American Earth Sciences*, 10(5-6), 345-359.

Robison, R., Pantoja-Alor, J. 1968. Tremadocian trilobites from the Nochixtlán region, Oaxaca, Mexico. *Journal of Paleontology*, 42(3), 767-800.

Rossello, E., Mozetic, M.E., 1999. Caracterización estructural y significado geotectónico de los depocentros cretácicos continentales del centro-oeste argentino. *5° Simposio sobre o Cretáceo do Brasil (Rio Claro), Boletim*, 107-113.

Ruiz, T., Tosdal, R.M., Restrepo, P.A., Murillo-Muñetón, G., 1999. Pb Isotope evidence for Colombia-southern Mexico connections in the Proterozoic. In: Ramos, V.A., Keppie, D., (eds.) *Laurentia Gondwana Connections before Pangea*. Geological Society of America Special Paper, 336, 183-198.

Russo, R.M., Silver, P.G., 1996. Cordillera formation, mantle dynamics, and the Wilson cycle. *Geology*, 24(6), 511-514.

Rutland, R.W.R., 1971. Andean orogeny and sea floor spreading. *Nature*, 233, 252-255.

Sadowski, G.R., Bettencourt, J.S., 1994. Amazon and Laurentia, former neighbours. *38°Congresso Brasileiro de Geologia (Camboriú), Resumos*, 110.

Salfity, J., 1982. Evolución paleogeográfica del Grupo Salta (Cretácico-Eogénico), Argentina. *V° Congreso Latinoamericano de Geología (Buenos Aires), Actas*, I, 11-26.

Salfity, J.A. (ed.), 1994. *Cretaceous Tectonics of the Andes*. Earth Evolution Series, 1-345, Fried. Vieweg & Sohn, Braunschweig/ Wiesbaden.

Salfity, J.A., Marquillas, R.A., 1994. Tectonic and sedimentary evolution of the Cretaceous-Eocene Salta Group basin, Argentina. In: Salfity, J.A. (ed.) *Cretaceous Tectonics of the Andes, Earth Evolution Series, 266-315, Fried. Vieweg & Sohn*, Braunschweig/Wiesbaden.

Sandeman, H.A., Clark, A.H., Farrar, E., 1995. An integrated tectono-magmatic model for the evolution of the Southern Peruvian Andes (13°-20°S) since 55 Ma. *International Geology Review*, 37, 1039-1073.

Sanjines-Saucedo, G., 1982. Estratigrafía del Carbónico, Triásico y Cretácico Boliviano en el borde oriental de las Sierras Subandinas Bolivianas. *5° Congreso Latinoamericano de Geología (Buenos Aires), Actas*, 1, 301-318.

Scheuber, E., Bogdanic, T., Jensen, A., Reutter, K.-J., 1994. Tectonic development of the North Chilean Andes in relation to plate convergence and magmatism since the Jurassic. In: Reutter, K.-J.-. Scheuber, E., Wigger, P.J. (eds.) *Tectonics of the Southern Central Andes, Structure and evolution of an active continental margin*, 121-139, Springer-Verlag.

Schmitz, M., 1994, A balanced model of the southern Central Andes. *Tectonics*, 13(2), 484-492.

Schubert, C., 1982. Neotectonics of Boconó fault, Western Venezuela. *Tectonophysics*, 85, 205-220.

Schubert, C., Vivas, L., 1993. *El Cuaternario de la Cordillera de Mérida, Andes Venezolanos*. Universidad de Los Andes, Fundación Polar, Mérida, 1-345 p.

Sébrier, M., Soler, P., 1991. Tectonics and magmatism in the Peruvian

Andes from Late Oligocene time to the present. In: Harmon, R.S., Rapela, C. (eds.) *Andean Magmatism and its Tectonic Setting.* Geological Society of America Special Paper 265, 259-276.

Sébrier, M., J.L. Mercier, F. Mégrad, G. Laubacher, E. Carey-Gaihardis, 1985. Quaternary normal and reverse faulting and the state of stress in the central Andes of south Peru. *Tectonics*, 7, 739-780.

Sébrier, M., Mercier, J.L., Macharé, J., Bonnot, D., Cabrera, J., Blanc, J.L., 1988, The state of stress in an overriding plate situated above a flat slab: The Andes of central Peru. *Tectonics*, 7, 895-928.

Sempere, T., 1995. Phanerozoic evolution of Bolivia and adjacent regions. In: Tankard, A.J., Suárez, R., Welsink, H.J. (eds.) *Petroleum Basins of South America.* American Association of Petroleum Geologists Memoir, 62, 207-230.

Sempere, T., Hérail, G., Oller, J., Bonhomme, M., 1990. Late Oligocene-early Miocene major tectonic crisis and related basins in Bolivia. *Geology,* 18(10), 946-949.

Smalley, R.F., Isacks, B.L., 1990. Seismotectonics of thin and thick-skinned deformation in the Andean foreland from local network data: evidence for a seismogenic lower crust. *Journal of Geophysical Research*, 95, 12487-12498.

Smalley, R., Pujol, J., Regnier, M., Ming Chiu, J., Chatelain, J. L., Isacks, B., Araujo, M. and Puebla, N., 1993. Basement seismicity beneath the Andean Precordillera thin-skinned thrust belt and implications for crustal and lithospheric behaviour. *Tectonics*, 12(1), 63-76.

Soler, P., Bonhomme, M., 1990. Relations of magmatic activity to plate dynamics in Central Peru from Late Cretaceous to present. In: Kay, S., Rapela, C.W. (eds.) *Plutonism from Antarctica to Alaska.* Geological Society of America Special Paper 241, 173-191.

Somoza, R., 1995. Paleomagnetismo de rocas cretácicas de la Patagonia y la curva de deriva polar aparente de América del Sur: implicancias geocinemáticas y tectónicas. PhD Thesis, Universidad de Buenos Aires (unpublished) 1-238 p., Buenos Aires.

Steinmann, M., Hungerbühler, Seward, D., Winkler, W., 1999. Neogene tectonic evolution and exhumation of the southern Ecuadorian Andes: a combined stratigraphiy and fission-track approach. *Tectonophysics*, 307, 255-276.

Stephan, J.F., 1982. Evolution geodynamique du domaine caraibe Andes et chaines caraibe sur la transversele de Barquisimeto, Venézuéla. Ph.D. Thesis, Université Pierre et Marie Curie, Paris, 1 and 2, 512 p.

Stern, C.R., 1990. Pliocene to present migration of the volcanic front, Andean Southern Volcanic zone. *Revista Geológica de Chile*, 16(2), 145-162.

Stern, Ch.R., 1991. Role of subduction erosion in the generation of Andean Magmas. *Geology*, 19(1), 78-81.

Stern, C., Kilian, R., 1996. Role of the subducted slab, mantle wedge and continental crust in the generation of adakites from the Andean Austral Volcanic Zone. *Contributions to Mineralogy and Petrology*, 123, 263-281.

Stern, Ch., Skewes, M.A., Durán, M., 1976a. Volcanismo orogénico en Chile Austral. *I° Congreso Geológico Chileno (Santiago), Actas*, II (F), 195-212.

Stern, Ch., de Wit, M.J., Lawrence, J.R., 1976b. Igneous and metamorphic processes associated with the formation of Chilean ophiolites and their implications for ocean floor metamorphism, seismic layering, and magmatism. *Journal of Geophysical Research*, 81, 4370-4380.

Suárez, G., Molnar, P., Burchfield, B. C., 1983. Seismicity, fault plane solutions, depth of faulting, and active tectonics of the Andes of Peru, Ecuador and southern Colombia. *Journal of Geophysical Research*, 88, 10403-10428.

Suárez, M., 1976. Plate tectonic model for Southern Antarctica Peninsula and its relation to Southern Andes. *Geology*, 4, 211-214.

Suárez, M., Bell, C.M., 1993. Triassic-rift related sedimentary basins in northern Chile (24°-29°S). *Journal of South American Earth Sciences*, 6(3), 109-121.

Suárez Soruco, R., 1999. Bolivia: Late Proterozoic - Early Paleozoic. *IV° International Symposium on Andean Geodynamics, (Goettingen), Proceedings*, 716-718.

Schwartz, D., 1988. Paleoseismicity and neotectonics of the Cordillera Blanca fault zone, Northern Peruvian Andes. *Journal of Geophysical Research*, 93, 4712-4730.

Taylor, G.K., Grocott, J., Pope, A., Randall, D.E., 1998. Mesozoic fault systems, deformation and fault block rotation in the Andean forearc: a crustal scale strike-slip duplex in the Coastal Cordillera of northern Chile. *Tectonophysics*, 299(1-3), 93-109.

Tebbens, S.F., Cande, S., Kovacs, L., Parra, J.C., LaBreque, J.L., Vergara, H., 1997. The Chile ridge: a tectonic framework. *Journal Geophysical Research*, 102(B6), 12035-12059.

Thorpe, R.S., 1984. The tectonic setting of active Andean volcanism. In: Harmon, R.S., Barreiro, B.A. (eds), *Andean Magmatism*. Cheshire, Shiva Pub. Ltd., 4-8.

Thorpe, R.S., Francis, P.W., 1979. Variations in Andean andesite compositions and their petrogenetic significance. *Tectonophysics*, 57, 53-70.

Tomlinson, A.J., Blanco, N., 1997. Structural evolution and displacement history of the west fault system, Precordillera, Chile: Part I: Synmineral history. *VIII° Congreso Geológico Chileno (Antofagasta), Actas*, 3, 1873-1877.

Tomlinson, A.J., Mpodozis, C., Cornejo, P., Fernández, C.F., Dumitru, T., 1994. El sistema de fallas Sierra Castillo-Agua Amarga: transpresión sinistral eocena en la Precordillera de Potrerillos-El Salvador. *7° Congreso Geológico Chileno (Concepción), Actas*, II, 1459-1463.

Toselli, A., Saavedra, J., Rossi de Toselli, J.N., 1996. Interpretación geotectónica del magmatismo del sistema de Famatina. In: Aceñolaza, F.G., Miller, H., Toselli, A.J. (eds.) *Geología del Sistema de Famatina*. Münchner Geologische Hefte, A 19, 283-292.

Toussaint, J-F., 1993. *Evolución Geológica de Colombia, Precámbrico-Paleozoico*. Universidad Nacional de Colombia, Medellín, 229 p.

Toussaint, J-F., Restrepo, J. J., 1982. Magmatic evolution of the northwestern Andes of Colombia. *Earth Science Reviews*, 18, 205-213.

Toussaint, J-F., Restrepo, J. J., 1994. The Colombian Andes during Cretaceous times. In: Salfity, J. (ed.) *Cretaceous Tectonics of the Andes*. Earth Evolution Sciences Monograph Series, 61-100, Friedr. Vieweg & Sohn.

Uliana, M.A., Biddle, K.T., 1988. Mesozoic-Cenozoic paleogeographic and geodynamic evolution of southern South America. *Revista Brasileira de Geociências*, 18, 172-190.

Uliana, M.A., Legarreta, L., 1993. Hydrocarbons habitat in a Triassic-to-Cretaceous sub-Andean setting: Neuquén basin, Argentina. *Journal of Petroleum Geologists*, 16(4), 397-420.

Uyeda, S. 1983. Comparative Subductology. *Episodes*, 1983(2), 19-24.

Van der Wiel, A. M., 1991, Uplift and volcanism of the SE Colombian Andes in relation to Neogene sedimentation in the Upper Magdalena Valley, Thesis University of Wageningen, Amsterdam, 208 p.

Van Thornout, F., Hertogen, J., Quevedo, L., 1992. Allochthonous terranes in northwestern Ecuador. *Tectonophysics*, 205(1-3), 205-221.

Vicente, J. C., 1972. Apercu sur l'organization et l'evolution des Andes argentino-chiliennes centrales au paralléle de Aconcagua: *XXIV° International Geological Congress (Ottawa), Section*, 3, 423-436.

Vicente, J.C., 1990. Early Cretaceous overthrusting in the Western Cordillera of southern Peru. In: Ericksen, G.E., Cañas Pinochet, M.T., Reinemud, J.A. (eds.) *Geology of the Andes and* *its relation to Hydrocarbon and Mineral Resources.* Circumpacific Council for Energy and Mineral Resources, Earth Sciences Series, 11, 91-117.

Vicente, J.C., Sequeiros, F., Valdivia, M.A., Zavala, J., 1979. El sobre-escurrimiento de Cuicha-Lluta: elemento del accidente mayor andino al noroeste de Arequipa. *Sociedad Geológica del Perú Boletín*, 61, 67-100.

Villamil, T., 1999. Campanian-Miocene tectonostratigraphy, depocenter evolution and basin development of Colombia and western Venezuela. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 153, 239-275.

Von Huene, R., Scholl, D.W., 1991. Observations at convergent margins concerning sediment subduction, subduction erosion and the growth of continental crust. *Review Geophysique*, 29(3), 279-316.

Von Huene, R., Weinrebe, W., Heeren, F., 1999. Subduction erosion along the Northern Chile margin. *Journal of Geodynamics*, 27, 345-358.

Webb, S.D., 1995. Biological implications of the Middle Miocene Amazon seaway. *Science*, 269, 361-362.

Woodward, H.P., 1958. Alleghany orogeny. *American Association of Petroleum Geologists Bulletin*, 42, 193.

Wörner, G., Moorbath, S., Horn, S., Entenmann, J., Harmon, R.S., Davidson, J.P., López Escobar, L., 1994. Large- and fine scale geochemical variations along the Andean Arc of Northern Chile (17.5°-22°S). In: Reutter, K.J., Scheuber, E., Wigger, P.J., (eds.) *Tectonics of the Southern Central Andes, Structure and Evolution of an Active Continental Margin*. Berlín, Springer Verlag, 77-91.

Zambrano, J.J., 1987. Las cuencas sedimentarias de América del Sur durante el Jurásico y el Cretácico: su relación con la actividad tectónica y magmática. In: Volkheimer W. (ed.) *Bioestratigrafía de los sistemas regionales del Jurásico y Cretácico de América del Sur, Comité Sudamericano del Jurásico y el Cretácico*, 1-48, Mendoza.

Zapata, T.R., Allmendinger, R.W., 1996. The thrust front zone of the Precordillera thrust belt, Argentina: a thick-skinned triangle zone. *American Association of Petroleum Geologists Bulletin*, 80(3), 359-381.

Zapata, T.R., Brissón, I., Dzelalija, F., 1999. The role of basement in the Andean fold and thrust belt of the Neuquén basin. In: McClay, K., (ed.) *Thrust Tectonics 99 (London), Abstracts*, 122-124.

Zentili, M., Maksaev, V., 1995. Metallogenetic model for the Late Eocene-Early Oligocene Supergiant porphyry event, Northern Chile. In: Clark, A.H. (ed.) *Giant Ore Deposits II*. Proceedings of the Second Giant Ore Deposits Workshop, Ontario, 138A-138N.

Zeil, W., 1979. *The Andes: A Geological Review*. Gebrüder Borntaeger, 1-260.

Mineral and Fossil Fuels Resources of South America

"The tectonic and stratigraphic evolution of the most representative basins of the South American Continent resulted in the creation of several classes of basins and hence in the generation of some very distinctive types of petroleum megasystems, that is, groups of basins presenting similar processes and conditions of hydrocarbon generation, migration and accumulation."

(Figueiredo and Milani, this volume)

"The coal deposits of South America occur mainly in two geological settings: in sedimentary successions that filled up some of the pre-Andean and Andean basins along the western border of the continent, and in the interior of the plate, related to Late Carboniferous to Permian deposits of the Paraná Basin."

(Lopes and Ferreira, this volume)

"The Andes Chain is one of the richest orogenic belts in terms of metallic ores in the world... Chile, alone, has about a quarter of the world's copper reserves and close to one third of those of molybdenum... On the other hand, important scientific studies have been dedicated to those types of ore deposits that are well represented in the Andean Belt..."

(Oyarzún, this volume)

"During the development of the South American Platform from the Archean to the Proterozoic, as well as during its tectonic evolution during the Phanerozoic, a number of mineral deposits were formed..."

(Dardenne and Schobbenhaus, this volume)

Petroleum Systems of South American Basins

Antonio Manuel Ferreira de Figueiredo and Edison José Milani

The sedimentary basins of South America may be ordered into a few classes related to the main tectonic-stratigraphic events that have originated and modified the South American Plate. These events were responsible not only for the present distribution of basins over the plate, but also for the volumes of hydrocarbons generated and effectively accumulated in these basins, and under an exploratory perspective, for their remaining potential.

The tectonic and stratigraphic evolution of the most representative basins of the South American Continent, discussed in this volume (see the text by Milani and Thomaz Filho), resulted in the creation of several classes of basins and hence in the generation of some very distinctive types of petroleum megasystems, that is, groups of basins presenting similar processes and conditions of hydrocarbon generation, migration and accumulation.

This concept was introduced by Dow (1974), while he was working for AMOCO. Since then, this concept has been extensively developed and used by most petroleum geoscientists. In this paper we will emphasize the source rock as the main factor to characterize the megasystem. With this in mind, the petroleum resources of the South American basins can be grouped into five petroleum megasystems:

- Subandean Megasystem, present from the N of Argentina to the S of Peru;
- Austral Rifts Megasystem, in the southern part of South America;
- Andean and Caribbean Foreland Megasystem, present from the N of Peru to Venezuela and Trinidad-Tobago;
- South Atlantic Rift Megasystem, present along the Atlantic coast of Argentina and Brazil;
- Intracratonic Megasystem, present in the interior sag basins of the Brazilian continental area.

These five megasystems include almost all reserves of gas and oil ever found in South America.

Reserves and Petroleum Megasystems

The petroleum systems of South America have been studied and analyzed in great detail by several authors and published mainly as Special Publications of the American Association of Petroleum Geologist. These special publications are very complete and comprhensive reviews, and include the Treatise of Petroleum Geology, the Atlas of Oil and Gas Fields (1991) and the Petroleum Basins of South America (Memoir 62, 1995). The Classic Petroleum Provinces (1990), published by the Geological Society of London, is also a very important reference. In more recent years other studies, made and published mainly by geoscientists on the geochemistry of petroleum, have contributed decisively to the understanding and clarification of the main aspects of these megasystems. Papers such as those of the AAPG Hedberg Research Conference of 1994, and of Mello and Trindade (1996), are excellent examples of these studies. Kronman *et al.* (1995) analyzed the oil and gas discoveries in the last decade in the South American basins, and forecast the remaining resources.

Excluding the Middle East, the South American countries, plus Mexico, have the largest volume of oil reserves of the Earth, being therefore of prime importance to future development, not only for the region, but also for the entire world. Taking into consideration only the South American basins Kronman's studies have indicated that the average success ratio for wildcat drilling during the last decade was between 20% and 30%, and is not declining. This may be interpreted as an indication that these basins may still hold large amounts of petroleum to be discovered. Even with the decrease of the numbers of wildcat wells drilled in the recent years, this ratio has been mantained and several giant oil and gas fields have been discovered very recently, including Marlim, Albacora and Roncador in the deep waters of the Brazilian Campos Basin; Cupiaga and Cusiana in deep reservoirs of the Llanos Basin in Colombia; the Camisea Gas Complex in the Peruvian jungle; and El Furrial in deep reservoirs of Maturín Basin, Venezuela. As a general result the oil and gas reserves of the South American basins have effectively increased, and the growing utilization of the most modern technology in petroleum research has been the fundamental key in achieving this goal.

Based on work of the authors cited above and data published around the world, it is possible to estimate the possible *in situ* volumes generated by the five megasystems:

Petroleum Megasystem	Age of Source Rock	Age of Reservoir	In Situ Volumes ($x10^6$bb/EO)
Subandean	Silur/Devon.	Paleoz/K/Terc	40 000
Austral Rifts	Jur/Eo-Cret.	Cret/Terc.	70 000
Andean/ Venezuelan Foreland	Mid-Cret/ Late Cretaceous	Cret/Terc.	2 000 000
South Atlantic Rifts	Early Cret/ Mid-Cretaceous	Cret/Terc.	100 000
Intracratonic	Devonian	Paleozoic	20 000

It becomes clear that the petroleum megasystem associated with the basins affected by the lateral displacement of the Caribbean Plate along the northern margin of the South American Continent in Venezuela and the foreland basins of Colombia, Ecuador and Peru, is, by far, the richest megasystem. This fact is related to the previous existence of optimum conditions for the development of thick and extensive sections of pelitic rocks with high content of organic matter, in geological conditions very similar to those occurring in the Tethys Sea of the Middle East, as well as favourable tectonic conditions for huge traps resulting from plate interaction and collision.

The BP Statistical Review of World Energy, published at the end of 1998, indicated that the proven world oil reserves are around 1.053 billion barrels, of which the contribution of the South American petroleum systems is a modest 89.5 billion barrels (8.3%). According to the same publication, the proven world gas reserves are about 5.170 trillion cubic feet, of which the contribution of South American basins is about 219 billion cubic feet (4.3%). It is believed that this large difference in terms of oil and gas content in the South American basins reflects the absence of a competitive and developed gas market on the continent, resulting in low prices for this energy source and the consequent lack of incentive to exploration, rather than the absence of gas in the basins.

If it is assumed that the current world production is about 75 million barrels of oil per day, the world reserve/production ratio (R/P) would be around 41 years for oil and 63 years for gas. For South America, these figures are 38 years and 71.5 years respectively. Again, it is clear that in the near future gas will become very important, and a growth of its presence in the energy matrix of the South American countries will be imperative.

The table shows the reserve and production figures for oil and gas in South America:

Country	Oil and Gas Reserve ($x10^9$bbl/TCF)	Oil and Gas Production ($x10^3$bbl/day-10^9m^3/year)
Argentina	2.6 – 24.1	890 – 29.3
Bolivia	0.5 – 4.3	30 – 3.2
Brasil	7.1 – 8.0	990 – 6.5
Colombia	2.6 – 6.9	765 – 6.3
Ecuador	2.1 – 3.7	385 – 0.0
Peru	0.8 – 0.5	115 – 0.0
T. & Tobago	0.5 – 8.3	135 – 8.6
Venezuela	72.6 – 142.5	3335 – 29.9

Source – BP Statistical Review

In order to understand and describe the petroleum megasystems referred to above, the characteristics of each megasystem in terms of the source rocks, main reservoirs, traps, migration pathways and timing will be described hereunder. Comments are made on the accumulation and preservation of the hydrocarbons, and an example of each type of petroleum megasystem will be made with reference to a specific basin and/or giant field.

Subandean Basins Petroleum Megasystem

In this petroleum megasystem are included several types of basin that were affected by the plate tectonics regime responsible for the generation of the Andean Chain that developed along the western margin of the South American Plate, resulting from collision of the South American and Nazca/Pacific plates.

These basins developed as the result of several cycles of subsidence and accumulation of sedimentary packages, alternating with periods of uplift and erosion (polycyclic basins). They occur from the northern part of Argentina (Northwestern Basin) and in Bolivia to the southern part of Peru (Ucayali/Madre de Dios basins).

In these basins (Fig.1) the petroleum megasystem is characterized, essentially, by oil and gas generation with source in Paleozoic shale (Late Silurian-Devonian) deposited in an open marine environment (marginal Panthalassan Basin). These rocks hold a high content of organic matter, and are associated with the basal section of a very thick (several hundreds of metres) pelitic transgressive sequence that onlaps sandy siliciclastic deposits accumulated in shallow epicontinental seas.

In the geological context of the Subandean Basin in northwestern Argentina and southern Bolivia there occurs the Kirusillas Formation (Late Silurian), and the Icla and Los Monos formations (Devonian), responsible for the generation of large amounts of oil and gas accumulated in intercalated sandstone sequences known as the Santa Rosa and Huamampampa formations. Besides, other Paleozoic (San Telmo, Tarija, Tupambi, and Escarpment), Cretaceous (Ichoa, Cajones and Petaca) and Tertiary (Tranquitas and Chaco) clastic reservoirs have their source in these rocks.

Andean tectonics was responsible for the development of a very large and continuous fault and fold belt striking N-S, and affecting mainly the western margin of the basin. Several pulses of compression originated huge structural traps residing in complex anticlinal systems associated with thrust faults with detachment in the less competent beds, resulting in shortening of several tens of kilometres and in the thickening of the Los Monos shaly section.

Asymmetrical folds with very tight limbs constitute most of these structures. However, stratigraphic and combined traps are also present, and these are mainly related to the channelling of fluvio-glacial sandstone deposited in large valleys sculptured by deglaciation-related mass flows during the Carboniferous.

The main epoch of generation and migration of hydrocarbons seems to be correlated with the last and very recent tectonic pulse of Andean uplift during the Miocene and Pliocene. During this event, the source rocks were rapidly depressed in the basins and folded into synclinoria, attaining both the oil and the gas window. As a result of this rapid deepening, large amounts of oil, and principally gas, were produced and trapped in the newly formed structural features. These accumulations in the deep reservoirs of Devonian age have been discovered only recently due to the technological advances that have permitted subsurface mapping and drilling of wells below depths of 5000 m through shallow high-pressures zones in search of hydrocarbons in the crest of the anticlines.

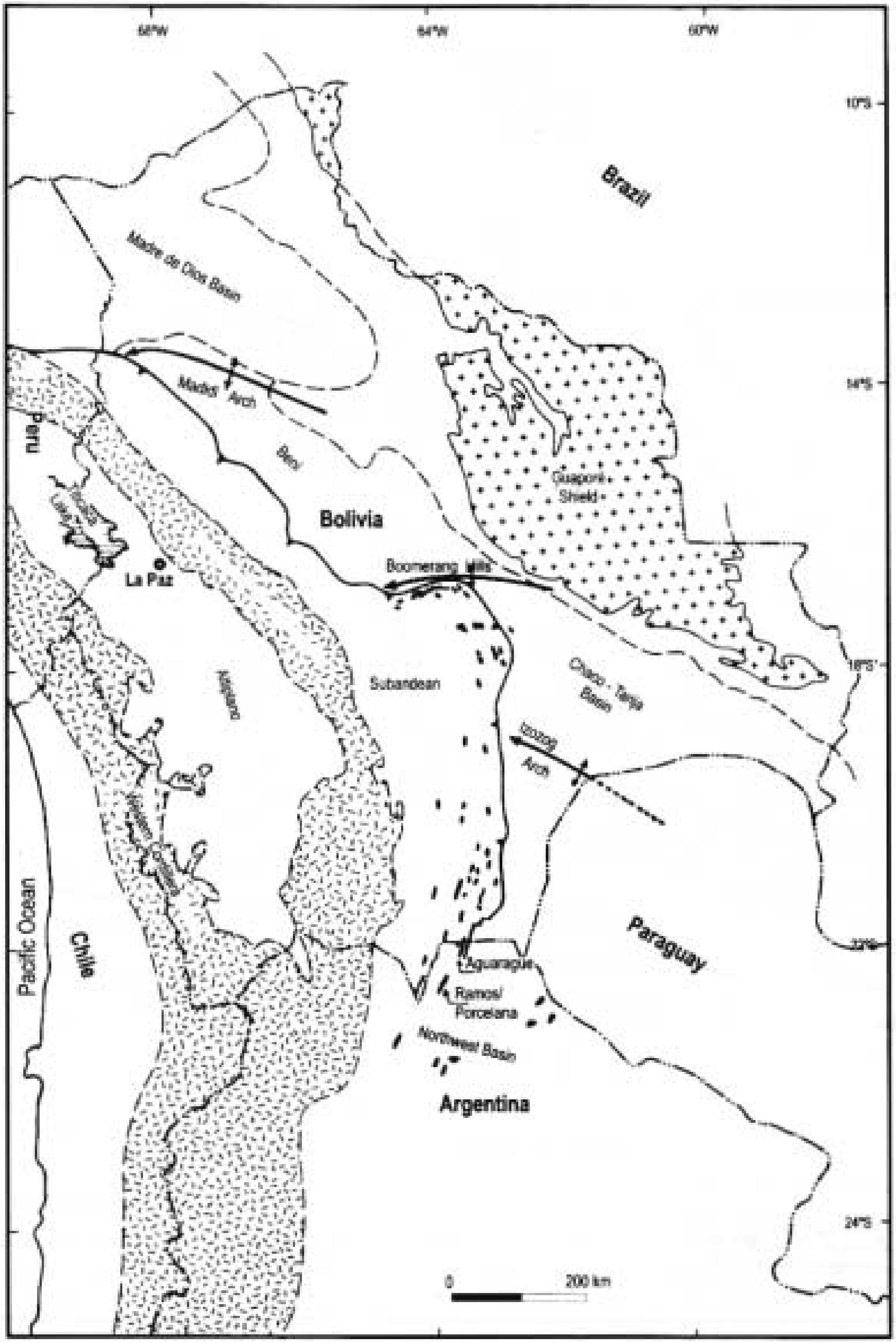


FIGURE 01 – Location map of the southern portion of the Subandean Petroleum Megasystem, in Argentina and Bolivia, displaying the main structural features, associated basins and major oil and gas fields.

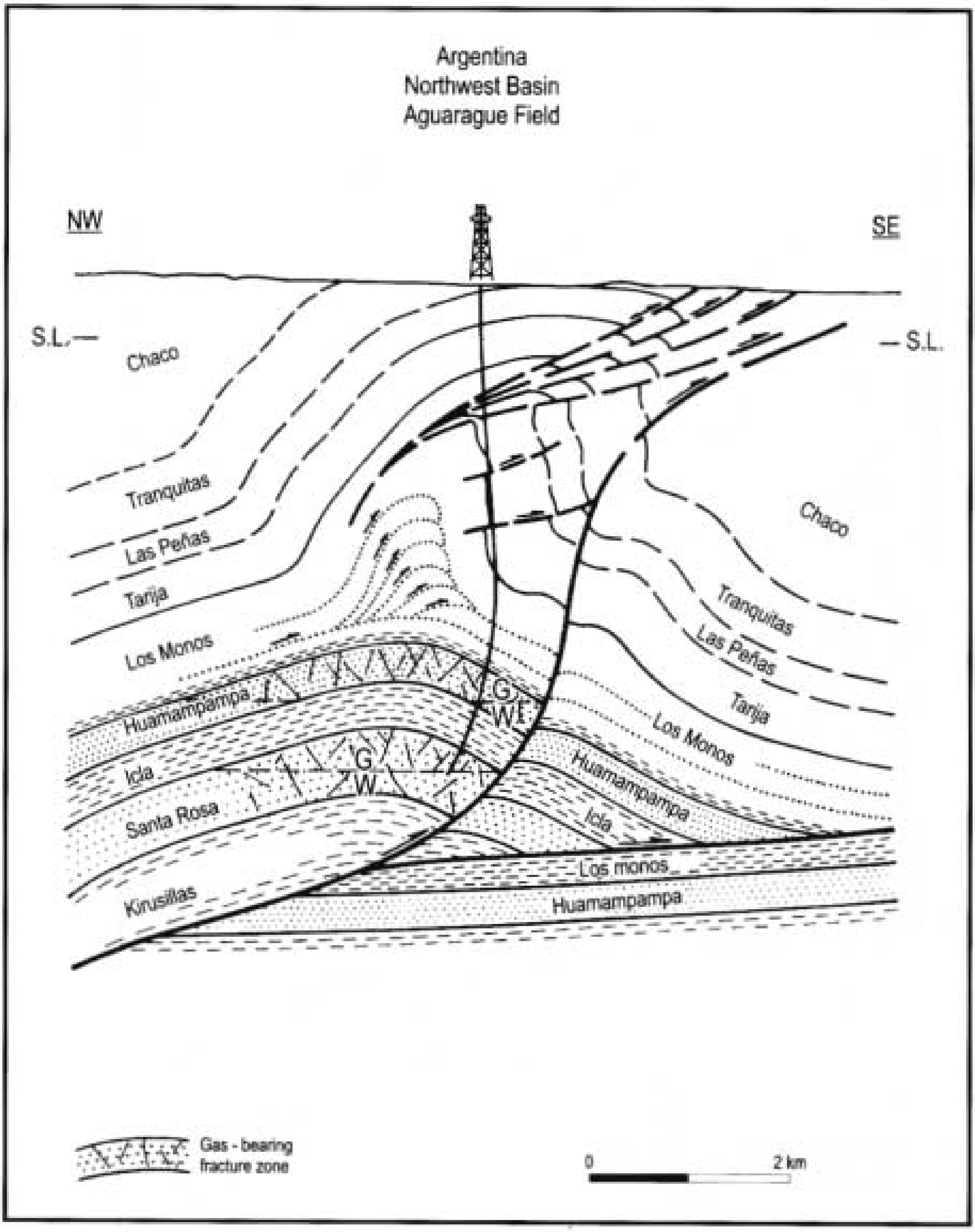


FIGURE 02 – Schematic structural cross-section of the Aguarague Field, Northwest Basin, Argentina (modified after Grieco, 1991)

The main reservoirs consist of very thick sandstone sequences composed of clean fine to very fine-grained quartzarenite beds, deposited on a shallow marine platform during the Devonian. Normally, these sandstone units are highly cemented by extensive secondary quartz overgrowths and, due to their high degree of compaction, developed a very dense network of natural fractures in order to accommodate the intense folding caused by the stress field.

The fracture zone is the key to the excellent reservoir properties displayed by these sandstone units, and this zone is only present as a narrow band associated with the crest of the anticlinal apexes. This feature causes great difficulties in the exploration for this target, since the position of these apexes is hard to interpret from seismic data. The poor quality of the seismic data is directly related to the rough topography and also to the structural complexity of the fold belt. Very often there occurs the lateral migration of the position of the crest of the fold with depth, resulting in problems in using surface mapping or shallow seismic horizons to orient the drilling objectives in the deeper levels of the anticline.

Included in this megasystem may be cited the giant gas pools occurring in the Northwestern Basin encompassing the Subandean folded belt in Argentina and Bolivia. Such pools include those of the Ramos, Aguarague, San Alberto and San Antonio, Colpa-Caranda and El Palmar fields. Small and shallow oil fields are normally associated with these gas pools. Although this type of field is not well known, recent technological advances in drilling and seismic surveying have led to the discovery of several very large gas fields with similar volumes as those found in the Camisea Complex of southern Peru that is also part of the same megasystem.

These basins contain a reserve having a potential of around 15 trillion cubic feet of gas (TCFG) in addition to the 11 TCFG already discovered in the Camisea Complex of the Ucayali Basin. These gas resources will supply the energy needs of the southern South American countries for very many years to come. It is now proposed to describe in further detail some of the representative giant gas accumulations associated with this megasystem.

Ramos-Porcelana and Aguarague Fields

These gas fields are situated in the Northwestern Basin, northern Argentina, in the vicinity of the City of Salta. This basin (Fig. 1) has geological continuity with the Subandean folded belt area of southern Bolivia, to the N, where very recent discoveries have been made in the same Subandean Ranges (San Alberto, Itau, and San Antonio fields).

The Ramos-Porcelana Field is situated on the same anticlinal structure, some 70 km long, in Argentina. On the Bolivian side, this structure is more than 70 km long, and at least two giant gas pools at its apex are being evaluated (San Alberto and Itau). The folded area is about 10 km wide, but the gas-bearing zone is restricted to the crest of the fold, where the Huamampampa reservoirs are intensely fractured. The original matrix porosity is reduced by cementation reaching values around 3 to 4%, which helps to accommodate large gas volumes.

These fields, as well as the Aguarague Field in a parallel anticlinal structure to the E may hold more than 3 TCFG each. The gas is contained in two main reservoirs situated below the Los Monos and Icla formations, the main Devonian source rocks, and above the Kirusillas Formation. These highly fractured siliciclastic sequences normally occur in the upper part of the Huamampampa, Icla and Santa Rosa formations (Fig. 2). The gas column in these reservoirs may be as thick as 600 m.

The natural fracture system created by the regional stress field may be sub-divided into two sub-systems. The first consists of large and open fractures responsible for the very high gas production, and the second consists of a very dense network of small and closely-spaced fractures that act together with the original matrix porosity as the stocking element to accumulate the large gas volume.

This dual system is extremely efficient, and allows open flow production that is locally greater than 1.5 million cubic metres of gas per day in each reservoir. Therefore, each well may present a production capacity exceeding 2 million cubic metres per day. This volume of gas is sufficient to sustain a thermoelectric plant producing 500 MW.

The source rock of these gas fields, and also of the small oil pools in the Late Carboniferous, Cretaceous and Tertiary sands, are the aforementioned Los Monos, Icla and Kirusillas formations of Late Silurian to Devonian age. The basal part of these marine shale sequences is very rich in algae-derived organic matter, and since they onlap over the main reservoirs the migration pathway becomes very short and only primary migration may be required. Fault zones also provided a way by which the oil and gas was able to reach the shallow reservoirs, locally only a few km above the source rock.

The traps are structural (Fig. 2), and the hydrocarbons are retained in the apexes of the anticlines, sealed by thick shale sections. The spill-point is apparently controlled by the low-angle reverse faults, and in some areas such as in Ramos Field there may occur the thickening of the reservoir due to the presence of duplex features affecting the Huamampampa sequence, doubling the length of the fractured pay-zones, and allowing the storage of very large reserves.

Camisea Complex

This complex of gas-bearing structures is situated in the southern part of the Ucayali Basin in Peru, close to the border between this basin and Madre de Dios Basin (Fitzcarrald Arch). This complex was found in the first years of the 1980s and is situated in the middle of a tropical rain forest, about 500 km S of the City of Lima, Peru (Fig. 3).

It is formed by at least, three gas accumulations known as Cashiriari, San Martín and Miyapa and holds around 11 TCFG and 600 million barrels of associated condensate, equivalent to a reserve of 2.6 billion barrels of oil.

The gas is trapped in anticlinal folds associated with low-angle reverse faults and several reservoirs are present in this structure (Fig. 4). The production capacity of these reservoirs is around 800 000 cubic metres per day.

Evaluation wells drilled very recently indicate that the main reservoirs, formed by Cretaceous sandstone units, deposited in fluvial and fluvio-deltaic environments (Cushabatay, Agua Caliente and Vivian formations), are extensively fractured by compression. However, long-duration

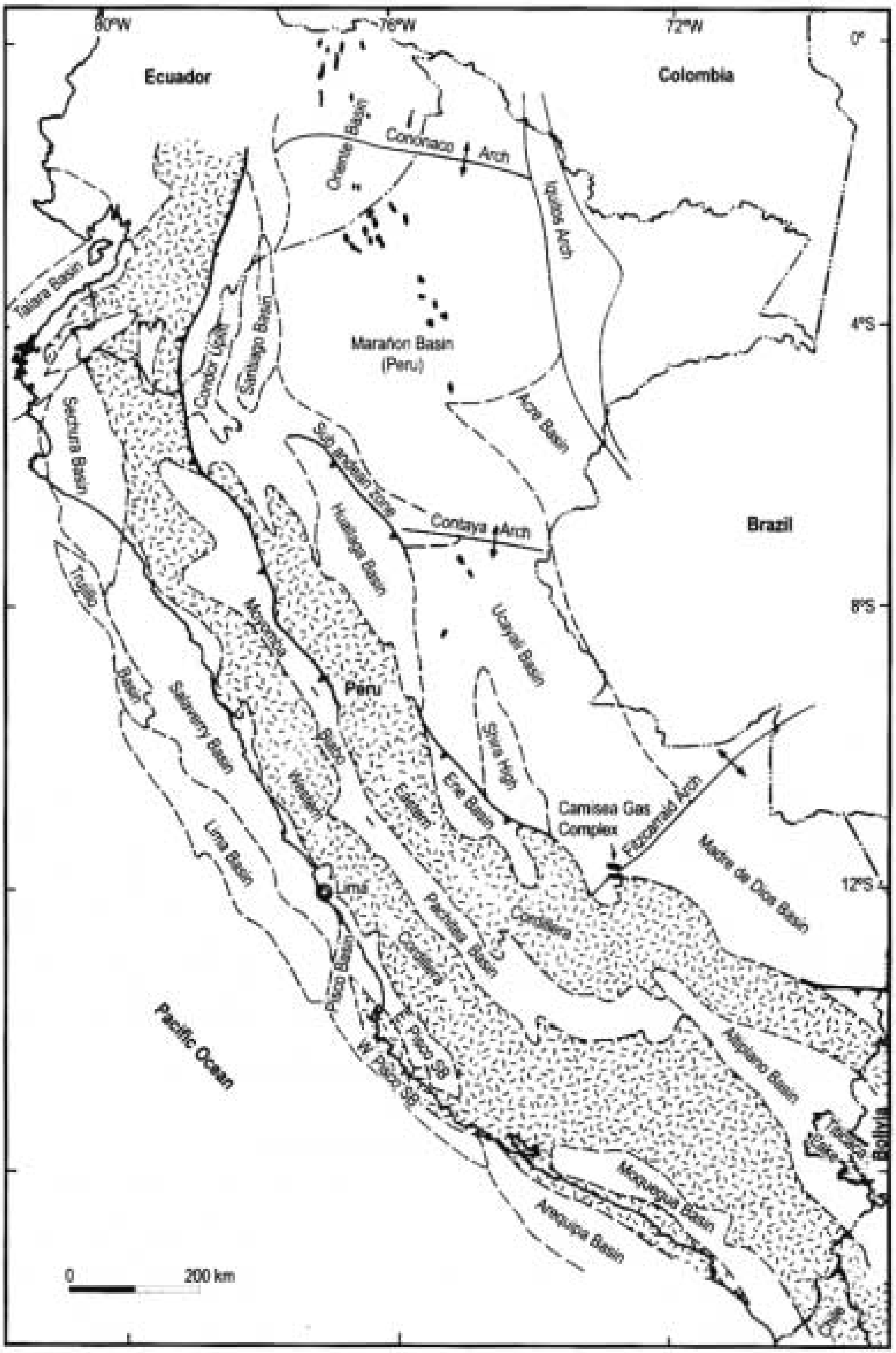


FIGURE 03 - Location map of the northern portion of the Subandean Petroleum Megasystem, and the southern portion of the Andean Foreland Petroleum Megasystem, in Peru, displaying the main structural features, associated basins and oil and gas fields (Camisea Gas Complex).

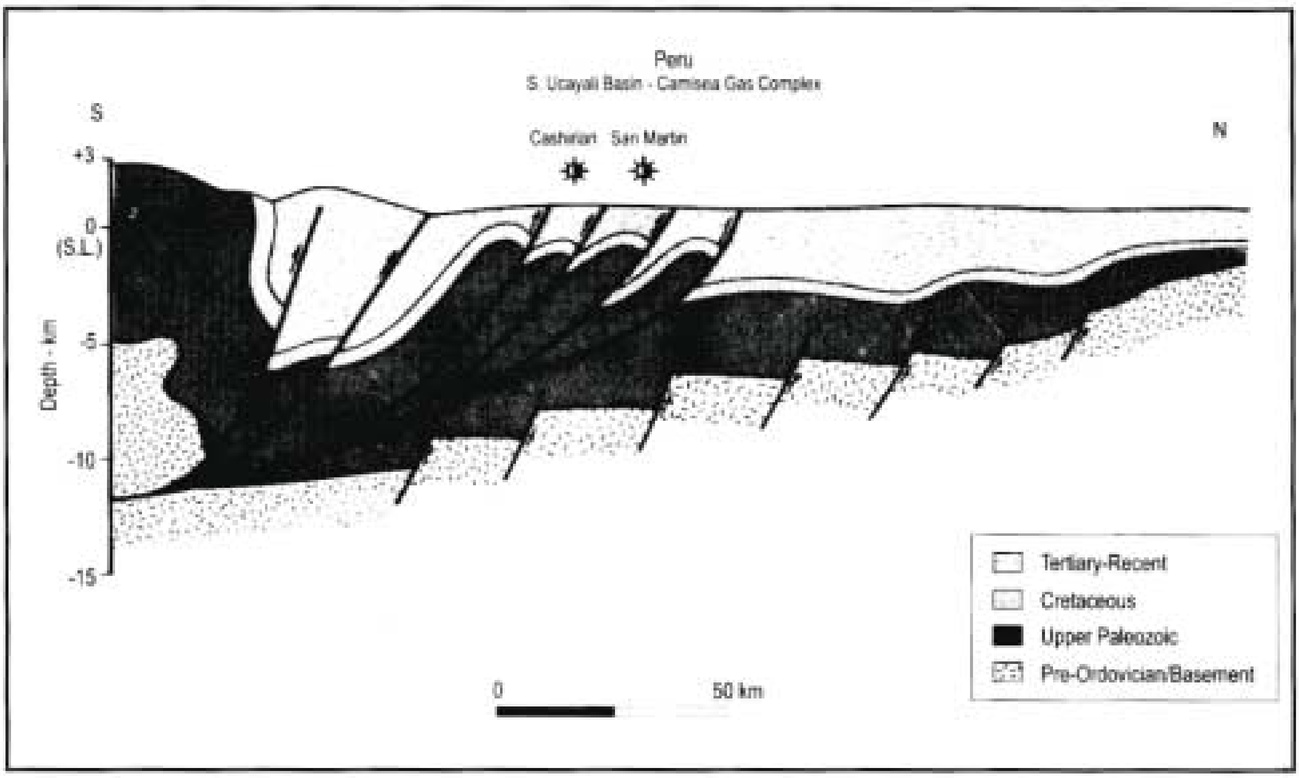


FIGURE 04 – Schematic regional cross-section of the southern Ucayali Basin, Peru, exhibiting the Camisea Gas Complex structural framework (modified after Mathalone and Montoya, 1995).

production tests also indicate a strong compartmentation of the structures, possibly by faults (strike-slip faults?) of less magnitude that are very difficult to define in seismic profiles. These barriers strongly affect the gas production capacity per well, and may result in heavy exploration and development costs due to the increase in the number of wells required to produce the same amount of gas.

The source of the Camisea Complex is still open to discussion.The existence of at least two potential source rocks in the basin is well known. The first are the Devonian marine shales of Cabanillas Formation that are comparable in terms of their thickness and areal extent to the Los Monos Formation of Argentina and Bolivia.The second are the shale and marlstone beds of Permo-Carboniferous age. The Permo-Carboniferous beds consist of deep-water and slope facies, very rich in algae-derived organic matter and time-equivalent to the high-energy platform carbonates of the Copacabana Formation. It may be that both source rocks were responsible for feeding gas and condensate to the huge structures of Camisea Complex.

Austral Rifts Petroleum Megasystem

The Austral Rifts petroleum megasystem occurs along the southernmost Andean margin in Argentina, and results from a major paleogeographic change in the southern part of the South American Continent. This change occurred during the earliest Mesozoic, and is associated with the shift of the magmatic arch westwards to a new position that approximates that of the present coast of Chile (Mpodozis and Ramos, 1989).

In this new plate tectonics scenario, a series of back-arc rift basin developed and remained active as typical extensional basins until the Middle Cretaceous, when a generalized tectonic event took place causing strong deformation in these basins. Amongst these back-arc rift basins are the Neuquén, San Jorge, Bolsones and Cuyo basins. The rock sequences in some of the basins were completely inverted during the Andean Orogeny, whereas others, mainly those in the E, were partially preserved (Fig. 5).

These rifts evolved from isolated deep fresh-water lakes to a more continuous gulf invaded by marine waters coming from the Pacific and filled up by siliciclastic sequences. In this geological context, thick pelitic units were deposited, providing excellent conditions for the preservation of organic matter, associated with deep lakes and relatively restricted marine conditions. The filling up of the basin led to the development of shallow water environments where fluvio-deltaic systems and carbonate platforms were installed resulting in the deposition of thick reservoir rocks, principally during the Late Cretaceous times.

Notwithstanding, that the original composition of the source rocks was more favourable for oil generation, they underwent deep burial in function of the volume of Andean cyclic sedimentation, reaching the gas window in most depocenters. Therefore, these basins have potential for both oil and gas. Typical examples of this petroleum megasystem may be found in the Neuquén Basin, mid-west Argentina.

Neuquén Basin

This basin has a triangular shape, covers an area of about 160 000 km^2 (Fig. 6), and, in like manner to all other austral rift basins, began as the result of an extensional episode in Late Triassic/Early Jurassic times with the

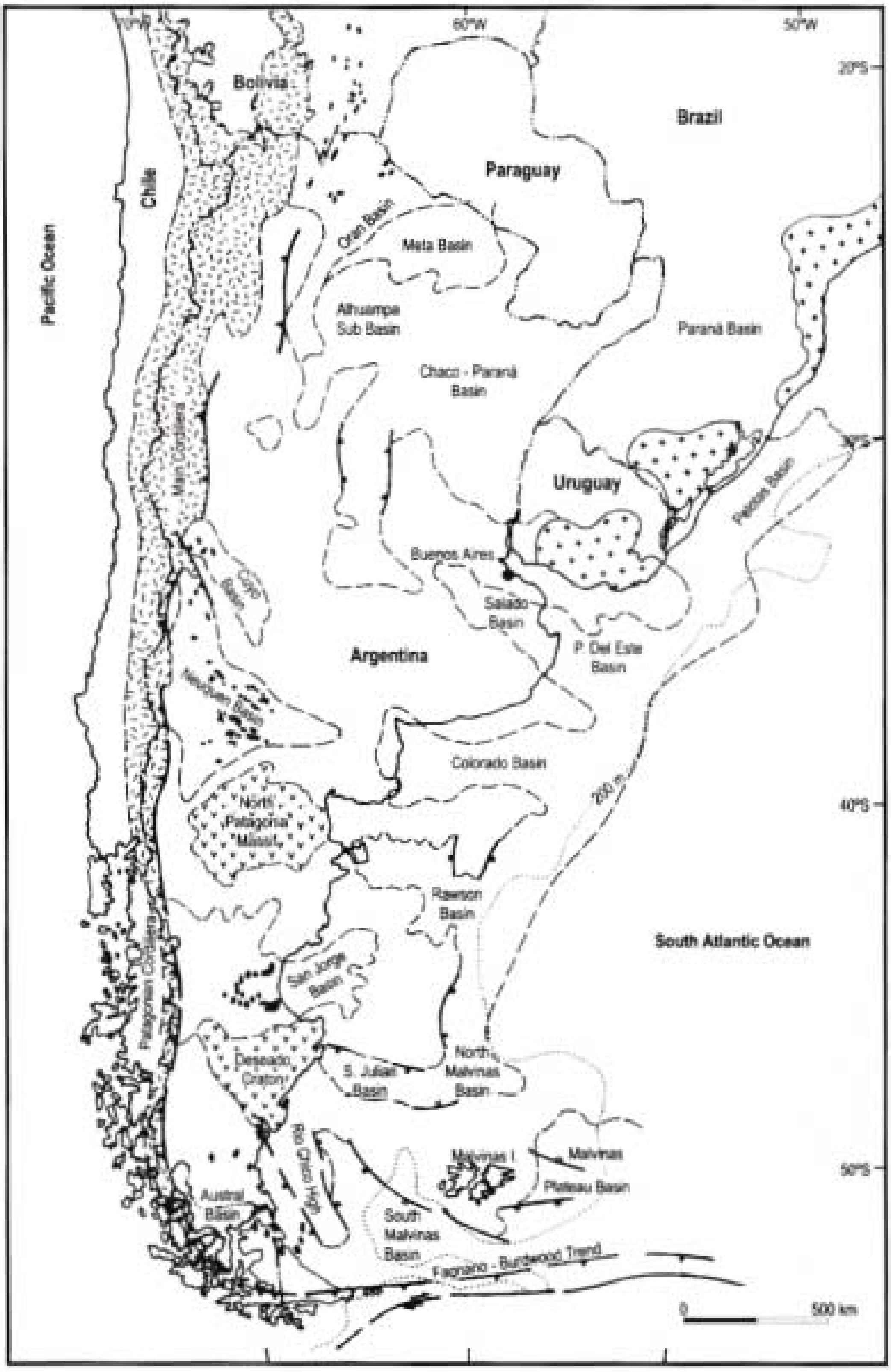


FIGURE 05 – Location map of the Austral Rifts Petroleum Megasystem, Argentina, displaying the main structural features, associated basins (Cuyo, Neuquén and San Jorge) and oil and gas fields.

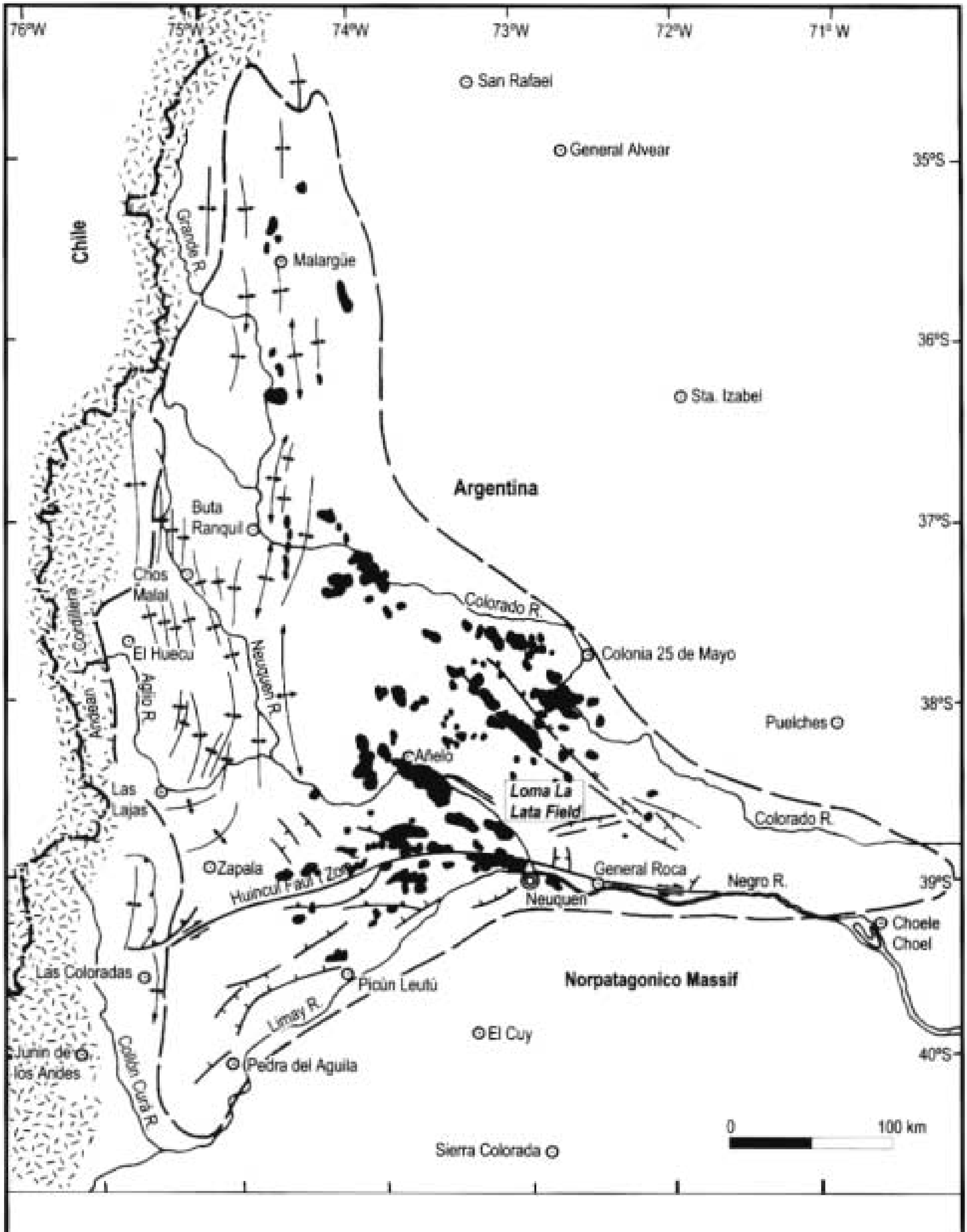


FIGURE 06 – Neuquén Basin map, Argentina, exhibiting the main structural features and oil and gas fields (modified after Legarreta et al., 1999).

widespread eruption of volcanic rocks. Following this event, shallow fresh-water lakes were developed in response to the stress field, since the connection with the Pacific was, at that time, limited by a submerged magmatic arch at the edge of the South American Plate (Legarreta *et al.*, 1999).

The shallow lakes evolved rapidly to deeper lakes with restricted circulation where dark shale beds with very high content of organic matter were deposited. Due to increasing aridity and evaporation, the salt content of the lakes reached the point that these became hypersaline. As a consequence of increased salinity, anoxic conditions developed and a permanent water stratification column was established giving rise to very favourable conditions for the preservation of the organic matter. The Puesto Kaufman Formation is representative of this rift stage, and may contain more than 2000 m of organic shale and coarse clastic sediments. The TOC content of this section varies from 2 to 8%, the hydrogen index is about 900 mgHC/g TOC, and the kerogen is type I, characterizing an excellent source rock.

The syn-rift isolated semi-grabens evolved to a more continuous basin (Gulf of Neuquén) that was invaded by normal marine Pacific salt water during the trangressive event associated with the Early Cretaceous rise in sea level. In this wider and shallower basin, a thick sequence of shale and marlstone beds was deposited, generating the best source rocks of the Neuquén Basin known as the Los Molles (Early Jurassic to Middle Jurassic) and the Vaca Muerta (Tithonian) formations. In the Los Molles Formation the TOC values are about 1 to 3%, with some intervals reaching 6%. The kerogen is of types I and II. The Vaca Muerta Formation consists of beds of bituminous marlstone and laminated shale. The TOC values of the Vaca Muerta Formation are around 2%, with peaks of 4%. The kerogen is of types I and II, and mainly derived from amorphous organic matter.

Besides these source rock sequences, there occur Hauterivian (Agrio Formation) sedimentary beds that are rich in organic material, also important as source rocks in some parts of the basin. The post-rift sedimentary sequence that fills the basin displays several transgressive-regressive cycles from the Jurassic to the Tertiary in response to the Andean tectonic pulses and sea level fluctuations. These cycles are limited by regional unconformities observed at the top of the Jurassic, in the Early Cretaceous (Kimmeridgian to Albian), and in the Late Cretaceous-Tertiary strata. These cycles are known as Rio Grandean and Andean cycles.

The stratigraphic column for these cycles consists of a very thick clastic wedge of continental origin at the base, covered by normal marine sediments. The marine beds change rapidly to those showing evidence for more arid and restricted marine and continental environments, including conditions favouring the accumulation of evaporite suites. Therefore an almost perfect association of source rock, reservoir and cap rock is present in most cycles. Linking this to the late basin structuration by the Andean Orogeny, and the creation of huge traps, this basin became highly favourable for hydrocarbon resources.

The Early Cretaceous cycle marks the maximum extent of the trangression when there occurred the deposition of abundant carbonate and marlstone beds as well as the sediments of the Vaca Muerta Formation, the main source rock. At the beginning of the Late Cretaceous, there started a new and very important tectonic pulse that was maintained during most of early Tertiary causing the deposition of a continental clastic wedge, subsequently covered by marine sediments. In the middle to late Tertiary there occurred the development of a pyroclastic complex that remained active until the Quaternary.

This basin displays a great variety of levels of reservoir-rocks throughout the stratigraphic column. Even the basal fractured and weathered volcanic Triassic rocks are hydrocarbon-bearing in some areas. However, the main reservoirs occur in the clastic sequences of the Punta Rosada, Tordillo, Mulichinco, and Agrio formations and the Neuquén Group.

The Punta Rosada Formation (Cuyo Group) reservoirs are representative of a marine marginal sequence consisting of fluvial and fan delta lobes having very good petrophysical properties. The Tordillo Formation reservoirs, also known as the Sierras Blancas Formation, consist of a thick sequence of coarse-grained fluvial-alluvial sandstone beds with provenance from the SE and SW that filled the basin and were subsequently buried by shale and marlstone of the Vaca Muerta Formation. During the Tithonian, shallow marine platforms were installed and thick carbonate and dolomite sequences were deposited. These sediments comprise the Loma Montosa Formation, and in some areas present very good reservoir properties due to intense dissolution and dolomitization.

At the top of the Vaca Muerta Formations there occur the clastic sediments of the Mulichinco Formation, deposited in fluvial, littoral and shallow marine environments. These beds are overlain by the clastic sediments of the Agrio Formation, consisting of eolian deposits displaying excellent reservoir qualities (Aviles Member).

During the Late Cretaceous, red beds of the Neuquén Group formed very thick deposits, mainly consisting of sandstone. The absence of significant pelitic intercalations may reflect the absence of commercial accumulations of hydrocarbons in these sediments; added to which there lacks an effective cap rock.

The very fine-grained shale and marlstone beds constitute the main cap rock beds in this basin, which are also responsible for the oil and gas generation. The evaporitic layers are present in the uppermost part of all the sedimentary cycles, acting as a very effective seal to hydrocarbon migration. The deposition of salt was mainly controlled by the dynamics of the magmatic arc in the western part of the basin. The tectonic movements of this arc were responsible for the large amount of water influx to the basin, and the establishment of a negative hydrological balance triggered the deposition of thick layers of anhydrite and halite. Subsequently, these plastic halite beds became very important as detachment zones for the propagation the Andean deformation throughout the basin, creating structural trends during the Cretaceous and Tertiary.

The several source rocks sequences, maturation and the main phase of oil expulsion and migration to the structural traps originated during the phase of Andean compression that started at the end of the Cretaceous. At present, the source rocks of the Vaca Muerta Formation have attained the initial phase of the gas window, whereas, the source beds of the Agrio Formation are still in the oil window. The older source beds of the basin are all in the gas window. This seems to explain why the Neuquén Basin contains such large volumes of oil and gas trapped within its fields.

The very intense tectonism developed at the end of the Cretaceous, and lasting throughout the Tertiary, oriented and controlled the initial migration of oil and gas in the basin. However, very important remigration must have taken place since the tectonic evolution gave origin to new hydrocarbon pathways along which the oil and gas was directed to newly-formed structures and/or combined traps associated with fractured reservoirs.

The main structural grain of the basin may be analyzed in two domains. The first domain is situated in the western part of the basin and was strongly affected by the Andean Orogeny, presenting very intense deformation and displaying faulted anticlinal trends. To the E of the volcanic arc, the deformation became more and more gentle and the structures are related to older basement faults and folds. Nevertheless, compression was the principal element by which structural trends were created in the basin, affecting not only the overall geometry of the basin, but also the facies changes and sandstone distribution. In this sense, the oil distribution is directly related to structural evolution of the basin.

The conspicuous Huincul Dorsal (Fig. 6) crosses the entire southern limb of the basin, and occurs as a very narrow deformation zone more than 200 km long. Along this zone, extensional and compression features are present, most of them *en echelon*, and interpreted as a huge transpression zone, active since the Jurassic. In this feature, several oil and gas fields were discovered, and a large volume of oil was trapped in very different types of trap. In the Neuquén Basin, most of hydrocarbons discovered are contained in 345 accumulations. The volume is estimated at 6.5 billion barrels of oil equivalent, a great part of which was found in combined traps. The Puesto Hernandez and Loma la Lata fields are very good examples of these types of accumulation. The Puesto Hernandez Field has an original oil recovery reserve of 620 million barrels, consisting mainly of good quality oil. The Loma la Lata Field has a reserve of 1.6 billion barrels of oil equivalent, composed essentially of gas. Steep dipping reservoir beds truncated by an important erosional unconformity and sealed by abrupt facies changes and permeability loss associated with the unconformity, are common features of both giant fields. Accordingly to Vergani *et al.* (1995) about 45% of Argentina's production comes from this basin.

Caribbean and Andean Foreland Petroleum Megasystem

In this megasystem are included almost all the petroleum occurrences along the transcurrent and foreland basins bordering the Caribbean coast in Venezuela and the Andean Chain of Colombia, Ecuador and northern Peru. These basins have in common the same rock sequence, deposited in a partially restricted marine environment from the Early to Middle Cretaceous, along the passive and mid-arc margin, developed along the northern and northwestern coast of South America.

Very favourable environmental conditions for generation and preservation of large amounts of organic matter were associated with the break-up of the South American and North American continents, originating a gulf known as Tethys Sea, a fore-runner of the present-day Gulf of Mexico. In this relatively shallow, warm and protected body of salt water that subsequently covered very large areas of the western margin of South America, reaching northern Peru (Cretaceous South American Seaway), thick sequences of shale and marlstone were deposited with a high content of algae-derived organic matter.

The Tethys Seaway is considered to represent the passive margin phase of a rifting process that took place in the area during the Triassic-Jurassic extensional event, similar to that responsible for the creation of the Austral Rifts. In this case the Tethyan Rift started its evolution in the Caribbean Region and propagated toward the S through Colombia, Ecuador, and northern Peru (Jaillard *et al.*, 1990).

At the end of Cretaceous and during the Tertiary this basin was submitted to the strong compressional effects of the Andean Orogeny (Daly, 1980). Subsequently, in the Neogene, the northern margin was also affected by the dextral shear stress field produced by the start of the Caribbean Plate motion. As a result of these complex geological events, the original basin, of enormous size, was divided into several segments by the uplift of regional arches, generating the present-day configuration.

In this megasystem the source rock sequence was essentially developed in a marine carbonate environment. These marlstone and shale beds, known by different lithostratigraphic names according to the specific basin in which they occur, include the Napo Formation in Ecuador; the Villeta and Gacheta formations in Colombia; and the La Luna and Querecual formations in Venezuela. All of these rocks have the same excellent characteristics of TOC content (higher than 5%), type I algae-derived kerogen and high hydrogen index, responsible for generating, where the thermal conditions have permitted, enormous amounts of liquid and gaseous hydrocarbons. To illustrate this petroleum megasystems some basins and giant oil fields will be described.

Oriente Basin

This basin (Figs. 3 and 7) covers an area exceeding 900 000 km², being known by different names in Colombia, Ecuador, Peru and Brazil: Putumayo, Napo, Marañon, Ucayali, Acre, Santiago, Huallaga and Ene basins. It contains several first-order sequences developed in a Silurian-Devonian back-arc basin, followed by those developed in another back-arc basin during the Permo-Carboniferous. In the Jurassic-Triassic it existed as a rift basin, and finally as a foreland basin during Cretaceous-Tertiary times.

In terms of the hydrocarbon potential, only the foreland basin is important as it contains both the source rocks and the reservoir rocks. The source rocks are marine marlstonestone and shale beds included in the Middle Cretaceous Napo/Villeta/Chonta formations, equivalent to the La Luna Formation in Venezuela. According to Carneiro and Cavalcanti (1994), the main depocenters where these

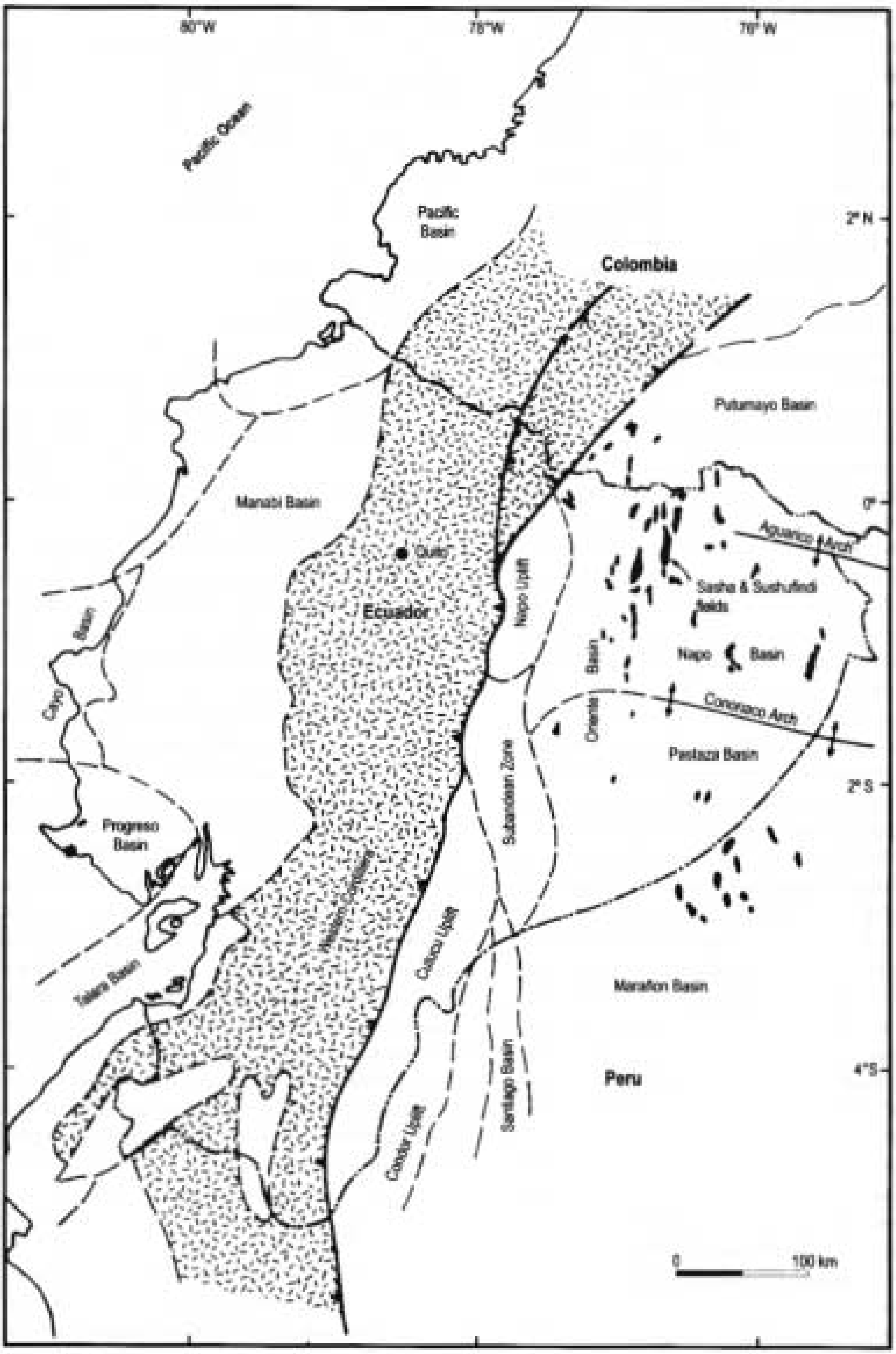


FIGURE 07 - Location map of the Oriente Basin (Marañon, Napo and Putumayo basins) of the Subandean Petroleum megasystem, in northern Peru, Ecuador and southern Colombia, displaying the main local features and oil and gas fields.

rocks attained the thermal conditions necessary to generate hydrocarbons are situated in the western part of Marañon Basin, Peru, very far from the main accumulation sites in northern Peru and Ecuador. Due to this fact, major secondary migration must have occurred over distances of some hundreds of kilometres.

The main reservoirs occur in the sandy units of the Hollin and Napo formations deposited in the Early and Middle Cretaceous. The Hollin reservoirs are also known as the Caballos and Cushabatay formations, the sediments of which comprise a regressive cycle associated with a generalized lowering of the sea level. The basal section consists of non-marine alluvial coarse-grained sandstone and conglomerate beds filling depressions and representing deposits characteristic of incised-valley channels. This sequence is covered by a very thick coarse to medium-grained sandstone unit associated with braided rivers. In some places eolian beds are developed at the top of this section.

This continental area of sedimentation was gradually inundated by a continuous sea level rise and covered by coastal sandstone deposits, strongly reworked by tides and waves, in addition to beds of marine shale and marlstone. The Hollin Formation lies at the base of the Cretaceous sequence throughout the entire basin, and attains a thickness exceeding 400 m in the depocenters.

The Napo Formation reservoirs are the least developed in terms of thickness and areal distribution, but still are present in most areas of the Oriente Basin. These reservoirs consist of sandstone intercalated with shale and marine carbonate units responsible for the oil generation. They were concentrated in three pulses of clastic input coming from the source area to the E where the granitic Brazilian Shield fed material to the basin. These clastic sedimentary pulses are known as the U, T and M sandstones or the Agua Caliente and Vivian formations, and were deposited during the progressive sea level rise and transgression of the Middle Cretaceous seaway during the Late Albian and Campanian.

They consist of deltaic and barrier-bar facies reworked by tides representing littoral, estuarine and shallow marine clastic deposits, developed over high-energy carbonate banks. This succession of deep-water marlstonestone and shale beds, carbonate bank deposits and clastics are interpreted to have formed as a consequence of tectonically-controlled third-order cyclic variations of the general trangressive sequence. Towards their source these three distinctive sequences grade laterally into a thick body of fluvial-alluvial sandstone units. This fact may explain the predominance of low API heavy and viscous oil in the accumulations of this basin, since a very intense process of fresh water invasion is present due to the existence of a very continuous carrier bed for the oil and water.

The main structural trends from where hydrocarbon production is obtained are related to folding and faulting resulting from the compression related to the Andean Orogeny. The oil is contained in relatively gentle anticlinal folds oriented N-S and associated with low angle thrust faults. The Late Cretaceous traps are usually filled with hydrocarbons, whereas the folds originated during the Tertiary are barren due to the absence of good cap rocks in the overlying continental beds of Tena-Tiyuyacu formations.

The uplift of the Andean Chain triggered the generation process only during the Neogene, when the source rocks present in the Napo Formation attained, in some depocenters, the thermal maturation conditions that were sufficient to expel the oil and gas to the Hollin and Napo reservoirs.

The presence of thick and laterally continuous sandstone beds in the Hollin and Napo formations, distributed over most of the basin permitted a good connection between the deeply buried reservoirs and the outcropping areas in both flanks of the basin. This is to say, the Brazilian Shield to the E, and the uplifted areas of the Andean Chain to the W. In consequence, massive fresh water percolation occurs in this basin, affecting directly the oil quality since a very active process of bacterial biodegradation is taking place in the accumulated oil. In this way most of the giant accumulations are to be found in the deeper parts of the basin, where this destructive process is less important, and where the reservoirs are more protected from water influx. The oil quality in these fields varies between 14° and 30° API. On the other hand, the presence of a regional aquifer in these reservoir rocks helps to maintain the pressure, allowing a very high recovery factor.

In Ecuador and northern Peru, the Oriente Basin has produced in the last 20 years more than five billion barrel of oil, from several oil fields including the giant Sushufindi and Sacha fields. Practically all of the remaining reserves in the basin, estimated at two billion barrels and the potential reserves are associated with this megasystem.

Llanos-Magdalena Basin

In Colombia, the Andean Chain is sub-divided into three branches and sedimentary basins are developed within these. The Eastern Cordillera consists of a sequence of rocks deposited from Jurassic and Early Cretaceous times, representing a rift phase that was covered subsequently by sediments related to a sag phase. In this phase Middle and Late Cretaceous marine beds and Tertiary continental sediments were deposited and subsequently deformed during the Andean Orogeny. The mountain building process of this orogeny permitted the preservation of large parts of the original basin, shaping lowland areas where large rivers came to be installed. Such rivers include the Madgalena and Cauca and their respective valleys.

In this context, the Magdalena Basin (Fig. 8) represents a large piece or relict of a huge ancestral basin deformed by the Andean Orogeny, comparable to the Oriente Basin in terms of its geological history, but completely deformed by the Andean compressional regime.

To the E of the uplifted Magdalena Basin and the Eastern Cordillera there occurs the almost undeformed part of this ancestral basin, preserved as a foreland basin and known as the Llanos Basin (Dengo and Covey, 1993), covering an area of about 200 000 km^2. The stratigraphic succession in both basins is very similar to that discussed above except that being closer to the source area of the Brazilian Shield the amount of clastic sediments is greater.

The first phase of the evolution of this basin compares with most others. However, in the Late Cretaceous and Tertiary strong and very consistent stress fields resulted in the development of a very large basin affected by folding, faulting and igneous activity. The evolution of the magmatic arc had ended by the Paleocene (Dengo and Covey, 1993).

The petroleum systems developed in this highly active tectonic environment were strongly controlled by tectonic evolution, seeing that the hydrocarbon generation and migration processes were strongly influenced by the thickening of the Middle Cretaceous shale beds, rich in organic material, and associated marlstonestone. Most of the oil accumulations are found in structural traps associated with regional thrust faults, low-angle reverse faults and Cretaceous and Tertiary detachment zones that propagated the compressional fronts throughout most of the basin.

In the foothills of the Llanos Basin of the Eastern Cordillera are found the most favourable sites for hydrocarbon accumulation, as shown by the presence of the giant fields of Cusiana and Cupiaga. In the less deformed areas, large fields such as the Caño Limon Field have been discovered. It is estimated that some 25 billion barrels of oil have been trapped in this basin.

In the Upper Magdalena Basin, oil and gas is being produced from the fluvial-alluvial sandstone beds of the Caballos Formation, time-equivalent to the Hollin Formation, as well as from the Monserrate and Guadalupe formations, consisting of fluvio-deltaic sediments of Late Cretaceous age. Representative fields of this basin are the San Francisco Field and the Dina-Tello fields, that originally contained reserves exceeding 400 million barrels of oil.

In the Middle Magdalena Basin, oil and gas were first discovered in Colombia. Nowadays there are about fifty producing oil and gas fields. The old La Cira-Infanta Field is a good representative of this oil system that originally contained about 800 million barrels of oil. In this type of accumulation, the reservoirs are found in the Tertiary sediments, consisting of sandstone and conglomerate, associated with fluvial and alluvial channel-fill environments (molasse sequence), with structural control by low-angle reverse faults.

The Cenomanian-Turonian marlstone and shale beds of marine origin provide all the hydrocarbons generated in these basins. These rocks are very rich in organic carbon derived from algae, presenting a very high conversion factor. They are included in the Villeta (Magdalena Basin) and Gacheta (Llanos Basin) formations, and are present in most areas of the basins, representing the continuation of the very same petroleum system present in Venezuela (La Luna Formation).

Cusiana and Cupiagua Fields

These fields were discovered at the end of the 1980s, and occur in the very highly deformed area in the transition zone between the Llanos Basin to the Eastern Cordillera. This area is known is the Llanos Foothills. In these fields, the traps form in very tight anticlines, associated with a basal detachment fault zone (Fig. 9). Hydrocarbon production comes from fluvio-deltaic and shallow marine sandstone units, assigned to the Guadalupe (Late Cretaceous), Barco and Mirador formations (Early Tertiary). This lies below 4000 m and locally below 5000 m.

The Guadalupe reservoirs were deposited when the relative fall of sea level permitted the deposition of two regressive-transgressive, third-order cycle sequences formed in fluvio-deltaic and offshore bars during the Coniacian and Santonian. The Barco and Mirador formations, deposited during the Paleocene and Eocene respectively, are the product of a regressive-transgressive sequence where channel-fill systems derived from the development of incised valleys were covered by transgressive estuarine sediments that rapidly evolved to a shallow sand-rich clastic platform where sand bars coalesced to form a continuous sandbody.

All of these reservoirs underwent very intense diagenesis due to the great depth of burial and the presence of tectonic conditions favourable to fluid circulation throughout the reservoir system. As a result of this process most of the reservoirs exhibit a very low porosity (normally below 10%) and permeability. However, the intense fracturing developed by several phases of tectonic activity, created a very complex system of fractures in these reservoirs, responsible for obtaining a very high production capacity; in some cases reaching values around 10 000 barrels/day per well.

The Cusiana reserves are estimated at 1.5 billion barrels of oil and 3.4 TCF of gas (2.1 barrel of oil equivalent). The Cupiagua Field was discovered later on in the same trend, but in an isolated anticline to the S (Fig.9). It contains about 500 million barrels of very light oil and condensate. Recently, some wells were drilled to investigate deeper pools in both anticlines, reaching depths exceeding 5800 m. These wells have found oil and gas in deeper reservoirs, still under evaluation, that may improve the reserves of these fields to around 3 billion barrels. The Cusiana Field came on stream recently and it is expected to reach the production peak of 500 000 barrels/day in the year 2000. To bring these fields into production it was necessary to construct a new oil pipe, crossing the very rough terrain of the Eastern Cordillera.

Caño Limon Field

The Caño Limon Field is shared by Colombia and Venezuela (Fig. 9). It was discovered in 1983, and covers and area of about 40 km². It continues to produces some 200 000 barrels/day of very good quality oil (30° API). The geology of this field is quite different to that of the Cusiana Field. It is situated in an area where the Andean Orogeny produced only gentle folds and minor faults. Due to this fact, the oil is trapped in an area of low relief and in a broad anticline, controlled by the motion of transcurrent faults active during the late Eocene and Oligocene. The lateral displacement of these faults is believed to be between 2 and 4 km (McCollough and Carver, 1990).

Deltaic sandstone pertaining to the Mirador Formation, with about 80% of the total reserve constitutes the main reservoir rocks of this field. These reservoirs are buried at a depth of 2200 m, and have an average porosity of around 25% and permeability greater than 1 Darcy. In function of these very favourable petrophysical properties, the capacity of production exceeds 20 000 barrels/day of oil per well. The low gas-oil ratio of this field is very effectively compensated by a strong and very active aquifer. The highly efficient water drive mechanism has permitted a very high recovery factor of about 60% over an original reserve estimated at 1.8 billion barrels of oil.

Maracaibo Basin

The Maracaibo Basin, in the northeastern Venezuela, covers an area of about 50 000 km², and has resulted from

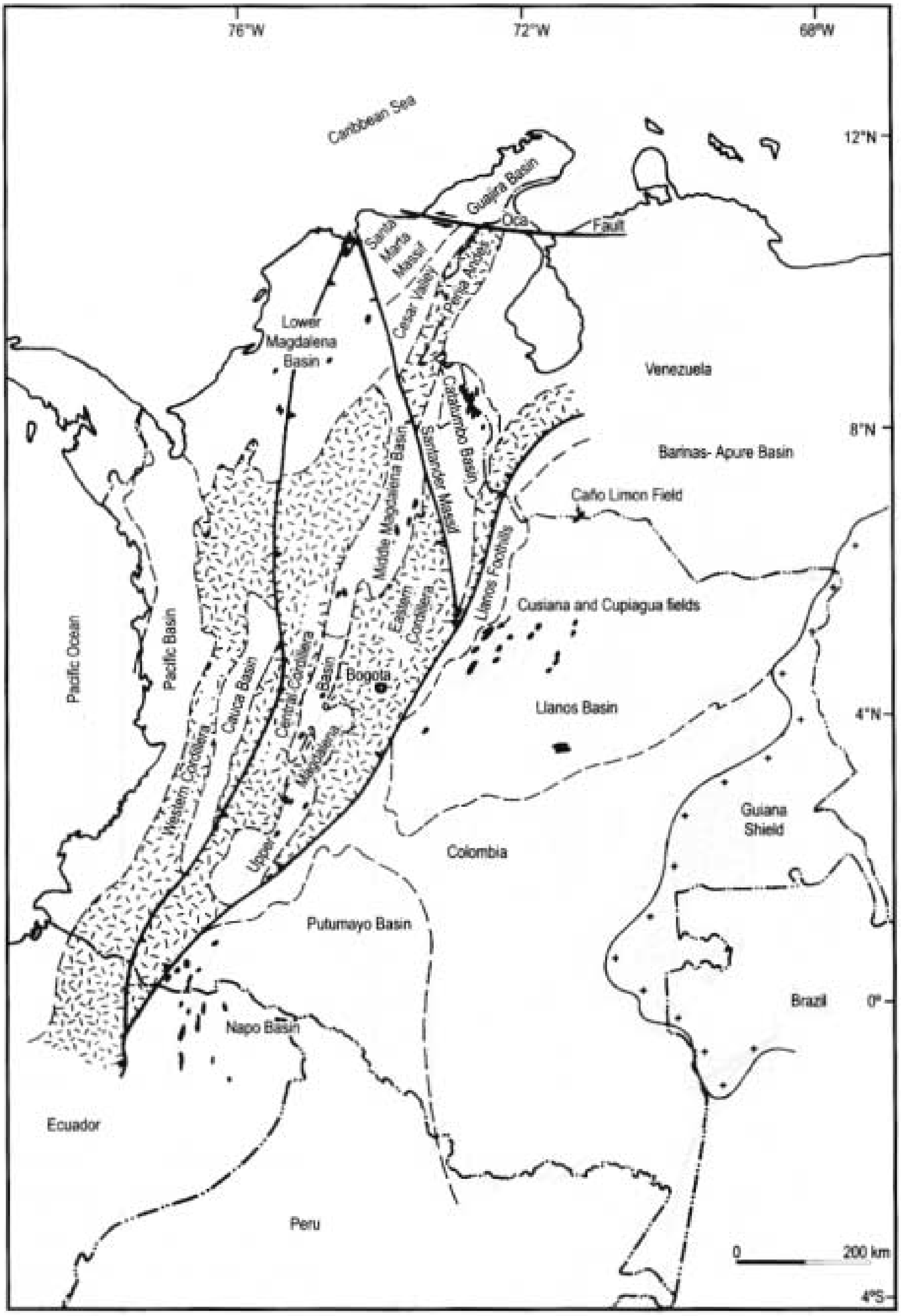


FIGURE 08 – Location map of northern portion of the Andean Foreland Petroleum Megasystem, Colombia, showing the main structural features, associated basins and oil and gas fields

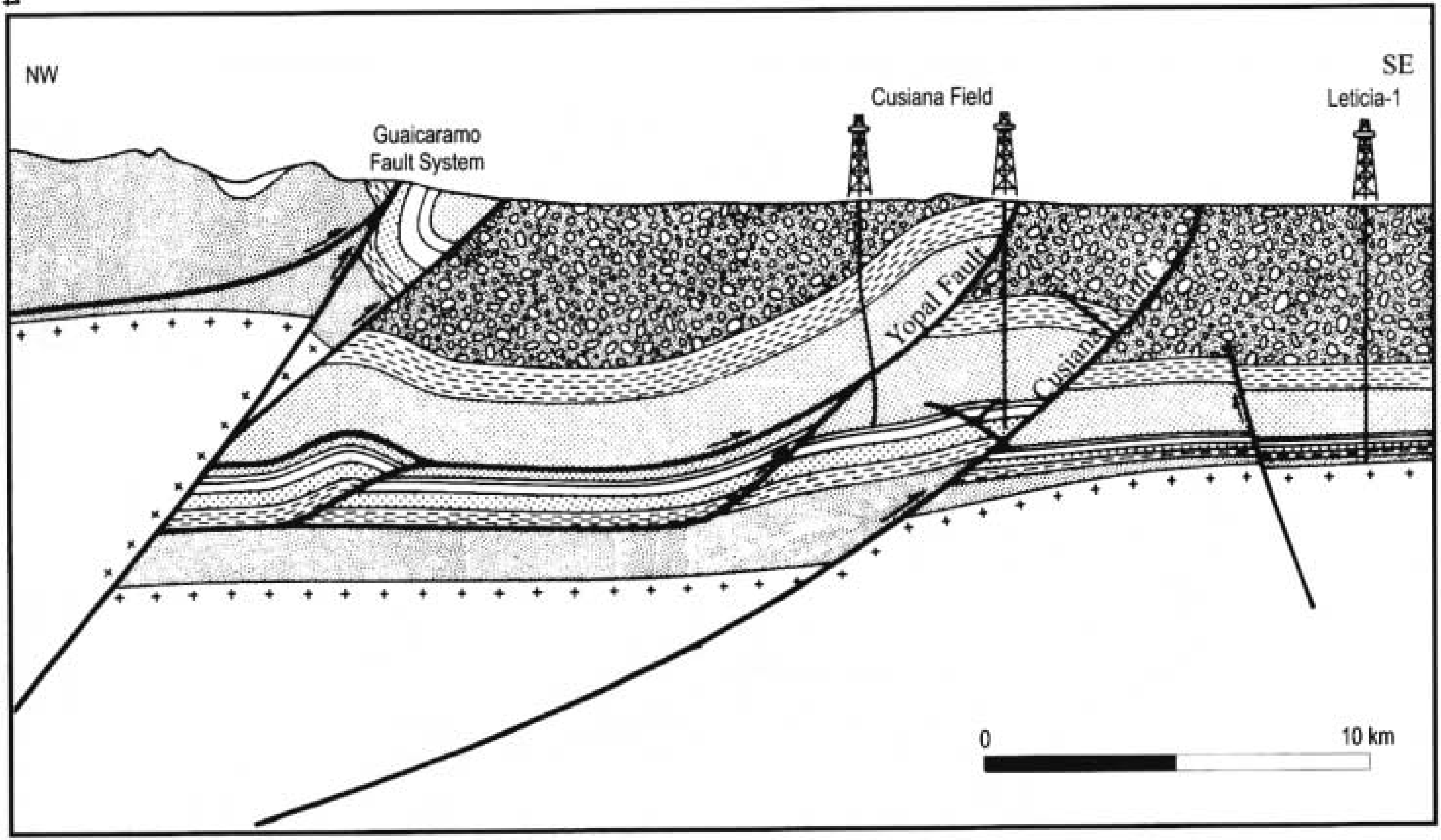


FIGURE 09 – Schematic regional cross section across the Llanos Basin, Colombia showing the Cusiana Field structural framework (modified after Cooper et al., 1995). No vertical scale.

the very complex interaction and collision of three plates: the Nazca, Caribbean and South America plates. Since the Early Paleozoic this basin has evolved from a back-arc basin, passing through a rift phase in the Jurassic; a passive margin phase during the Cretaceous and attained a foreland phase during the Paleogene (Lugo and Mann, 1995). Its present-day configuration is controlled by the Mérida Andean Chain to the E, and by the Sierra de Perijá to the W, as well as the megasuture represented by the Oca, Perijá and Boconó transcurrent fault system to the N (Fig. 10).

In the last fifty years the Maracaibo Basin has produced around 35 billion barrels of light to medium quality oil, and the remaining reserves are estimated in about ten billion barrels of oil. This type of oil accounts for almost 50% of the Venezuelan reserve. The petroleum resources of this basin are entirely related to a fault system formed by parallel transcurrent faults with sinistral motion, striking approximately N-S. This system originated by the introduction of the Caribbean Plate during the Neogene, and most of the giant accumulations such as Mene Grande, Bacachero, Lagunillas and Ceuta-Tomoporo fields, are aligned along the western margin of Lake Maracaibo, and are controlled by this tectonic event. These fields are collectively known as the Bolivar Coastal Fields (Fig.10).

The marlstonestone, shale and fine-grained carbonate beds of the La Luna Formation have a very high content of algae-derived amorphous organic matter, being the source rock of this huge amount of hydrocarbon. These rocks were deposited during the Cenomanian-Turonian anoxic event and are related to a period of maximum flooding that can be correlated worldwide.

Samples collected from immature areas have an average TOC content of about 5%, and it is estimated that around 290 million barrels of oil may be produced from each cubic kilometre of source rock (Ramírez and Marcano, 1990). This means that if the entire area that attained the required thermal conditions to produce hydrocarbons is taken into account, then the La Luna Formation may have generated more than one trillion barrels of oil to feed the traps in the Maracaibo Basin.

Bolivar Coastal Fields

This group of giant petroleum fields was discovered shortly after oil exploration in Venezuela began. Here, exploration was oriented by the presence of abundant oil seepages along the coastal areas of Lake Maracaibo. These accumulations are found in structures following a N-S trend for more than 70 km along the eastern margin of the lake and may reach until 30 km of width in lake waters (Fig. 10).

The productive section occur in Eocene deltaic sandstone beds (Trujillo and Misoa formations) and in Miocene coarser clastics of fluvial origin (Lagunillas Formation). These depositional systems were controlled by a foreland tectonic environment (Lugo and Mann, 1995). The entire sequence formed by successive intercalations of shale and sand may reach a total thickness of about 10 000 m in the main depocenter, located in the foothills of the Mérida Andes.

The trapping mechanism responsible for the huge amount of hydrocarbons is closely associated with transcurrent faults and folds related to the oblique collision of the Caribbean Plate against the South American Plate, that occurred throughout the Late Tertiary. Combined and stratigraphic traps are also present, some of them having as the top seal asphalt lakes derived from seepage and biodegradation of the oil.

More than 200 different accumulations have been found in this area, at depths varying from 170 m to 3000 m, containing oil that varies from 12° to 43° API. In the light of recent technological advances, such as 3D seismic data acquisition and processing, sequence-stratigraphy analysis, reservoir delineation and multi-lateral and horizontal drilling technologies there has been a significant increase in the recovery factor of these huge accumulations. The application of new technology to seismic surveying has brought about a boom in the exploration of deep targets including those in fractured Cretaceous carbonate of the La Luna Formation, at depths of between 4000 and 6000 m. The production mechanism usually present in these accumulations is water drive associated with gas dissolution, allowing very good recovery factors.

Maturín Basin

The Maturín Basin is part of the Eastern Venezuela Basin and is the second oil producer in this country, being active in oil exploration since the beginning of this century (Fig. 10). The basin is limited to the N by the Serrania del Interior range formed by a belt of folded and faulted rocks produced by the oblique collision of the Caribbean Plate along the El Pilar Trancurrent Zone, and to the S by the granitic terrane of the Guiana Shield.

The oil exploration began in this basin at the end of the XIX century, due to the presence of very large seepage areas occurring close to the foothills of the mountains. Commercial production began in 1913, when some very shallow oil occurrences were discovered in Tertiary reservoirs (Aymard *et al.*, 1990). In the last eighty years this basin has produced more than 1.8 billion barrels of oil (Prieto and Valdes, 1990), mainly from these shallow reservoirs. In the last fifteen years, exploration using more advanced technology resulted in the discovery of a few giant oil fields in deep reservoirs. Amongst these accumulations there may be cited the El Furrial Field, discovered in 1984 in reservoirs at a depth of 4500 m (Fig. 11).

The source rock sequence of the Maturín Basin has the same characteristics as the La Luna Formation of the Maracaibo Basin, but in the Maturín Basin these are known as the Querecual and San Antonio formations. These rocks were deposited in a bathyal marine anoxic environment, responsible for the accumulation and preservation of huge amounts of algae-derived organic matter. Recent quantitative geochemical studies indicated TOC content varying between 2% and 6%, and a potential of 56 to 252 million barrels of oil per cubic kilometre of mature rock. Therefore, taking into account the area of occurrence of this sequence, and assuming the required thermal conditions throughout, a resource exceeding two trillion barrels of oil in this basin can be estimated.

Along the southern border of the basin there occurs a very large belt of heavy oil as the result of a massive up-dip migration of oil from the deeper parts of the basin through the very permeable Tertiary reservoirs toward its outcrop areas. This belt, known as Orinoco Heavy Oil Belt, contains an estimated volume of heavy oil of about one trillion barrels (Fig. 10). The oil is stored in reservoirs with porosity values over 30% of the Las Piedras, Oficina and Merecure formations (Oligocene to Pliocene). This accumulation is sealed at the surface or at a shallow depth by asphalt lakes that have formed due to the biodegradation of the crude oil by bacterial action.

El Furrial Field

This field is situated in the tectonic domain of the Serrania del Interior range, consisting of a very complex reverse fault and fold system. The trap is an anticline associated with a very large thrust fault. It has an area of about 80 km², and vertical closure of around 900 m (Fig. 11). The main reservoir of the El Furrial Field lies below 4000 m, but it still has very good petrophysical properties including an average porosity of 15%. The reservoir rocks consist of deltaic and estuarine sandstone deposited as offshore bars and barriers bars during the relative lowering of the sea level during the Oligocene. The net pay zone occurring within the Naricual Formation is 276 m thick. In addition to these reservoirs there occur important production zones in the deeper parts of Late Cretaceous deposits.

The source rocks of the El Furrial Field are contained in the above-cited Querecual and San Antonio formations, buried deep below the Serrania del Interior. Here these rocks attained conditions favourable for the generation of huge volumes of mature and light oil and gas expelled to the shallow parts of the basin from the Miocene to the present-day.

El Furrial Field reservoir yields a good quality oil, around 29°API, and the production wells may yield over 12 000 barrels/day each. The fields do not have any oil-water contact and the main production mechanism is the pressure solution and fluid expansion in the reservoir. Because of this, a project of water injection was implemented in order to maintain the reservoir pressure to sustain the production level at about 380 000 barrels/day. The original *in situ* reserve is estimated at 6.8 billion barrels, and the remaining reserve is around 2.6 billion barrels.

South Atlantic Rift Petroleum Megasystem

In this megasystem are included a very large number of basins developed along the margin of the South Atlantic, off the coasts of Argentina, Uruguay and Brazil. All these basins have a common genesis linked to the continental break-up and drift of South America and Africa since the Late Jurassic. Continental break-up resulted in the generation of a series of linked rift basins which evolved to a restricted seaway or proto-gulf and, in most cases this opened up to create fully interconnecting marine passive margin basins, resulting in the present-day South Atlantic Ocean (Chang *et al.*, 1992).

Evolutionary events resulted in the deposition of two very important source rock sequences that are responsible for most of the oil and gas found in these basins. The older beds consist of continental shale deposited in a lacustrine environment. These lacustrine sediments may have been deposited in initially shallow depressions associated with the first phases of distension and rift formation, or related to half-graben development during the taphrogenic phase of rift evolution, when anoxic deep lakes were formed. Furthermore, the chemistry of the lake waters may have varied from fresh, as in the Recôncavo Basin, Brazil; or brackish to salty, such as in the Campos and Austral basins, in Brazil and Argentina, respectively.

The younger source rock sequence is related to the development of a proto-gulf or marine seaway at the end of the rift phase (Barremian to Aptian), when the half-grabens and shallow platforms where invaded by the sea water, interconnecting most of the basins and developing a relatively narrow seaway. This proto-gulf was 5000 km long and relatively shallow, extending from the Austral Basin off Argentina to the Sergipe-Alagoas Basin off northeastern Brazil. In this seaway, the Brazilian waters became progressively hypersaline due to the presence of a volcanic barrier situated in the present-day northern Pelotas Basin. These restricted conditions were highly favourable for the preservation of a thick sequence of black shale beds associated with a marine evaporitic suite. As a consequence, a very extensive and thick sequence of rocks with high organic matter content was deposited almost along the entire length of the South Atlantic margin of Brazil.

In this megasystem the oil derived from one of the above mentioned source rock or from both petroleum sub-systems migrated to traps and reservoirs during the rift, marine-evaporitic and passive margin sequences. Thus, giant oil accumulations developed in clastic reservoirs deposited immediately below or above the source rocks, or in reservoirs deposited during the Late Cretaceous and Tertiary and situated up to some kilometres above the source rocks. Vertical secondary migration is proposed to explain this fact, and very effective salt tectonics have played an important role in the development of this petroleum sub-systems. The Austral, Recôncavo and Campos basins have sub-systems that serve to illustrate the megasystem.

Austral Basin

The Austral Basin, also known as the Magallanes Basin, is situated in the extreme S of the South American Plate. It is limited to the W by the Andean Cordillera, and extends out into the Argentinean Atlantic Ocean as far as the Rio Chico High (Fig. 5). It covers an area of about 170 000 km² of which almost 23 000 km² are on the continental platform.

This basin developed during two distinctive rift phases in the Middle and Late Jurassic, respectively, and terminated with the extrusion of a very thick sequence of volcanic material known as the Tobífera Series of Early Jurassic age. In this phase the crustal opening diminished in intensity, and the basin was covered by continental clastics, filling shallow depressions associated with half-grabens. Following this terminal rift phase, a sag phase occurred during the Early Cretaceous, and typical transgressive sequences were deposited, known as the Springhill Formation, during the Early Valangian and Hauterivian. In Middle and Late Cretaceous the Austral Basin evolved to a special or discrete type of foreland basin, in part associated with the Patagonian and Andean tectonism (Robiano *et al.*, 1996).

In this basin, the source rocks in the Early Cretaceous

FIGURE 10 – Location Map of the Caribbean Foreland Petroleum Megasystem, Venezuela, showing the main structural features, associated basins and oil and gas fields.

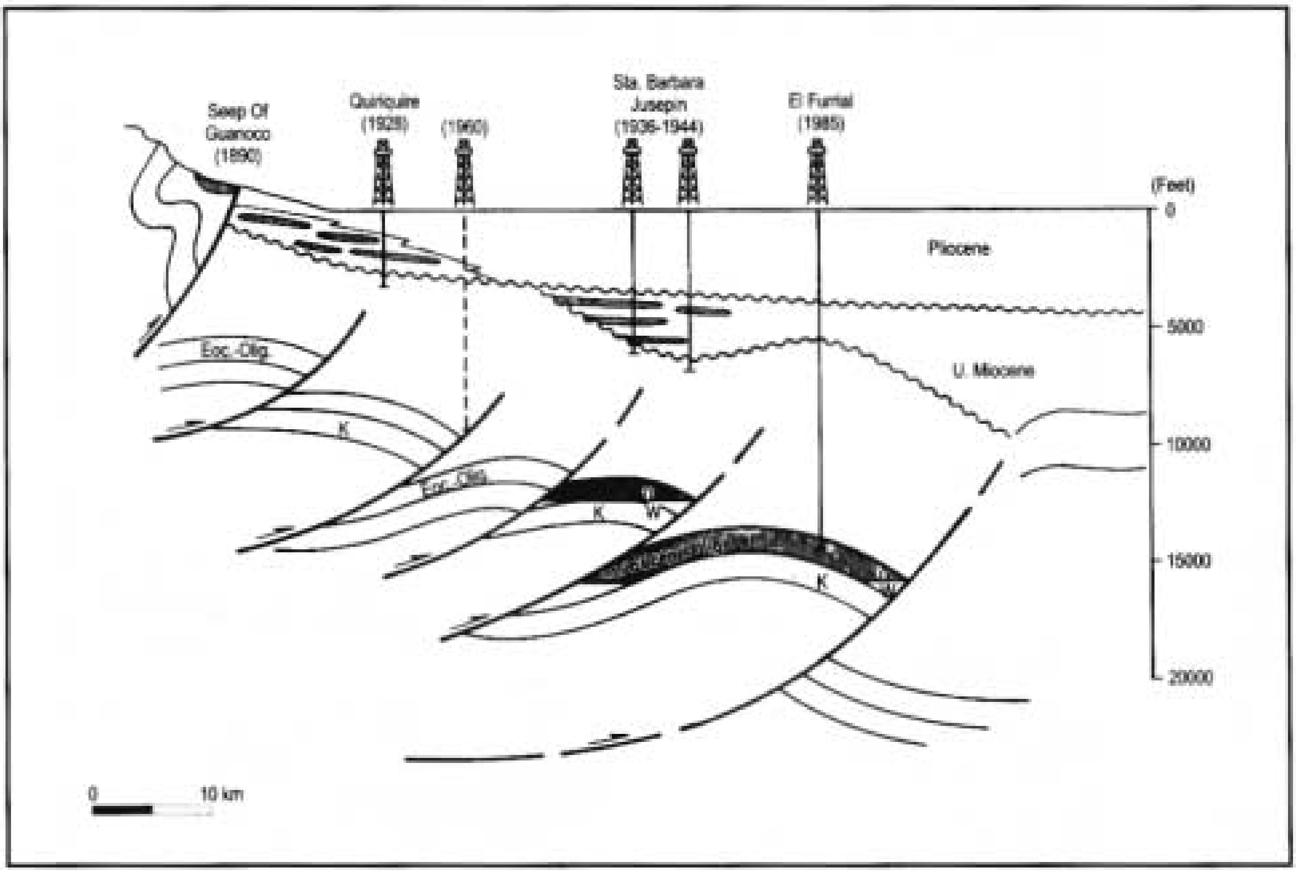


FIGURE 11 – Schematic regional cross-section across de Maturin Sub-basin, Venezuela, showing the El Furrial Field structural framework (modified after Prieto and Valdés, 1990).

sedimentary section (*Inoceramus*/Springhill formations), where basal continental shale beds that grade to beds of marine marlstone and shale with encroachment of the sea. As a result, most of the source rocks show a strong continental contribution and influence, and the organic matter present in these rocks has a mixed origin.

In the western part of the basin these rocks are deeply buried and display thermal conditions favourable for the production of gas. On account of this, the resources found in this basin are composed of 70% gas and 30% mature oil. The fluvial and shallow marine sandstone units of the Springhill Formation form the main reservoirs of this basin. The source rocks directly cover these reservoirs, and in this sense only a very short vertical distance was required to introduce the oil and gas.

Structural or combined features constitute most of traps present in the basin. Normal faulted blocks associated with the final phase of rift evolution produced the structural features and the combined features were associated with the onlap of the coastal sandstone beds, deposited during a regressive phase. During the Tertiary, the basin was affected by extensional and transpressive movements originated by the Andean Orogeny, forming small faulted blocks and pull-apart basins of little economic interest.

A conspicuous characteristic of this basin is the very large lateral migration. As previously stated, the source rock covers an excellent carrier bed (Springhill Formation reservoirs) and itself was acting as a very effective seal. This combination allowed the oil to migrate a very long distance, charging the existent traps or reaching the surface. Pittion and Arbe (1997) studies indicated that, in some areas, oil migration reached about 200 km.

Only a small number of hydrocarbon accumulations found in the basin were developed, since most of the discovers made were gas pools, some with an oil ring at the base. The Hidra Field has produced around 46 million barrels of oil. The gas is sold in Buenos Aires. The proven gas reserves already found and certified in the Austral Basin exceed 100 billion cubic metres.

Recôncavo Basin

This basin is situated in northeastern Brazil and occupies an area of 11 500 km² in the State of Bahia. In a regional tectonic context this basin represents the southern part of an aulacogen developed from a triple junction situated in the position of the City of Salvador. This branch of an aborted rift was isolated at the end of the Cretaceous from the other two branches that gave origin to the Atlantic Ocean (Fig.12).

According to Szatmari *et al.* (1985), the Late Jurassic-Early Cretaceous stress field, influenced by pre-existing weakness zones in the basement, created a small crustal block known as East Brazilian Microplate (EBM). The independent counterclockwise movement of this microplate in relation to the nascent South American Plate permitted the development of an intracontinental rift in this area. In the Aptian, the movement of the microplate ceased and, as

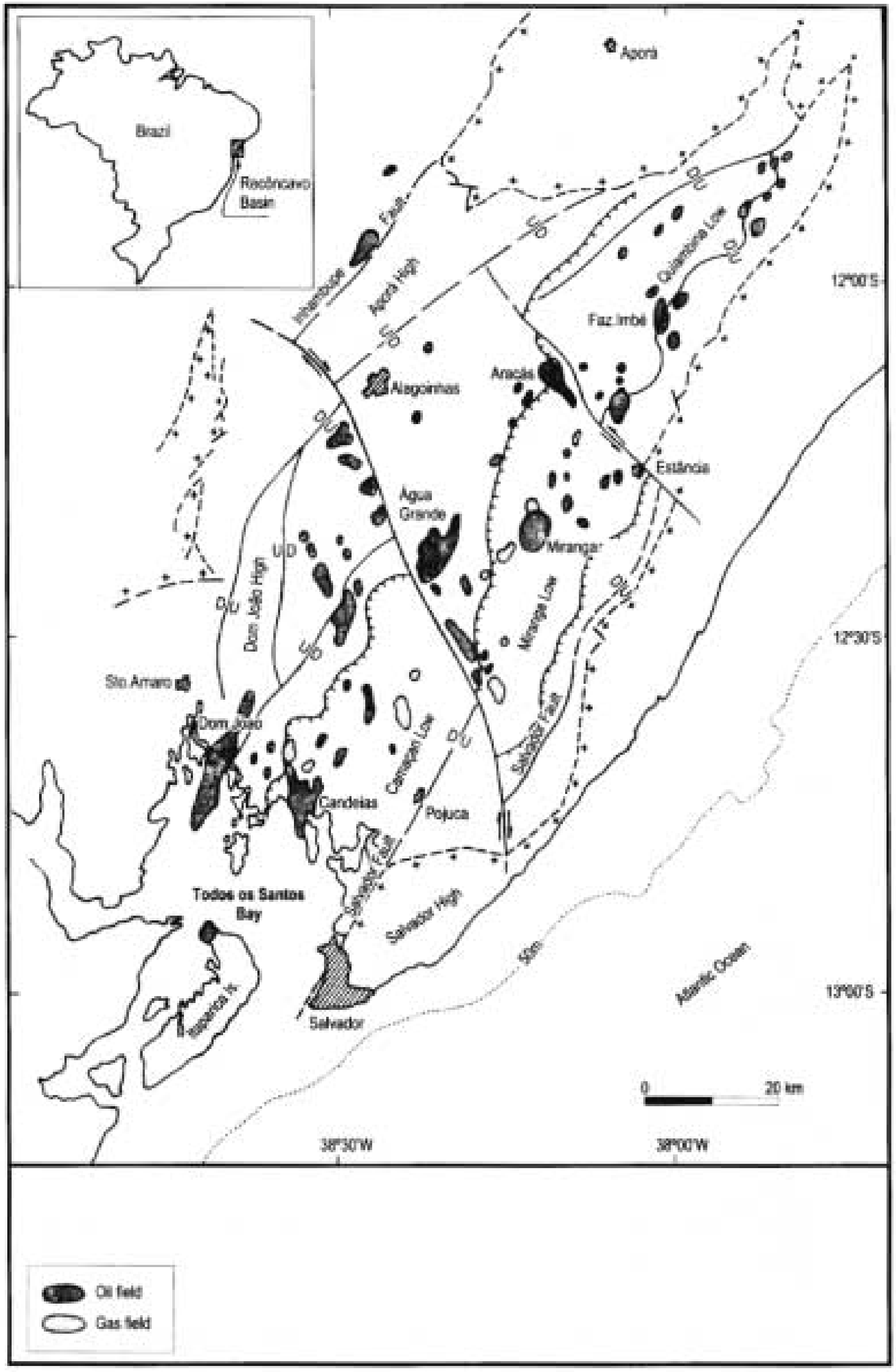


FIGURE 12 – Recôncavo Basin map, onshore northeast Brazil, showing the main structural features and oil and gas fields (modified after Figueiredo et al., 1994).

a consequence, the interior rift branch was abandoned.

In the Recôncavo Basin the main source rocks are fresh water lacustrine shale beds. When the rift evolution started there developed shallow lakes in a depression formed by the gentle extension of the crust, resulting in the generation of extensive argillaceous deposits with organic matter that covered almost all the basin. Subsequently, with the paroxysm of the rifting event, very deep (1000 m) lakes were formed (half-grabens) and a thick sequence of shale beds, rich in organic matter, was deposited and preserved in anoxic conditions confined to these depocenters. The first type of source rock is found in the Tauá Formation; whereas, the second is contained in the Gomo Formation.

Both source rock-types have an average organic matter content (TOC) below 2%, and residual generation potential between 5 and 10 kg HC/ton of rock. These relatively low values are interpreted as being related to the very high convertibility of the original organic matter into hydrocarbons, already expelled to the traps.

The main reservoirs of the basin are found in the pre-rift section (Late Jurassic) and in the syn-rift section (Early Cretaceous). The pre-rift reservoirs consist of alluvial-fluvial sandstone of the Sergi and Água Grande formations that floor most of the basin and have a uniform thickness of about 150 m. Overlying these sands there occur well-developed eolian beds presenting favourable petrophysical properties as reservoir-rocks.

The syn-rift reservoirs are distributed along the entire section of the basin fill. In the main depocenters the deep lakes were totally filled by a package of shale beds (source rocks) and sandy sediments. These sediments were deposited by high and low-density density turbidity currents, locally presenting good reservoir quality (Candeias/Maracangalha formations). During this phase, the tectonic control over deposition prevailed, and in the active eastern border conglomerate units were stacked resulting in the deposition of a wedge, locally up to 2000 m thick (Salvador Formation).

Following this first tectonic pulse there occurred a more quiescent phase. The progressive filling of the basin permitted the rapid advance of sand-rich fluvio-deltaic fronts over the shallow basin, responsible for the development of laterally continuous reservoirs during the syn-rift phase (Marfim/Pojuca formations and Catu Member). During the final phase of rifting, the basin subsided slowly and was filled by coarse-grained sandstone and shale beds with a very high sand/shale ratio, deposited under fluvial conditions (São Sebastião Formation).

The main trappping mechanisms in this rift basin are related to faulted blocks. These structures are related to basement-involved normal faults affecting the pre-rift and syn-rift sediments by the development of horst and block-faults. They may also be related to the development of shale diapirs and listric normal faults detached in the Candeias Formation, with associated rollover anticlines, affecting only the syn-rift section. Associated with the depocenters, and controlled by the half-graben, there also occur combined and stratigraphic traps, affecting mainly the reservoirs formed by turbidite beds.

For more than sixty years the Recôncavo Basin has produced more than 6 billion barrels of very good quality waxy and paraffin-rich oil (30° to 40° API) and more than 100 billion cubic metres of gas. Remaining reserves are estimated at about 300 million barrels of oil and 44 billion cubic metres of gas.

Campos Basin

The Campos Basin is situated on the offshore part of the State of Rio de Janeiro, southeastern Brazil, close to the state capital of Rio de Janeiro (Fig. 13). The Campos Basin is the most productive oil basin along the entire American side of the Atlantic Ocean, and it contain about 80% of the oil found and produced in Brazil. The basin covers an area of about 100 000 km^2 extending from the coastline to the 3000 m isobath. It is limited to the N and to S by shallow water basement highs separating the basin from the Espírito Santo Basin and Santos Basin, respectively. Both these basins are oil and gas producers. In the deep water domain there is no physical separation among these basins.

This basin is an excellent representative of the so-called Atlantic passive margin basins, present along the eastern coast of South America and western coast of Africa. The basin has developed on an extrusive volcanic substratum that resulted from the final phase of the break-up of Pangea (Cabiúnas Formation). This unit is present all over the southern submarine part of South America and extends inland over the Paraná Basin where the basalt sheets may exceed 2000m (Serra Geral Formation).

The evolution of the basin started with a rifting episode in the Early Cretaceous, causing the development of relatively shallow half-grabens where continental lacustrine sediments were deposited in fresh to brackish water. These lacustrine sediments consist of coarse-grained basal conglomerate containing basalt fragments, overlain by fine-grained siliciclastic sediments deposited in an alkaline environment. These siliciclastic sediments include talc-stevensite mudstone with very high organic matter content (Lagoa Feia Formation). Also present in this thick rift sequence are carbonate banks built of pelecypod and ostracod shells (coquina).

In the Barremian to Aptian transition, a strong erosional angular unconformity peneplained the basin, and over this flat surface there occurred the first marine transgression from the S. Once a more continuous water body along the Brazilian continental margin (proto-gulf or seaway) had developed, the prevalent arid conditions and the presence of a barrier to the S permitted the deposition of a thick sequence of evaporite beds. These evaporite beds consist mainly of cyclic deposits of halite and anhydrite and, in the present day deep waters of Santos and Campos basins, is estimated the existence of an original thickness of almost 2000 meters of halite (Retiro Member).

Continuous plate drifting and deepening due to active thermal subsidence enlarged the seaway. During the Albian, an extensive ramp-type carbonate platform, more than 600 m thick, developed in the shallower areas. This ramp consists mainly of grainstone (oolitic and pisolitic banks) grading to mudstone and marlstonestone in the deeper parts. During the Cenomanian to the Campanian times this high energy carbonate ramp was flooded by a continuous sea level rise and deep-water carbonate facies and

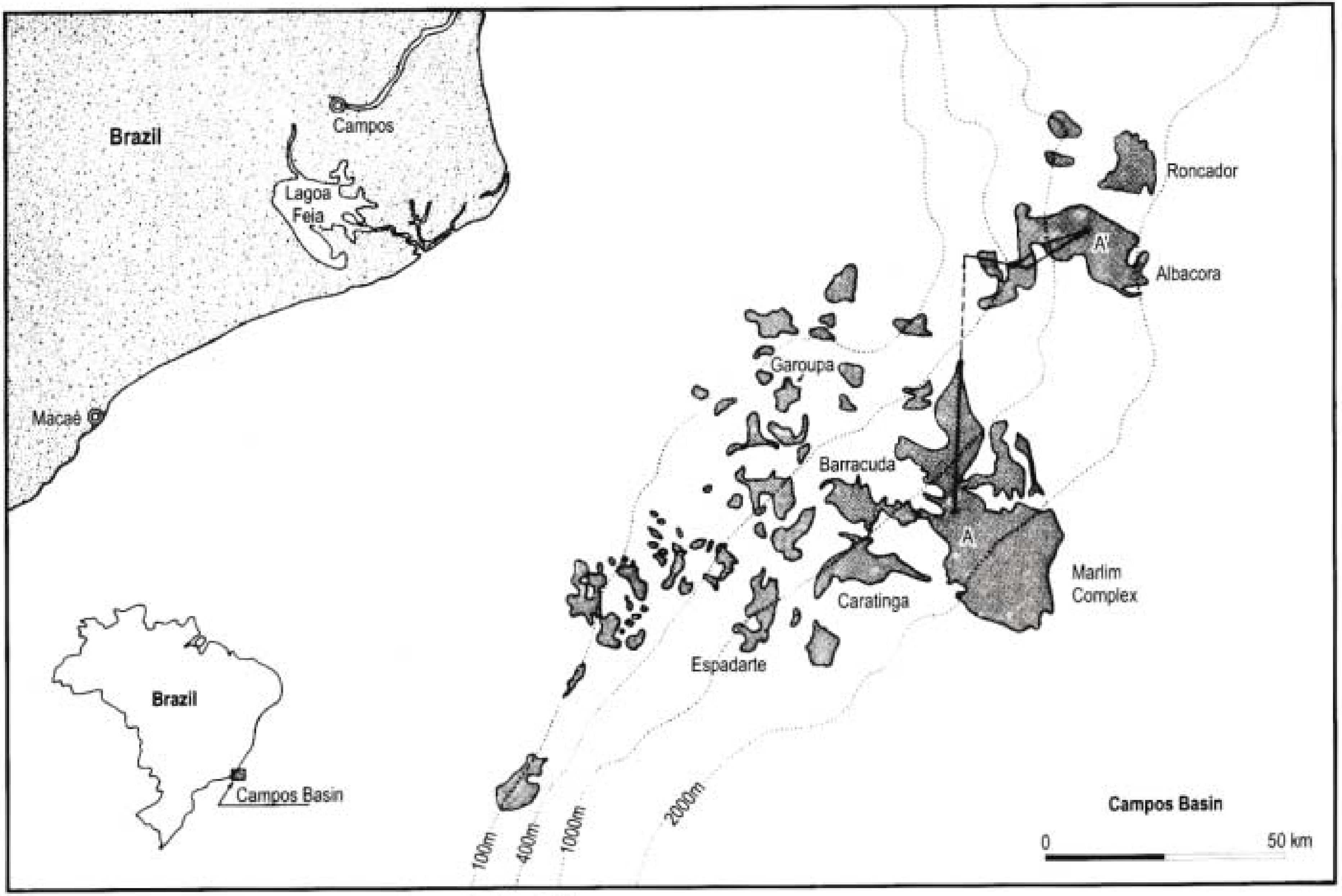


FIGURE 13 – Campos Basin map, offshore southeastern Brazil, showing the oil and gas fields.

marlstone beds (Macaé Formation) sealed the grainstone banks.

During the development of this marine carbonate environment, siliciclastic rocks and turbidite beds were also deposited, reaching the low areas between banks and carried out to the deep basin by density currents. These coarse to fine-grained sandstone beds are known as Namorado Sandstone.

The rapid subsidence of the basin during the Late Cretaceous and Tertiary was responsible for the accumulation of more than 3000 m of sediments. This sequence consists of siliciclastic and carbonate beds associated with the growth of the continental margin, and the development of a prograding complex consisting of shallow platform and deep slope to basin sediments. These sediments consist of coastal fluvial and shallow marine sandstone and shale beds, carbonate bank sediments and slope and basin floor shale and marlstone. Intercalated with the fine-grained sediments deposited on the slope and in the deeper parts of the basin, there occur zones of coarse clastic sediments resulting from periods of a relative fall in sea level and the consequent input to the deep basin of platform sandstone by density and turbidity flows. These flows were able to transport large amounts of clastic sediments to the basin floor, giving rise to thick turbidite fans, locally covering areas exceeding 200 km^2 (Carapebus Formation).

The carbonate and clastic beds of the Middle to Late Cretaceous and Tertiary section were strongly affected by intense salt tectonics, with the development of salt pillows and domes, as well as the development of listric faults with detachment surfaces in the evaporite section. Salt tectonics is responsible not only for the traps for hydrocarbon accumulation but also for the distribution of most of the turbidite units, as the bottom topography of the basin was totally affected by the movement of salt.

The Early Cretaceous lacustrine shale of Lagoa Feia Formation deposited in a shallow rift in which the environment was alkaline represents the main source rock of the basin. This Barremian sequence displays total organic carbon contents of about 4 to 6%, amorphous organic matter (type I), and yielding capacity of around 38 kg of hydrocarbon for ton of rock (Guardado *et al.*, 1997).

Biomarker data of this source rock suggest deposition in a brackish to salty water lacustrine environment, developed in shallow depressions of the rift phase. These rocks occur very extensively all over the basin, and reach a thickness varying between 100 and 300 m. The oil window was reached during the Middle Cretaceous, but only in the Miocene the maturation peak was attained. Most of the source rock sequence is still in the oil window, and only in a few deep grabens these rocks penetrated the gas window (Mello *et al.*, 1994). As a consequence the Campos Basin has a low gas/oil ratio, and the oil quality is relatively poor, varying from 14° to 32° API.

The reservoirs in this basin occur almost throughout the stratigraphic column, from the fractured basalt and coquina in the rift section; the oolitic carbonate of the shallow platform; and the turbidite beds of the deep siliciclastic basin. The main giant oil fields occur in clastic reservoirs formed from Tertiary turbidite beds.

In the rift package the oilfields are situated in internal horsts where fractured basalt and thick banks of coquina are present. These lacustrine carbonates have a porosity in the order of 15% to 20% and medium permeability values. The complex Badejo-Linguado-Trilha and Pampo fields, situated in the Badejo High, southern Campos Basin, produce oil from this interval. The original *in situ* volume is given as 660 million barrels.

The main producing Albian reservoir consists of marine oolitic carbonate. In this reservoir more than five billion barrels *in situ* original oil volume was found in a N-S oriented narrow belt of bars structured by halokinesis. The petrophysical characteristics of these carbonate banks are excellent, but may vary very rapidly in response to facies change to low-energy carbonate beds (lagoonal and open marine facies).

In the Late Albian to Turonian deep-water sequence, there occur thick units of turbiditic sandstone, associated with a phase of trangression and flooding of the basin. These turbidites known as Namorado Sandstone display a typical channeled geometry and were also strongly structured by the initial phase of salt movements. Due to the gravitational tectonics, the original low areas that captured the sand-rich density flows were completely reversed to high blocks forming natural traps to oil accumulation. The Namorado Field containing more than 120 million barrels of proven original oil, constitutes the typical oil field in this type of reservoir. It is estimated that in this petroleum sub-system more than two billion barrels of oil were accumulated *in situ*.

However the giant fields are contained in the turbiditic Tertiary section known as Carapebus Member, where more than 30 billion barrels of oil *in situ* were discovered. The turbiditic fans were formed from the Late Cretaceous to the Miocene, but the huge volume of sandstone, deposited as basin floor fans reworked by deep tidal currents are concentrated in the late Oligocene and early Miocene. This vast influx of sand in the deep basin is associated with rapid but violent sea level falls related to climatic change, creating conditions favourable for the transport of very large volumes of clean platform sands, via slope channels and, in some case, via canyons to the deep basin floor. The basin floor fans consist essentially of coalescing sandbodies up to 120 m thick and covering areas of several hundreds km^2. They have remarkable laterally and vertically continuity, with porosity values around 30% and permeability of several Darcies. These sandbodies may produce around 20 000 barrels/day per well.

In these Tertiary reservoirs the main trapping mechanism is the pinchout of the sand against the fine-grained sediments of the slope. However, halokinesis also plays an important role, since the listric normal faults generated in the salt section are responsible for the migration routes that allow the oil to reach the post salt reservoirs coming from the deep parts of the rift basin.

The migration pathway model to explain the distribution of a so large a volume of hydrocarbons in the Campos Basin was controlled by a number of factors. These controls include the presence of rift paleo-highs responsible for collecting oil from the adjacent lows; salt displacement to open window for oil ascension through the excellent seal represented by the evaporite beds; listric growth faults acting as conduits for the hydrocarbons; and finally, by the carrier beds represented by the laterally continuous carbonate and turbidite reservoirs.

Accordingly Bruhn (1998, 1999) the turbidite reservoirs *contain about 71% of the oil and 60% of the gas found in* the Brazilian basins, being responsible for 89% of the proven reserves of oil and 54% of gas reserves, estimated in 14.1 billion barrels of oil and 15.4 TCF, respectively.

Albacora Field

The Albacora Field was the first discovery in the deep waters of Campos Basin, Brazil, at the end of 1984, and it lies at water depths varying from 250 m to 2000 m. The field contains six stacked reservoirs consisting of turbidite beds, the oldest deposited in the Early Cretaceous and the youngest deposited during the Tertiary (Eocene, Oligocene and Miocene). These reservoirs lie between 2300 and 3300 m below sea level.

The field complex covers an area of about 235 km^2 and contains an *in situ* oil volume of 4.5 billion barrels of oil and 68 billion cubic metres of gas. The deepest and oldest reservoir consists of the Namorado Sandstone (Albian-Cenomanian) lying in a horst-type trap developed by salt tectonics. The other five reservoirs are combined or stratigraphic traps, related to depositional characteristics of turbidite basin-floor fans, gently accommodated over the older structures (Fig. 14).

The total thickness of the oil-bearing sandstone may reach 110 m and the porosity varies from 17% in the older body to 30% in the Miocene sandstone. Permeability values are extremely high, in the range of 1 to 3.5 Darcies, mainly in the Oligocene sandbody, where most of the oil is trapped. The Miocene reservoir extends to very deep waters and has a very large gas cap.

The oil quality is variable in function of the reservoir depth and biodegradation processes suffered by the younger reservoirs due to their maturation low temperatures. In the Namorado Sandstone, below 3000 m, the oil presents values around 30° API, whereas in the shallower Miocene reservoirs the oil presents only 17° API (Candido and Corá, 1990).

This field is being developed since 1987, when an early production system was installed to drain and produce the shallower water reservoirs (Namorado Sandstone and Eocene beds). Only recently, a more complete development system was installed, being composed of several floating production storage and offloading (FPSO) barges and semi-submersible platforms, with capability to produce oil and gas to water depths of 1200 metres. The Miocene oil and gas reservoir, located beyond these depths will be subsequently put into production (Albacora Leste Project). The production peak of this system is estimated to reach 320 000 barrels and 5.4 million cubic metres per day of associated gas.

Marlim Field

Marlim Field is situated in the deep waters of Campos Basin, Brazil. It was discovered in 1985 by the drilling of a wildcat in a water depth of 835 m. This well, a world record for the petroleum industry at that time, had the purpose of testing a stratigraphic trap mapped and delineated on the basis of an anomaly detected on 2D seismic lines.

This seismic anomaly covering an area of about 140 km^2, was mapped in the Oligocene section and related to the presence of turbidite bodies already known in the basin *as potential producers of oil and gas in shallow water.* At about 2600 m the drill intersected massive oil-bearing sandstone some 73 m thick, displaying a very good correlation with the seismic anomaly. The production test showed a production capacity around 3800 barrels per day of oil of 19° API.

Contrary to the previously discussed Albacora Field, the Marlim Field only has one reservoir of exceptional size and lateral and vertical continuity, covering an area of about 152 km^2 (Fig. 14), and lying at water depths between 600 and 1200 m. Subsequently, several other petroleum accumulations were discovered close to the Marlim Field, in other Oligocene basin-floor fans, and are collectively known as Marlim Complex. This complex (Fig. 14) covers an area of about 350 km^2 and contains around 14 billion barrels of oil *in situ*.

The Marlim reservoir consists of several stacked and coalescing deep water sand-rich non-confined fan lobes, deposited rapidly during the late Oligocene, and controlled by a relative fall in sea level. The amalgamation of these fan lobes produced a very massive and homogeneous medium to fine-grained sandstone, with an average thickness of 47 m, an average porosity of 25% and permeability ranging from 1.2 to 5.4 Darcies.

The absence of fine-grained beds, usually present at the very top of each turbiditic event, is being tentatively interpreted as the result of strong and deep submarine tidal currents, known as contour currents. The presence of this type of currents would be responsible for the removal of the argillaceous material from the section and consequent cleaning of the turbidite deposits.

The main trapping mechanism acting in the Marlim Field may be described as a combined one, since to the W, N, and S there is a sand pinchout, and to the E the accumulation is terminated against a listric fault plane (Fig. 14). This fault provided the main migration pathway for oil ascension, from the source rocks in the pre-salt rift section.

Although the Marlin Field reservoir consists of a single and continuous sandstone unit the oil produced from this reservoir may vary in quality from 19° to 26° API. This variation is related to the presence of, at least, two oil migrations and filling pulses. The first one occurred probably in the late Oligocene, which is very early in the reservoir history, and may be due to the shallow burial and low temperature caused the partial alteration of the incoming oil by bacterial degradation. The second pulse occurred in the late Miocene and recharged the reservoir with lighter and more mature oil, preserved by the deeper and higher temperature conditions reached by the reservoir. The final product is an accumulation with a mixture of oils. In some places there is prevalence of one type (heavier) with respect to another (lighter), in function of some aspect of the migration patterns and pathways that is not completely understood.

The Marlim Field went on stream in 1992, through an early production system installed in the northern part of the field. Here the water depth is approximately 600 m, which permitted a better knowledge of the reservoir production capacity and pressure maintenance. This was important because only a very small amount of water had been detected in the upper parts of the reservoir. This

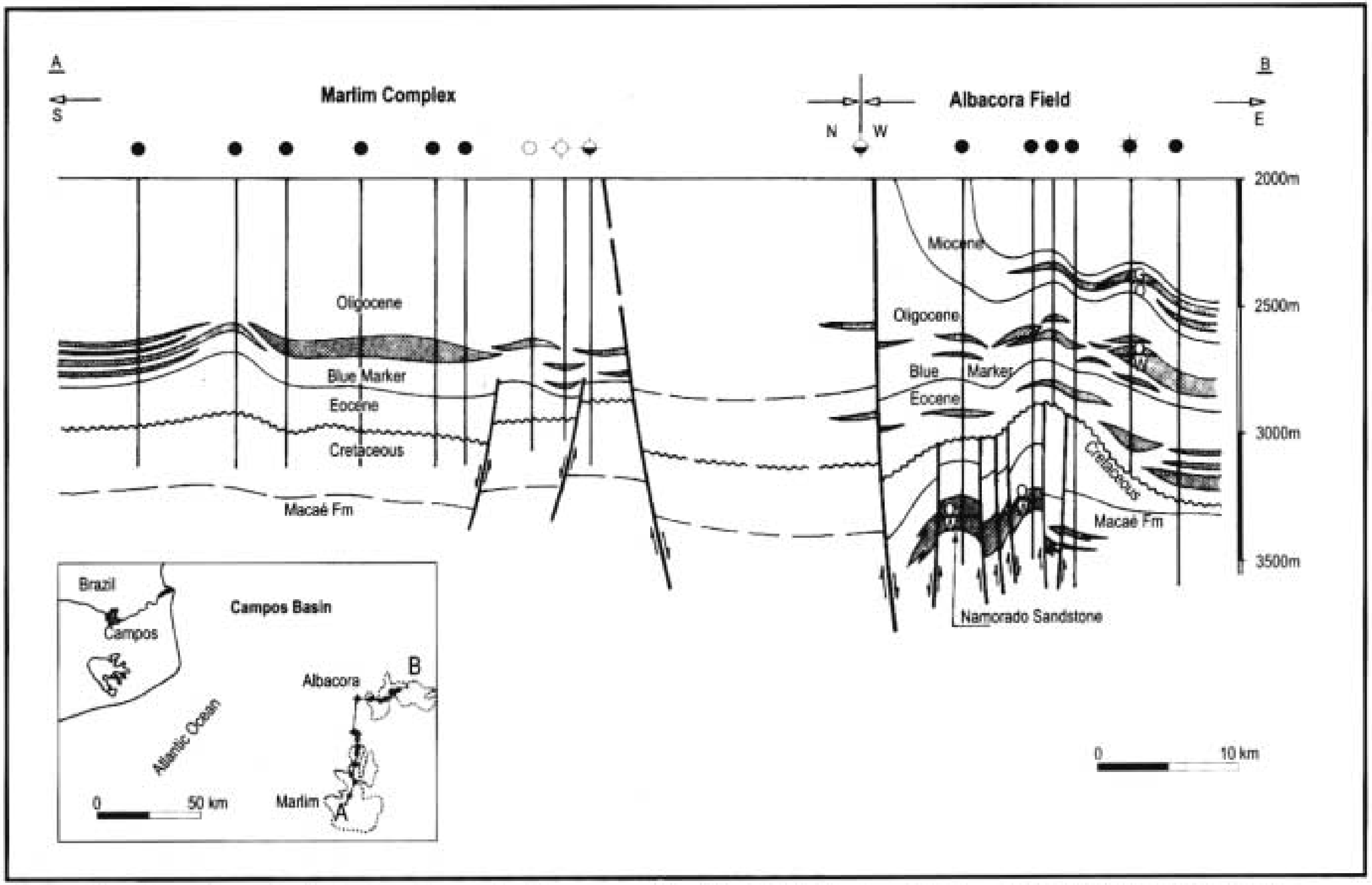


FIGURE 14 - Schematic structural cross section showing the tectonic framework and deep-water reservoir distribution of the Marlin Complex and the Albacora Field (modified after Candido and Cora, 1990).

aquifer is probably not active, and pressure maintenance measures have been required since production started.

The development project of the Marlstoneim Field will be almost complete by the end of 1999. It will consist of about 100 producing wells and 30 water injector wells drilled from six semi-submersible floating platforms (FPSO). A production peak of about 500 000 barrel of oil per day is estimated from this system. The part of the Marlstoneim Complex, lying in deeper water will be in production in the next few years.

Intracratonic Petroleum Megasystem

The Intracratonic Petroleum Megasystem is developed in the Paleozoic interior sag basins of South America. These huge, elliptical to circular shaped basins, with areas ranging between 500 000 to more than 1 000 000 km^2, share some general characteristics including the remarkable presence of large volumes of Mesozoic basaltic rocks and a relatively simple stratigraphy and structural framework (Figure 15).

On the other hand, a very complex and unusual petroleum geology should be included in this list of similarities, and it was certainly responsible for the slow response to the exploration efforts performed in these basins since the early fifties.

The South American Paleozoic syneclises were filled dominantly by siliciclastic sequences, with a notable exception represented by the Late Carboniferous-Early Permian carbonate-evaporite section present in the Solimões, Amazonas and Parnaíba basins, and to a lesser extend by the Late Permian carbonate-shale rhythmic beds of the Paraná Basin. Also remarkable was the glacial influence over the sedimentation, that was extreme during the Late Carboniferous in the Paraná Basin. The tectono-stratigraphic development of these basins started in the Late Ordovician and reached the Cretaceous (Paraná and Parnaíba), or proceeded up to the Quaternary (Solimões and Amazonas), being their stratigraphic record a succession of large-scale cratonic sequences separated by interregional unconformity surfaces.

In all these basins, the most effective source rock interval is located in the Devonian, represented by marine black shale with ages varying from Frasnian (Solimões, Amazonas and Parnaíba) to Emsian (Paraná). The reservoirs are essentially formed by sandstone units accumulated in different environments, with ages from the Devonian (Itaim and Cabeças formations in Parnaíba Basin), through the Carboniferous (Juruá and Monte Alegre formations in Solimões and Amazonas basins), to the Permian (Rio Bonito Formation in Paraná Basin). Seal is provided by thick evaporite sections in the Solimões Basin, but a Mesozoic diabase sill retained the gas in the Barra Bonita Field of the Paraná Basin.

The main trapping mechanism is structural. Mesozoic fault-propagation anticlines associated with contractional faults hold the majority of the oil and gas reserves in the Solimões Basin. Permian transpressional folds are responsible for gas accumulations in the Paraná Basin. Several minor occurrences and small oil accumulations of stratigraphic nature were discovered in the Devonian reservoirs of the Solimões and Amazonas basins and in the Permian reservoirs of the Paraná Basin (Milani and Zalán, 1998, 1999). Of these, the Solimões Basin holds most of the reserves of hydrocarbons related to the Intracratonic Petroleum Megasystem.

But the main aspect that makes the Intracratonic Petroleum Megasystem of South American basins unique is how the maturation stage of the Devonian source rocks was attained. In none of these basins, can a simple model for maturation evolution, based solely on subsidence and uplift rates and thermal input derived from crustal stretching, explain any of the observed accumulations of petroleum and associated maturation indicators.

In all these basins, the introduction of thick Mesozoic diabase had enormous thermal effects upon the maturation of the organic matter, on the level of such maturation, and also on the transformation ratio of previously accumulated hydrocarbons. In some of these basins, maturation of the organic matter was achieved only by the additional thermal input of intruding diabase into the source rocks. In others, the oil-window level of maturation was suddenly increased to the gas-window level by these intrusion. An excellent example of this type of megasystem is found in the Solimões Basin, Brazil.

Solimões Basin

The petroleum system of the Solimões Basin is formed by Late Devonian source rocks of the Jandiatuba Formation, attaining a thickness of 40 m and reaching a TOC mean value of 4% (Eiras, 1998). Vitrinite reflectance is above 1.00% all over the basin. The Late Carboniferous eolian, tidal plain and shallow marine sandstone beds of the Juruá Formation are the best reservoir in the basin. The sandy package may be as thick as 40 m, displaying porosity values around 18% and good permeability mainly in the eolian facies. Efficient seal is provided by evaporites found at the base of the Pennsylvanian Carauari Formation.

Very extensive magmatism occurred during Late Triassic to Early Jurassic and that was the time when most of the generation, expulsion, migration and accumulation of petroleum took place in the Solimões Basin. Three main, basin-scale sills were intruded roughly parallel to argillaceous bedding planes, serving as stratigraphic reference levels. Fault-propagating folds, the classic trapping style in the basin, were formed by dextral wrenching during Late Jurassic to Early Cretaceous times (Figure 16).

Pioneer exploration activities in the Solimões Basin, an area located in the middle of the Amazon jungle, began in the late 50's, but the main cycle of discoveries of hydrocarbons only started in the late 70's. Most of hydrocarbons found in this basin are composed of gas, but in several pools an oil ring is present at the base of the accumulation. This light oil and the associated liquids formed with gas (condensate) started to be produced in the 80's, after the design and construction of a complex production system in order to minimize the environmental impact of this activity in this type of tropical rainforest.

Up to now, four oil and gas provinces have been discovered in the basin: Juruá and Copacá (10 gas fields), Urucu (5 light oil/gas/condensate fields), and São Mateus

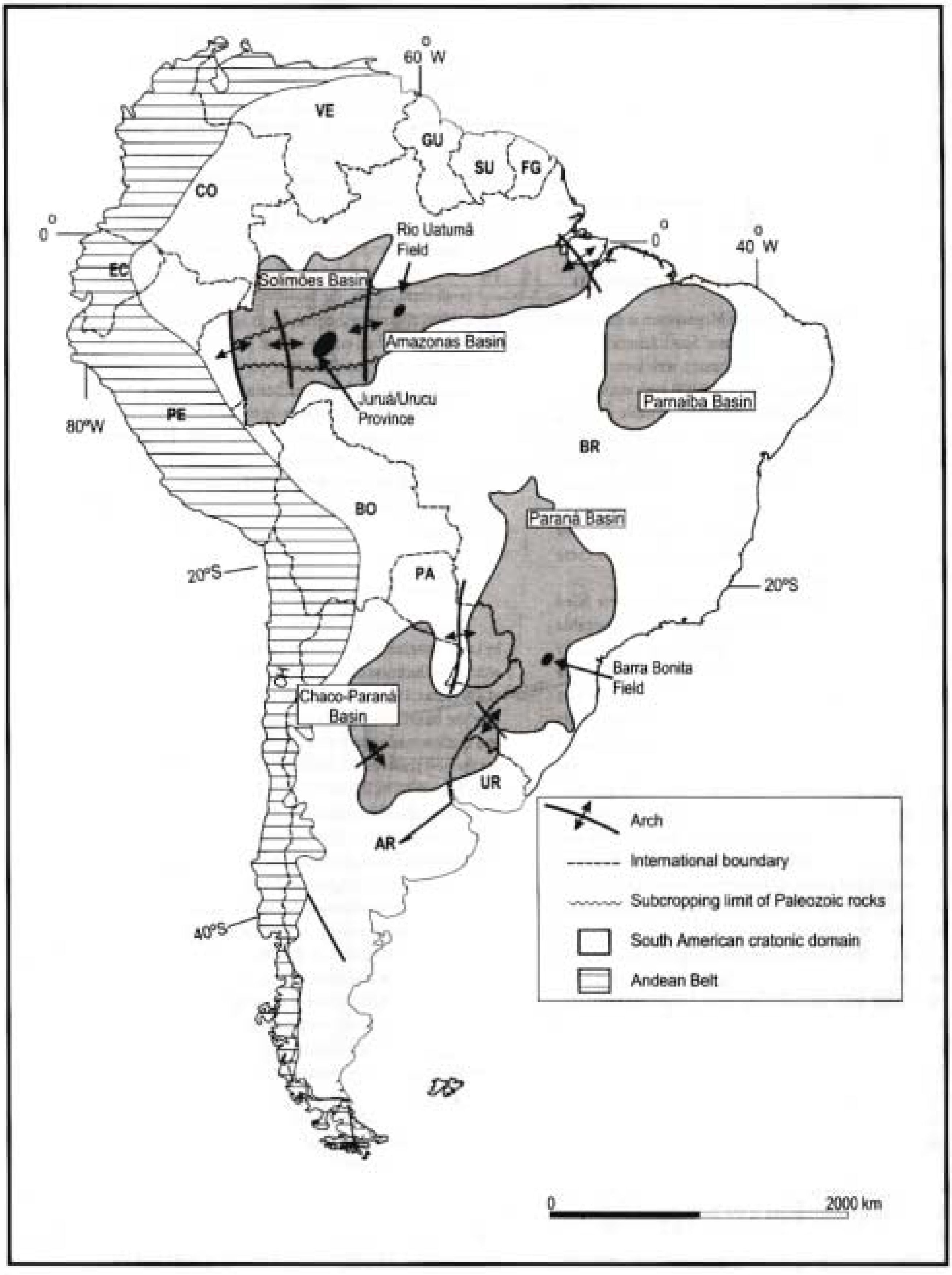
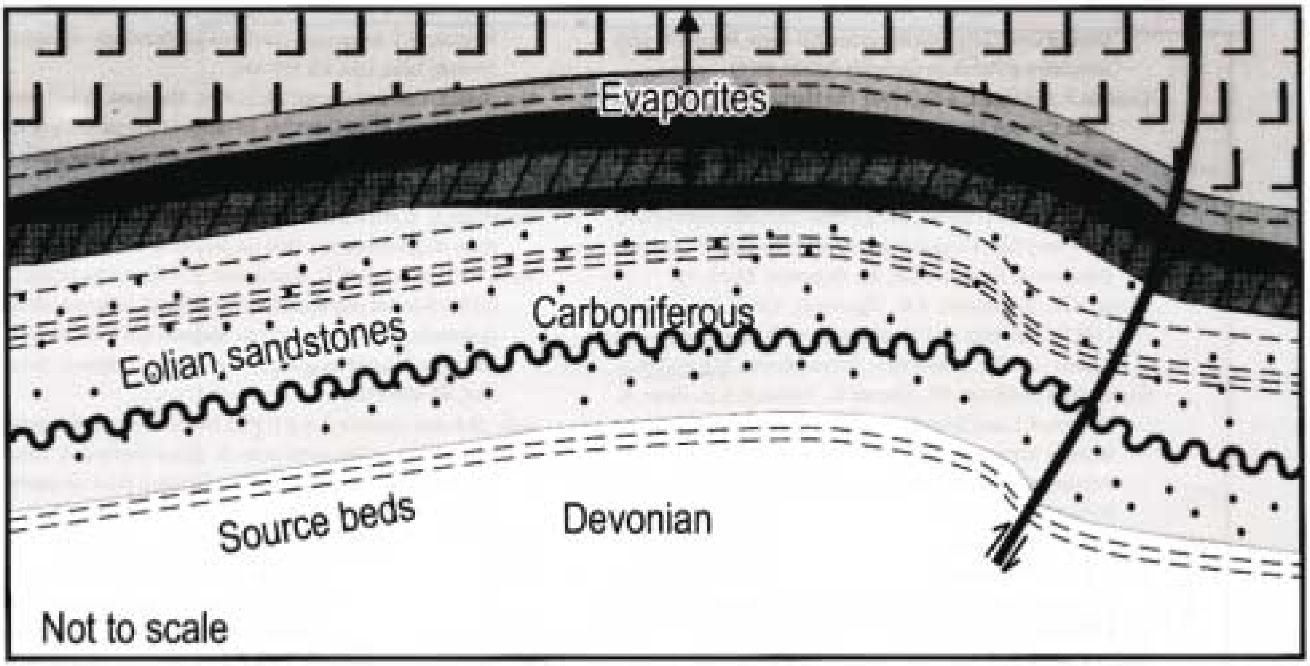


FIGURE 15 – Location map of Intracratonic Petroleum Megasystem showing the Paleozoic interior sag basins of the South American Continent, and associated oil and gas fields.

FIGURE 16 - Typical structural trap in the Solimões Basin, Brazil, where most of the reserves are located in fault-propagation folds associated with reverse faults of Mesozoic age (after Milani and Zalán, 1998).



(1 gas/condensate field and 2 gas field) provinces, holding in place reserves of 170 billion m^3 of gas and 121 million m^3 of oil and condensate. The present production is around 35.000 barrels of oil per day and 1.7 MMm^3/day of gas, making the Solimões Basin the most important petroleum-producing Paleozoic interior sag basin of South America.

Considering the huge size of these basins and the sparse coverage of seismic data, due to adverse local conditions, it is believed that the Intracratonic Petroleum Megasystem still is one of the most promising exploration targets for the next century.

Final Remarks

After one century of active petroleum exploration, the South American basins present different stages of maturity in terms of oil and gas exploration and production. It is believed that most basins may be still classified as being in the initial stage of exploration, principally when occurring in large areas covered by dense tropical jungle, or submerged in deep waters of the continental margins of the Atlantic and Pacific oceans. A few basins such as some onshore basins in Argentina, Brazil and Venezuela may be considered as in an advanced or in a more mature stage of exploration.

Nevertheless, even in these apparently well explored basins, a remaining exploration potential is present, taking into consideration the existence of untested deep objectives for gas production, the presence of strongly deformed area affected by tectonics, and several types of pure stratigraphic traps, still badly imaged by the 2D seismic data. The advent of 3D seismic processes associated with visualization techniques, as well the development of drilling and production techniques to reach deep targets and produce oil and gas from multi-lateral and extended-reach horizontal wells brought to these basin new chances to add huge untapped reserves.

On the other hand, improvement in the extraction techniques of already discovered oil and gas pools, using 4D seismic data, and a more severe control in the production parameters, pressure maintenance, reservoir characterization and reserve evaluation will bring the recovery factor of this accumulation to levels around 60% to 70%, adding a large amount of hydrocarbons to the world reserve.

According to Kronman *et al.* (1995) the probability of finding new giant oil and gas fields (above 500 million barrel of oil equivalent of original reserve) is very high in the Campos, Maracaibo, Llanos and Maturin basins. Subgiant accumulations and medium-size fields (between 50 and 250 million barrels of oil equivalent) may occur in Neuquén, San Jorge, Austral, Tarija, Oriente (including Marañon and Putumayo) and Magdalena basins.

A global estimation for the future reserves to be found in the South American basins is around 40 to 60 billion barrels of oil equivalent. Considering that the world tendency is to sustain or improve the global reserve, the South American Continent should keep the reserve/production ratio (R/P) around fifty years in respect to oil and seventy years in respect to gas for at least one more decade. This fact and the continuous transportation and market integration of most countries will keep the energy matrix essentially based on hydrocarbons for the first half of the next century.

REFERENCES

Aymard, R., Pimentel, L., Eitz, P., Lopez, P., Chaouch, J., Navarro, J., Mijares, J. and Pereira, J.G. (1990). Geological integration and evaluation of Northern Monagas, Eastern Venezuelan Basin. In: Classical Petroleum Provinces, ed. Brooks, J. *Geological Society Special Publication*, 70, 37-53.

Bruhn, C.H. (1998). Deep-water reservoirs from the Eastern Brazilian rift and passive margin basins. In: *Petroleum geology of rift and passive margin turbidite systems; Brazilian and worldwide examples*. AAPG/ABPG International Conference and Exhibition Short Course Notes 6, Rio de Janeiro, Brazil.

Bruhn, C.H. (1999). Reservoir architecture of deep lacustrine sandstones from the Early Cretaceous Reconcavo Rift-basin, Brazil. *American Association of Petroleum Geologists Bulletin*, **83**(9).

Candido, A. and Cora, C.A.G. (1990). The Marlin and Albacora giant fields, Campos Basin, Brazil. In: Giant oil and gas fields of the decade 1978-1988, ed. Halbouty, M.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, **54**, 123-135.

Carneiro, E.P. and Cavalcanti, A.R.S. (1994) The Cretaceous Oriente Basin revisited: a regional overview. In: *Anais do V Congresso Brasileiro de Petróleo e Gás*, Rio de Janeiro, Brazil, 9 p.

Chang, H.K., Kowsmann, R.O., Figueiredo A.M.F., Bender, A.A. (1992). Tectonics and stratigraphy of the East Brazil Rift System (EBRIS): an overview. *Tectonophysics*, **213**, 97-138.

Cooper, M. A., Addison, F.T., Alvarez, R., Hayward, A.B., Howe, S., Pulham, A.J. and Taborda, A. (1995). Basin development and tectonic history of the Llanos Basin, Colombia. In: (eds.) Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J.. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 659-665.

Daly, M.C. (1989). Correlations between Nazca/Farallon Plate kinematics and forearc basin evolution in Ecuador. *Tectonics*, **8**, 769-790.

Dengo, C. A. and Covey, M.C. (1993). Structure of the Eastern Cordillera of Colombia: implications for trap styles and regional tectonics. *American Association of Petroleum Geologists Bulletin*, **77**, 1315-1337.

Dow, W.G. (1974). Application of oil-correlation and source-rock data to exploration in Williston Basin. *American Association of Petroleum Geologists Bulletin*, **68**, 1752-1763.

Eiras, J.F. (1998). Geology and petroleum system of the Solimões Basin, Brazil. *Rio'98 AAPG International Conference and Exhibition, American Association of Petroleum Geologists/ Associação Brasileira de Geólogos de Petróleo, Extended abstracts volume*, p. 446.

Figueiredo, A.M.F., Braga, J.A.E., Zabalaga, J.C., Oliveira, J.J., Aguiar, G.A., Silva B., Mato, L.F., Magnavita, L.P., and Bruhn, C.H. (1994). Recôncavo Basin, Brazil: a prolific intracontinental rift basin. In: Interior rift basins, ed. Landon, S.M., *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **59**, 157-204.

Grieco, L. F. (1991), Distribución de fluidos en la estructura de Aguarague. *Boletín de Informaciones Petroleras*, **26**, 61-66.

Guardado, L.R., Spadini, A.R., Mello, M.R. and Brandão J.S.L. (1997). Petroleum systems of the Campos Basin, Brazil. In: *AAPG/ABPG Hedberg Research Symposium, Petroleum systems of South Atlantic Margin. Extended abstract.* Rio de Janeiro, Brazil.

Jaillard, E., Soler, P., Carlier, G. and Mourier, T. (1990). Geodynamic evolution of the northern and central Andes during the early and middle Mesozoic times. *Journal of Geological Society of London*, **147**, 1009-1022.

Kronman, G. E., Rushworth, S. W., Jagiello, K. and Aleman, A. (1995), Oil and gas discoveries and basin resources prediction in Latin America. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 53-61.

Legarreta, L., Laffite, G. A. and Minniti, S. A. (1999). Cuenca Neuquina: múltiples posibilidades em las Series Jurásico-Cretácicas del depocentro Periandino. In: *Actas del IV Congreso de Exploracion y Desarollo de Hidrocarburos*, Mar del Plata, Argentina, **1**, 145-175.

Lugo, J. and Mann, P. (1995). Jurassic-Eocene tectonic evolution of Maracaibo Basin, Venezuela. In: Petroleum basins of South America, eds Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir* , Tulsa, USA, **62**, 699-726.

Mathalone, J.M.P. and Montoya M.R. (1995). Petroleum Geology of the Sub-Andean Basins of Peru. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 423-444.

McCollough, C.N. and Carver, J.A. (1990). The giant Caño Limon Field, Llanos Basin, Colombia. In: Giant oil and gas fields of the decade 1978-1988, ed. Halbouty, M.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **54**, 175-196.

McGettigan, C. K. and Hunt, D. G. (1996). Colombia continues to yield major oil, gas discoveries. *Oil & Gas Journal Special*, July, **15**, 40-45.

Mello, M.R., Mohriak, W.U., Koutsoukos, E.A.M. and Bacoccoli, G. (1994). Selected petroleum systems in Brazil. InThe petroleum systems from source to trap, eds. Magoon, L.B. and Dow, W.D. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **60**, 499-512.

Mello, M.R. and Trindade, L.A.F. (1996). Nuevas metas exploratorias em cuencas latinoamericanas de aguas profundas: cómo proceder con el concepto de sistema petrolero. *Oil & Gas Jounal, Revista Latinamericana*, **2**(1), 25-32.

Milani, E.J. and Zalán, P.V. (1998) *Brazilian Geology Part 1: the Geology of Paleozoic Cratonic Basins and Mesozoic Interior Rifts of Brazil.* Rio'98 AAPG International Conference and Exhibition, American Association of Petroleum Geologists/ Associação Brasileira de Geólogos de Petróleo, Short Course Notes, Rio de Janeiro, Brasil.

Milani, E.J. and Zalán, P.V. (1999). An outline of the geology and petroleum systems of the Paleozoic interior basins of South America. *Episodes*, **22**(3), 199-205.

Mpodozis, C. and Ramos, V. (1989) The Andes of Chile and Argentina. In: Geology of the Andes and its relation to hydrocarbon and mineral resources: Circum-Pacific Council for Energy and Mineral Resources, eds. Ericksen, G.E., Cañas Pinochet, M.T. and Reinemund, A. *Earth Science Series*, **11**, 59-90.

Pittion, J.L. and Arbe, H. (1997). Petroleum systems in the Austral Basin. In: *AAPG/ABPG Hedberg Research Symposium, Petroleum systems of South Atlantic Margin, Extended abstract.* Rio de Janeiro, Brazil.

Prieto, R. and Valdes, G. (1990). El Furrial oil field a new giant in an old basin. In: Giant oil and gas fields of the decade 1978-1988, ed. Halbouty, M.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **54**, 155-163.

Ramírez, E. and Marcano, F. (1990) Ceuta-Tomoporo Field, Venezuela. In: Giant oil and gas fields of the decade 1978-1988, ed. Halbouty, M.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **54**, 163-174.

Robbiano, J.A., Arbe, H.A. and Gangui, A. (1996). Cuenca Austral Marina. In: Ramos, V.A. and Turic M.A. (eds.). *XIII Congreso geológico Argentino y III Congreso de exploracion de hidrocarburos. Geología y recursos naturales de la Plataforma Continental Argentina*, pp. 323-341. Buenos Aires, Argentina.

Schamel, S. (1991). Middle and Upper Magdalena basins, Colombia. In: Active margins basins, ed. Biddle, K.T. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **82**, 283-301.

Szatmari, P., Milani, E.J., Lana, M.C., Conceição, J.C.L. and Lobo, R.P. (1985). How South Atlantic rifting affects Brazilian oil reserves distribution: *Oil and Gas Journal*, **83**, 107-113.

White, H.J., Skopec, R.A., Ramirez, F.A., Rodas J.A. and Bonilla G. (1995). Reservoir characterization of the Hollin and Napo formations, western Orient Basin, Ecuador. In: *Petroleum basins of South America*, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 573-596.

Vergani, G.D., Tankard, A.J., Belotti, H.J. and Welsink, H.J (1995). Tectonic evolution and paleogeography of the Neuquen Basin, Argentina. In: Petroleum basins of South America, eds. Tankard, A.J., Suárez Soruco, R. and Welsink, H.J. *American Association of Petroleum Geologists, Memoir*, Tulsa, USA, **62**, 383-402.

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An Overview Of The Coal Deposits Of South America

Ricardo da Cunha Lopes and José Alcides Ferreira

The coal deposits of South America occur mainly in two geological settings: in sedimentary sequences deposited in some of the pre-Andean and Andean basins along the western border of the continent, and in the Late Carboniferous to Permian deposits of the Paraná Basin. Coal is mined for usage in power generation (thermal coal) and/or in steel making (coking coal) in six of the 13 South American countries: Argentina, Brazil, Chile, Peru, Venezuela and Colombia. Two of these countries, principally Colombia, and to a lesser extent Venezuela, also export coal. Four of the 13 countries, Ecuador, Bolivia, Paraguay and Uruguay, have coal resources that have not yet been evaluated, and the information available on the coal resources of Guiana, Suriname and French Guiana is very limited.

Coal basins

The major coal deposits of South America are situated on the western side of the continent (Fig. 1). They are associated with the basins that evolved along the convergent margin of the plate, ranging in age from the Mesozoic to the Tertiary. Inside the continent coal deposits are found mainly in the Paraná Basin, a large intracratonic basin formed by lithospher sagging, and are Permian in age. In both settings coal is mainly of the bituminous class, with smaller deposits of lignite, anthracite and peat.

In the basins of the Andean Cordillera and in Patagonia, coal is mainly of the bituminous to sub-bituminous types. Anthracite deposits are restricted to basins in the central region of the Andes. The coal deposits of the Paraná Basin are considered to be high volatile bituminous coal, almost without exception. Lignite is found in the Amazon region near the frontier between Brazil and Colombia. In the Caribbean region, coal is bituminous or rarely sub-bituminous.

The coal deposits of the Paraná Basin are Palaeozoic (Late Carboniferous to Permian) in age, and they are attributed mainly to accumulations of organic matter in ancient coastal regions in which there developed deltaic and barrier-lagoon systems; low-energy environments favourable to the growth of luxuriant vegetation. By comparison, the coal deposits of the Andean Cordillera are Mesozoic to Tertiary in age and are interpreted to have formed in a limnic environment.

Coal rank is very much a function of depth of burial and time. Thus those coals that developed under the low geothermal gradient of the Paraná Basin have dominantly a low rank, whereas those of the sub-Andean basins have a somewhat higher rank. However, in some of the basins along the Andean Cordillera where the geothermal gradient was still higher, the highest rank coal (anthracite) in South America can be found.

A total of 22 coal basins is recognized in South America (Olade, 1984) (Fig. 1). Twenty of these are found along the Andean Cordillera: Oriental and Maracaibo in Venezuela; Cesar, Magdalena, Llanos, Sinu-Atlantica and Cauca in Colombia; Cañar-Azuay/Lojas-Malacatos in Ecuador; Tumbres, Gayllarisquizga and Arequipa in Peru; Occidental and Oriental in Bolivia; Arauco, Valdivia, Osorno Llanquihue and Magallanes in Chile; Precordillera La Rioja-San Juan and Precordillera Mendoza-Neuquén in Argentina and Magallanes (shared with Chile). One basin is situated on the Patagonian Platform, the Patagonia Central Basin in Argentina, and the other is the South Brazilian Paleozoic Paraná Basin. Exposures of coal in the Paraná Basin occur mainly in Brazil, with minor areas in Uruguay and Paraguay. The coal basins may contain several deposits to which regional or geographical names have been assigned.

Coal deposits by country

Colombia

Colombia is South America's main coal producer and exporter with potential for expansion. Coal is mined in two distinct regions: the Andean Highlands, and the Caribbean Shore Zone. Recent production figures are about 30 Mt/year. Exports amount to 80% of the production, which places Colombia in ninth position amongst the world's coal exporters. The Colombian coal is Cretaceous to Tertiary in age.

The largest deposit, El Cerrejón, occurs in the La Guajira Department of the Caribbean Shore Zone. It is divided into northern, central and southern sectors of which the northern sector is the most important economically on account of the before mining reserves amounting to 3 Gt to 300 m depth. Around 40 economically recoverable seams have been identified in the district, that reaches a total area of 120 km^2.

Cerrejón's deposits are of Tertiary age, accumulated between 55 and 25 Ma ago, and contain several coal seams intercalated between shale and siltstone. Structural contortion of the strata by thrust faults affected the coal mesures locally. Regional metamorphism affected the coal deposits and resulted in increases in rank, with original low-rank resources being upgraded to high-volatile bituminous types A, B, and C. The El Cerrejón coal has general poor coking

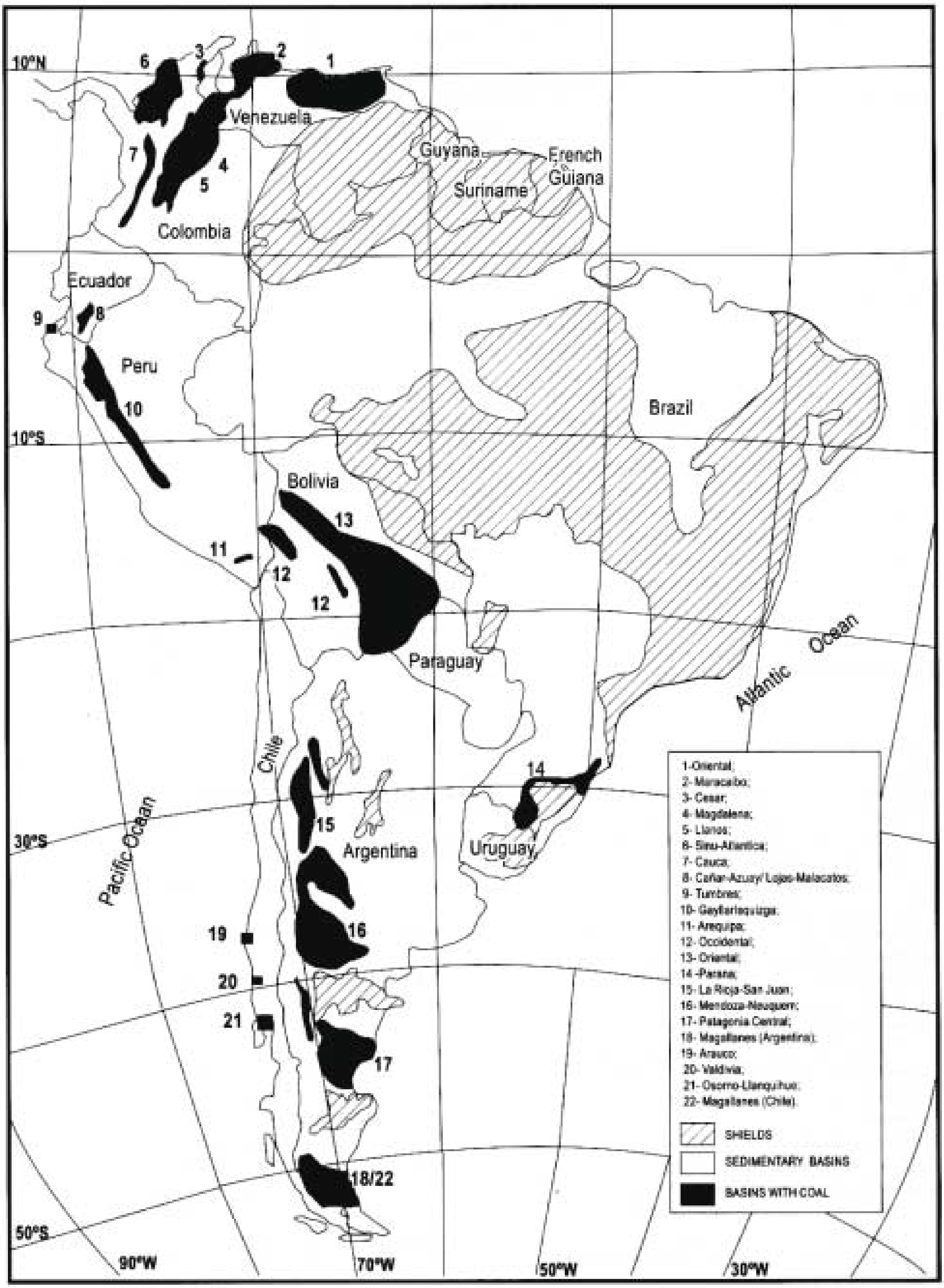

Figure1: - Main basins with coal in South America. (Adapted from OLADE,1984)

properties. However, it has low ash (7.5%) and sulphur (0.7%) content that make it suitable for the generation of electricity in thermoelectric plants as well as for usage in industry. This mine is currently being worked under a joint venture agreement between Carbocol, a state-owned Colombian enterprise, and a subsidiary of Exxon.

In the Andean Highlands, the mines are small probably due to the highly folded and faulted condition of the deposits. The deposits have been mined extensively, and include those situated in the Cundinamarca, Santander and Antioquia departments. Resource estimates made in 1987 range from 16 to 40 Gt, with reliable surveys covering only some 20% of the deposits. Production is destined mainly for the domestic market, although many seams contain coking coal with high bituminous to medium volatile rank with low ash and sulphur content.

Brazil

Brazil has a long-established coal mining industry in the southern part of the country that dates from the last century. There are eight large deposits associated with Permian sedimentary successions of the Paraná Basin, seven of which occur in the State of Rio Grande do Sul, and one in the State of Santa Catarina (Aboarrage and Lopes, 1986). In addition to these eight large deposits, there occur several small workings, mostly in the State of Paraná.

From SW to NE the main deposits are Candiota, Capané, Iruí, Leão, Charqueadas, Morungava-Chico Lomã, Santa Terezinha, in the State of Rio Grande do Sul; and Sul-Catarinense in the State of Santa Catarina. Total resources are estimated at 30 Gt, with economically recoverable reserves estimated at 20 to 30% of the total resource (Gomes *et al.*, 1998). Production figures in recent years show 10 Mt/year ROM. However, environmental constraints have lowered this figure to about 4.5 Mt/year of saleable coal.

Gondwana coals of Southern Brazil are associated with the transitional deposits of the Rio Bonito Formation, of Artinskian-Kungurian (Early Permian) age (Corrêa da Silva, 1989). The sequence reaches a maximum thickness of 300 m and is composed dominantly of shale and sandstone containing remains of the *Glossopteris* flora. Different paleodepositional environments are associated with the origin of these coals, with the dominance of subaqueous facies holding the coal mesures. More recently, the occurrence of cyclically-stacked peat swamps in the Permian of the Paraná Basin has been attributed to high frequency sea level variations and development of a pattern of transgressive parasequences (Alves and Ade, 1996).

The Brazilian coal rank determined by reflectance measurements varies from the sub-bituminous C-type to the high volatile, bituminous A-type (Corrêa da Silva, 1989). Anthracite, attributed to the thermal effects of Mesozoic intrusions, occurs locally. The coal from the State of Santa Catarina is of coking grade. However, it contains unacceptable ash and sulphur content, and therefore its usage in the metallurgical industry has been abandoned. Better coking grade coal is found in some deposits in the State of Rio Grande do Sul but this has not been exploited for this purpose yet. The main reason for this is the availability of metallurgical-grade charcoal. Nowadays, Brazilian coal is used in thermoelectric plants as well as in the cement, ceramics and petrochemical industries.

Venezuela

Coal deposits in Venezuela occur in the western part of the country, in the Guasare region of the State of Zulia. They may be compared, geologically, with those of El Cerrejon in neighbouring Colombia. Current coal production comes mainly from the large Paso Diabo Mine, owned by the AGIP/Shell joint venture. The Guasare coal field contain an estimated 2 Gt of coal of which some 200 Mt is high quality bituminous coal with low sulphur content. Low rank coals are found in the northern region and along the valley of the Orinoco River, but these are not being exploited.

Between 1987 and 1997, coal production rose from hundreds of thousands of tonnes to 5.5 Mt. Exports amounted to about 90% of the production, principally to Europe. Venezuela imports a small amount of metallurgical-grade coal. A production increase to tens of millions of tonnes over the next few years is planned.

Argentina

There are many occurrences of coal in Argentina. These include those found in the Carboniferous to Tertiary basins of the Precordillera region, including La Rioja-San Juan and Mendoza-Neuquén. Anthracite occurs at Cervantes. In the Patagonia Central Basin (Tertiary) several coal showings have been reported. Deposits at Rio Turbio, an extension of the Chilean Magallanes Basin, have been mined for many years, but the production levels are small. In general, the economic potential for coal is low.

Chile

Chile produces small quantities of coal from the Arauco-Valdivia Basin of Tertiary age, as well as from other basins in the southern part of the country.

Ecuador

Coal occurrences are found in the Cañar-Azuay Basin of Tertiary age in southern Ecuador. Coal seams have also been reported from Mesozoic sediments of the Napo Basin the western part of the country. None of these occurrences, classified as sub-bituminous C, are considered to have economic potential.

Peru

The main occurrences of high-rank coal in Peru, among the highest in South America, are found in the Gayllarisquizga Basin of Cretaceous age, in which several deposits of anthracite are known. In the Arequipa Basin in the southern part of the country, the coal rank varies from bituminous to anthracite. However, Peruvian coal production is small because the deposits are found in regions of difficult access. Furthermore, the seams are strongly folded and fractured, and the coal has high ash and sulphur content.

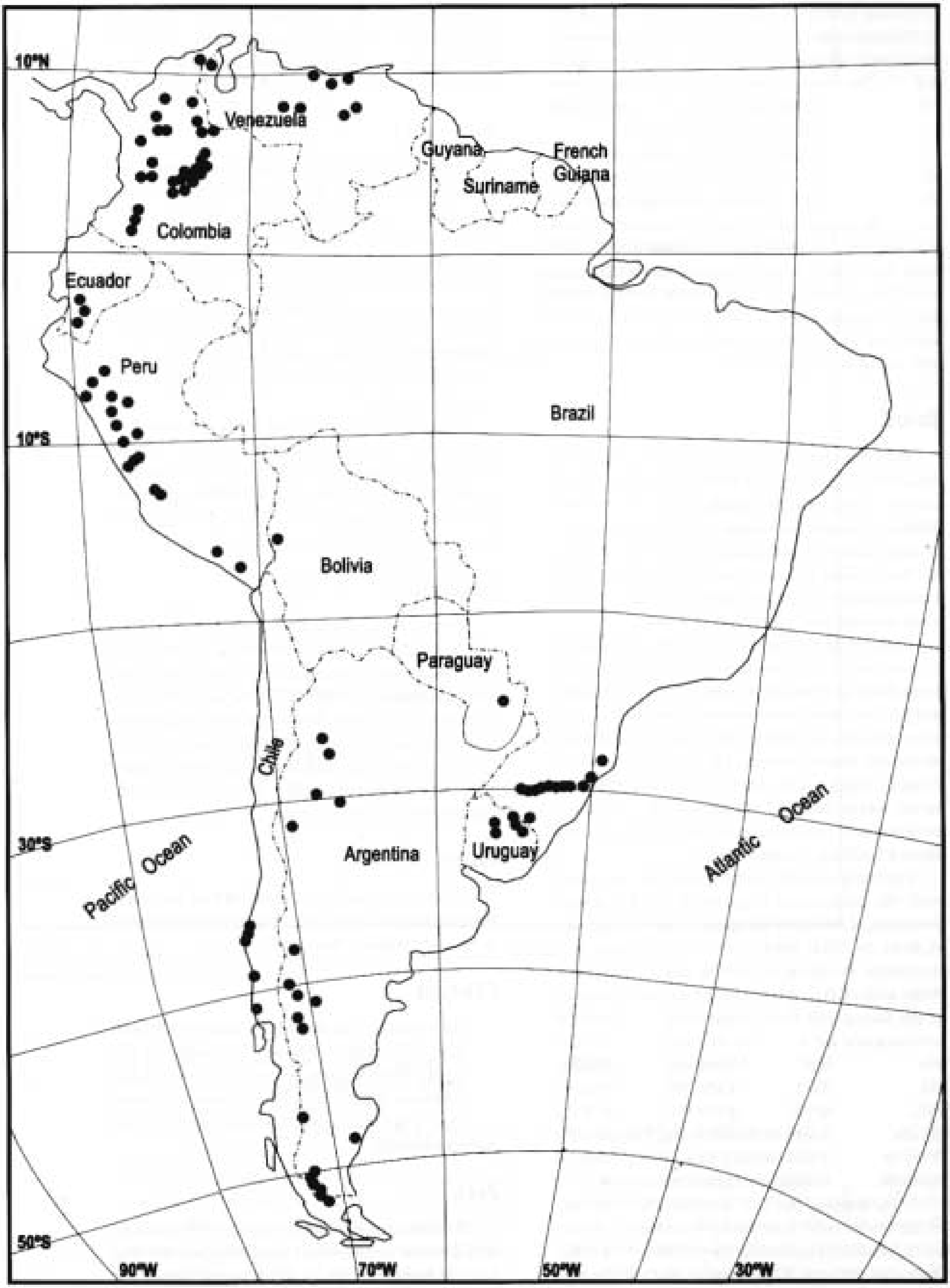

Figure2: - Coal occurrences and deposits in South America.(adapted from OLADE,1984)

Bolivia

In Bolivia, coal occurs in the Carboniferous sedimentary sequences of the Occidental Cordillera where only one deposit, Copacabana-Isla del Sol, is being worked. Although the Occidental Cordillera consists of a large expanse of sedimentary rocks of Carboniferous to Tertiary age, only minor coal occurrences have been reported.

Uruguay

Only a few coal showings are known in Uruguay. These are found in Permian sediments of the Paraná Basin where this extends southwards from the Candiota Deposit in the neighbouring State of Rio Grande do Sul, Brazil.

Paraguay

In like manner to Uruguay, the coal showings in Paraguay are associated with the Permian sediments of the Paraná Basin and are without economic potential.

REFERENCES AND SOURCES OF DATA

Aboarrage, A.M. and Lopes, R.C. (1986) *A Borda Leste da Bacia do Paraná: integração geológica e avaliação econômica*. DNPM/CPRM, Porto Alegre, Brazil (unpublished report).

Alves, R.G. and Ade, M.V.B. (1996) Sequence stratigraphy and coal petrography applied to the Candiota Coal Field, Rio Grande do Sul, Brazil: A depositional model. *International Journal of Coal Geology* **30**: 231-248.

Corrêa da Silva, Z.C. (1989) The rank evaluation of South Brazilian Gondwana coals on the basis of different chemical and physical parameters. *International Journal of Coal Geology* **13**: 21-39.

Gomes, A.P., Ferreira, J.A.F., Albuquerque, L.F.F., and Süffert, T. (1998) Carvão fóssil. Universidade de São Paulo, *Estudos Avançados* **12**: 89-106.

Olade (1984) Inventario de Recursos de Carbón Mineral en Paises no Productores de America Latina. Quito, Ecuador.

United States Energy Information Administration. [On line]: Argentina: energy, oil, natural gas, electricity, environment, profile, 1998. [11/1998]. Available in http:// www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Bolivia.1998. [12/1998]. Available in http://www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Brazil: energy, oil, natural gas, coal, electricity, environment, privatisation profile, profile. 1998. [10/1998]. Available in http://www.eia.doe.gov/international/.

United States Energy Information Administration. [On line]. Chile: oil, natural gas, electricity, 1998. [10/1998]. Available in http:/ www.eia.doe.gov/international/.

United States Energy Information Administration. [On line]. Colombia: oil, natural gas, coal, electric power, environment, profile, 1998. [07/1998]. Available in http://www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Ecuador. 1998. [12/1998]. Available in http:/www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Paraguay: energy highlights, profile. 1997. [09/1997]. Available in http://www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Paraguay: energy highlights, profile. 1997. [09/1997]. Available in http:// www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Peru. 1999 [02/1999]. Available in http://www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Peru: Mining and Energy 1992. [07/1992]. Available in http:// www.eia.doe.gov/international/.

United States Energy Information Administration. [On line]. Uruguay. 1999 [04/1999]. Available in http://www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. Venezuela: oil natural gas, coal electricity. 1998. [09/1998]. Available in http://www.eia.doe.gov/ international/.

United States Energy Information Administration. [On line]. An Energy Overview of Venezuela. 1999. [04/26/1999]. Available in http://www.eia.doe.gov/international/.

United States Energy Information Administration. [On line]. An Energy Overview of Brazil. 1999. [05/10/1997]. Available in http://www.eia.doe.gov/international/.

ANDEAN METALLOGENESIS

A SYNOPTICAL REVIEW AND INTERPRETATION

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In geological terms, the Andean Belt has special importance as a model for the evolution of magmatic arcs developed close to the continental crust, on an active, plate consuming convergent border. Although the magnetic anomalies of the ocean floor permit us to follow the convergence history of the margin only as far back as the Cretaceous, there is geological evidence for plate tectonic activity in the Andean Domain during Paleozoic times. In consequence, the geological evolution of the Andes offers a most interesting frame for describing the metallogenetic development of the belt, and answers that explain the origin and geological evolution of its mineral deposits.

The Andean Belt is a complex orogenic system that has its maximum width (approximately 800 km) at around 18 °S. Within this domain there are several cordilleras, sierras, plateaux, basins and valleys. Three well-defined cordilleras and one sierra occur in Colombia, while only one cordillera exists in southern Ecuador. The present configuration of the belt is relatively recent. Thus, the Bolivian Altiplano was a zone of subsidence until its Pliocene uplift. The valleys, or longitudinal depressions, present areas of rapid subsidence in some sectors (*e.g.* Colombia, southern Chile), where the accumulated Plio-Quaternary sedimentary and volcanic deposits are 5000 to 10 000 m thick.

The present Andean cordilleras rise over the western and northwestern border of the South American Tectonic Plate and face four other tectonic plates; three of which are of the oceanic type: the Nazca, Cocos and Caribbean plates; and one is of the oceanic-continental type, the Antarctica Plate. Only the Cocos, Nazca, and Antarctics plates show active subduction, the relative motion of the Caribbean Plate being of the transcurrent type. Seismic activity affects the entire Andean Belt, but the Benioff Zone under the continent exhibits important differences in definition and angle of dip, attaining a maximum depth of about 350 km in the central part of the Andean Chain. The oceanic trench also attains a maximum depth of about 8 km between latitudes 22°S and 25°S, where it runs parallel to the coast, some 100 km to the W. Thus the topographic contrast between the summits of the Andean Chain and the bottom of the neighbouring trench is the greatest on Earth.

The thickness of the continental crust varies along the belt. It attains a maximum of 70 km under the Principal Cordillera, between 14 °S and 22 °S, a figure that corresponds to that of the continental crust under the Himalayas. On the other hand, the thickness of the continental crust is small or nihil under a part of the coastal regions of Colombia and Ecuador. The thickness of the continental crust also varies across the belt; first increasing, then diminishing, until it stabilizes at 30 - 35 km under the continental shield.

The presence of major longitudinal and transverse faults is an important feature of the Andean geology. The major longitudinal faults have controlled the vertical displacement of the longitudinal tectonic blocks, as well as the emplacement of magmas and the distribution of ore deposits. Several of these faults, such the Romeral (Colombia) and Atacama (Chile) faults, have a history that began, at least, during the Early Mesozoic. The transversal faults linked to differential displacement of the continental plate have also played an important role in the distribution of some ore deposits. A good example is the Chaucha porphyry copper, Ecuador (Goossens and Hollister, 1973).

The Andean Belt contains hundreds of stratovolcanoes, many of them reaching high altitudes. The stratovolcanoes occur in three main active segments: 5°N - 2°S (andesitic-basaltic), 16°S - 28°S (andesitic) and 37°S - 46°S (andesitic-basaltic). Only five stratovolcanoes are known in the more southerly parts of the Andean Chain (48°S - 56°S), and their composition is andesitic. The principal volcanic segment, 16°S - 28°S, contains around 150 000 km^2 of Miocene-Pliocene rhyo-dacitic ignimbrite flows some of which are associated with very large calderas, up to 30 km in diameter (Francis and Baker, 1978). Some of the Andean volcanoes have provided important clues for understanding the genesis of the ore deposits. Two good cases of this are San Fernando, Ecuador (Goossens, 1972a), and El Laco, Chile (Park, 1961).

Although some authors, such Auboin *et al.* (1973) and Zeil (1979), maintain that there exist fundamental differences between the Paleozoic and post-Paleozoic geological development of the Andean Belt, these differences depend on the specific Andean segment under consideration and the time. Neither the episodes of marginal basin development nor the stages of strong horizontal compressive tectonics are exclusive aspects of the Paleozoic evolution. On the other hand, important Paleozoic sedimentary basins are characterized by vertical tectonics. Also, calc-alkaline magmatism, so typical of the Mesozoic-Cenozoic Andean belts, was equally widespread during the Paleozoic, attaining a peak during the Permian. Thus, the Permian-Triassic transition occurs in geological continuity. Finally, the Paleozoic and post-Paleozoic tectonic strikes are similar, and the Paleozoic metallogenesis includes the same metals as those formed in Mesozoic and Cenozoic times, notwithstanding that the areal distribution of the metallic belts is different. Porphyry copper deposits, a main feature of the Cenozoic Andean metallogenesis, were formed in the Andean Domain at least since the Carboniferous (Sillitoe, 1977).

Nevertheless, some of the characteristic aspects of the Andean Belt, *e.g.*, the generation of large amounts of calc-alkaline rocks, increased in Mesozoic and Cenozoic times, whereas other aspects, such as the accretion of oceanic prisms, diminished in relative importance. The separation of South America from Africa that began during the Jurassic, did not imply radical changes in the evolution of the Andean Belt, which remained basically a magmatic belt (Zeil, 1979). As suggested by Coney (1970) and other authors, this evolution may be described in terms of the superimposition of magmatic arcs over the edge of the continent. This description, which is valid for the central Andean segment, should also include periods of accretion to the continent of oceanic magmatic-sedimentary prisms at the northern and southern ends of the belt.

Ensialic basin development was also important during Mesozoic and Cenozoic Andean evolution. However, some of these Mesozoic basins (*e.g.*, the Neocomian basin in central Chile; Aberg *et al.*, 1983), attained during their evolution several characteristics of the marginal basins.

The Andean Belt shows the imprint of several important compressive episodes. However, their intensity was different along the belt. Besides which, intensive folding only affected the miogeosynclinal rocks occurring between the volcanic terranes to the W and the continental terranes to the E.

Mesozoic Andean magmatism includes tholeiitic, calc-alkaline and alkaline series. The tholeiitic series are characteristic of the accreted oceanic prisms of the Northern Andes, whereas calc-alkaline magmatism predominated along the principal magmatic arc of the belt. Alkaline magmatism occurs in small amounts as intrusive and extrusive bodies in the back-arc region. The presence of shoshonitic rocks has been reported both in the Jurassic-Early Cretaceous magmatic arc in Central Chile (Levi *et al.*, 1988) as well as in the Tertiary back-arc region of north-western Argentina (Sasso and Clark, 1988).

With regard to the older ore deposits in the Andean Domain, the only ones that have a possible Precambrian age are some Ni and Cr ores in ultrabasic rocks of the Eastern Cordillera of Peru, as well as some Ni-Cr deposits in ultrabasic rocks; Cu-Fe deposits in amphibolite and W deposits in granulite of the Pampean Ranges of Argentina, of minor economic importance (Di Marco and Mutti, 1996; Stoll, 1975).

Though Paleozoic and post-Paleozoic Andean ore deposits contain basically the same metals, there are some differences regarding the type of deposit (*e.g.*, post-Paleozoic BIFs are not known). However, the main difference concerns the huge amounts of ores formed after the Paleozoic, especially in the Central and Southern Andes, and which are generally associated to sub-volcanic igneous rocks.

The post-Paleozoic metallic provinces appear as 50 to 300 km wide belts elongated and parallel to the Andes. Between latitudes 14 °S and 30 °S, where three or all four provinces are present, the iron belt appears close to the Pacific coast, followed by the copper, polymetallic and tin belts, the easternmost of which lies some 500 km from the coast. Although these major provinces are defined by the predominance of one or two principal metals, they contain ore deposits of different ages, typologies and paragenesis. Both, the copper and the polymetallic provinces are present along most of the Andean Chain, although some segments are extremely rich, whereas in others, metal deposits are scarce or do not occur at all. However, the tin and the iron provinces are more restrained along the Andean Belt but show a homogeneous distribution of the ore deposits.

The Andes Chain is one of the richest orogenic belts in terms of metallic ores in the world. Several Andean countries are amongst the world's top ten producers or in terms of their geological reserves of antimony, barium, beryllium, bismuth, boron, copper, indium, iodine, lead, molybdenum, nitrate, platinum, rhenium, selenium, silver, tellurium, tin, tungsten and zinc (Petersen, 1977). Chile, alone, has about a quarter of the world's copper reserves and close to one third of those of molybdenum. The tin-silver province of southern Bolivia is well known for its fabulous deposits, such as Cerro Rico, Potosí, which produced some 60 000 t of silver, in addition to high-grade tin ores. Equally famous are the polymetallic and copper provinces of Peru. Besides which, the last 30 years have been generous in terms of the discovery of new world class ore deposits, such as El Indio and Escondida in Chile. Indeed, the Andes Belt is still considered a first class target, attracting about 15% of the world's investment in mineral exploration. On the other hand, important scientific studies have been dedicated to those types of ore deposits that are well represented in the Andean Belt. This is the case of porphyry copper, epithermal Au-Ag deposits, the Sn-Ag sub-volcanic deposits in Bolivia and the zoned polymetallic deposits of Peru. Also, some industrial minerals of the Andes are of special interest, as in the case of the evaporite deposits of Chile, Bolivia and Argentina that contain huge amounts of potassium, lithium, iodine, nitrate and borate.

This review will now describe the different metallic provinces of the Andes, including a special section on precious metal deposits. The review will terminate with a discussion of the main factors involved in the metallogenetic interpretation of the belt.

Metallic Provinces in the Andes

The iron belt

The iron ore deposits of the Andean Domain (Fig. 1) may be grouped in four types: BIF-type deposits of the Nahuelbuta Belt (Chile); oolitic iron formation deposits in northwestern Argentina and Colombia; Kiruna-type deposits in the coastal ranges of northern Chile and Peru, and skarn-type Fe-Cu deposits of the Andahuaylas-Yauri zone of Peru. The magnetite deposit of the El Laco volcanic structure in northern Chile is included in the third group, but will be considered separately, because of the special characteristics of this district.

The BIF-type iron ores of Nahuelbuta occur in high-pressure metamorphic rocks (pelitic schist, chert and greenschist) that have a Lower Carboniferous metamorphic age, and belong to an accreted terrain (Aguirre *et al.*, 1972). The oceanic volcano-sedimentary prisms contain, and to the magnetite ores, some podiform chromite deposits and some pyritic Cu-Zn massive sulphide bodies. The principal iron mineralization, that is interbeded with micaschist, crops out in three main areas, situated between 38°05'S and 38°30'S, close to 73°15'W. Ore reserves are about 100 Mt, containing 30% Fe (Oyarzún *et al.*, 1984).

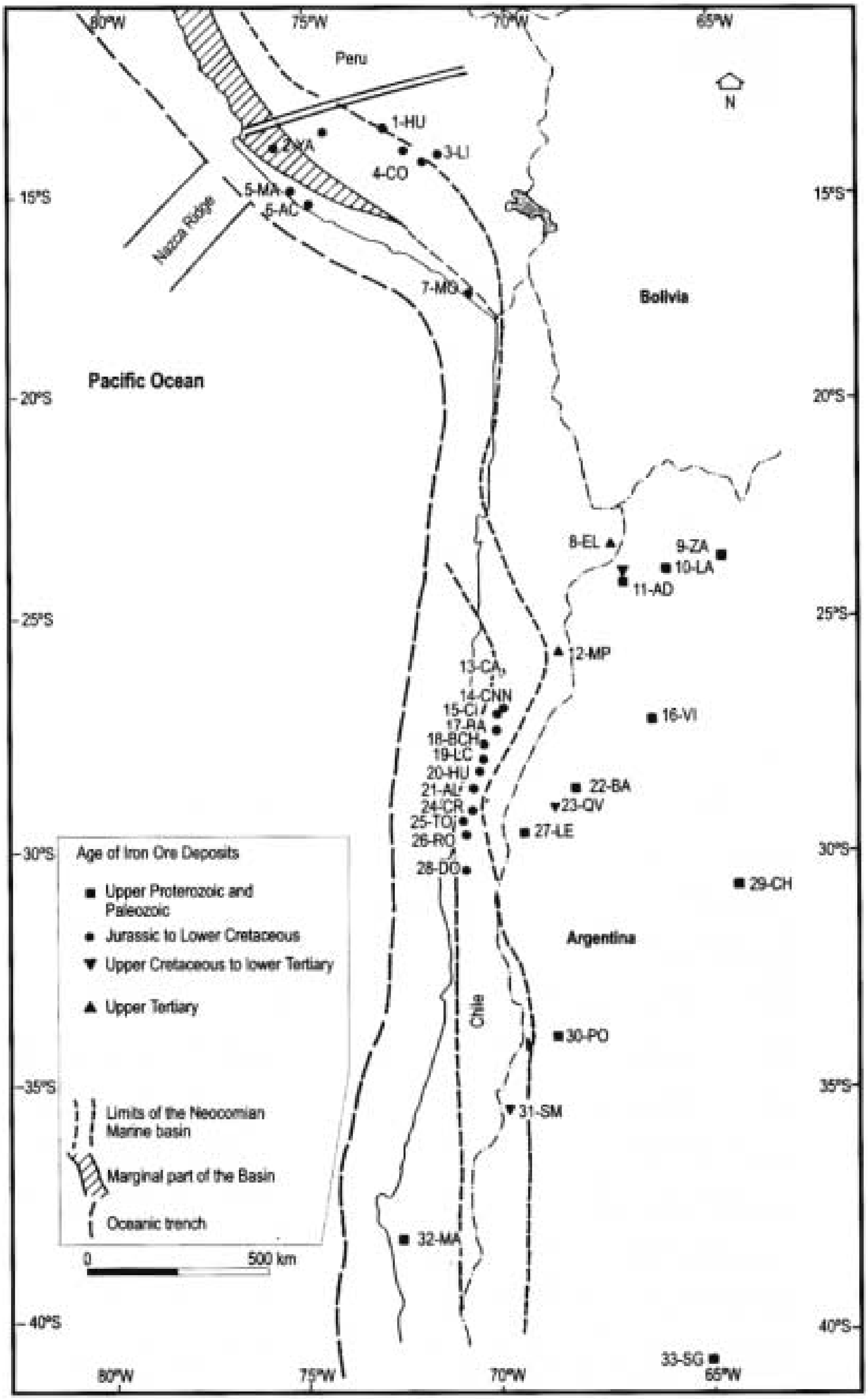


FIGURE 1 - Iron ore deposits in southern Peru, Chile and Argentina. List of deposits: 1- Huancabamba. 2- Yaurilla. 3- Livitaca. 4- Colquemarca. 5- Marcona. 6- Acari. 7- Morrito. 8- El Laco. 9- Zapla. 10- Lagunillas. 11- Agua del Desierto. 12- Magnetita Pedernales. 13- Carmen. 14- Cerro Negro Norte. 15- Cerro Imán. 16- Vizvil. 17- Bandurrias. 18- Boquerón Chañar. 19- Los Colorados. 20- Huantemé. 21- Algarrobo. 22- Borde Atravesado. 23- Quebrada Varela. 24- Cristales. 25- El Tofo. 26- Romeral. 27- Leoncito. 28- El Dorado. 29- Characato. 30- Potrerillos. 31- Sur de Mendoza. 32- Mahuilque. 33- Sierra Grande (modified after Angelelli et al., 1970; Bellido et al., 1972; Grez et al., 1991; Ruiz et al., 1965).

The oolitic iron formation deposits are found in northwestern Argentina, where they have a Lower Silurian age, and in Colombia, where they are upper Eocene in age. The Argentinean deposits are in coastal marine facies at the eastern border of a central craton. The ores are oolitic, and the iron beds, deposited during a marine transgression, contain chamosite (partly altered to hematite) as the principal iron mineral (Bossi and Viramonte, 1975). Although the productive formation (Zapla) crops out along a N-S strike for hundreds of kilometres, the principal deposits occur between 24°S and 25°S, close to 65°W. These deposits are: Zapla, Río Iruya and Unchimé. Their total pre-mining reserves are about 300 Mt ore, containing 40% Fe (Angelelli *et al.*, 1970).

The oolitic iron formation deposits of Colombia are part of a belt some 650 km long and trending NNE, between Lagunillas (Venezuela) and Sabanalarga (Colombia), that contains at least four zones of mineralized outcrops. The principal zone is that of Paz del Río (6°11'N / 72°43'W), where the oolitic iron formation is 0.5 to 8 m thick and crops out along a strike length of 57 km trending N30°W. The maximum width is 8 km. The ores contains 42 - 47% Fe and 0.8 - 1.2% P. The Paz del Río deposit, in like manner to the other deposits in the belt, was formed during the upper Eocene, as the result of a marine transgression. The oolitic iron formation units form part of a sequence of sandstone and shale beds deposited in the beach-lagoon transition zone. The iron ore is aluminous and the silica content in the form of chert is low. Some 70 km S from Paz del Río, there are exposures of the Sabanalarga Formation that contains oolitic iron formation units similar to those observed at Paz del Rio (Angulo, 1978).

The Kiruna-type iron deposits of northern Chile are distributed along a narrow belt striking N-NNE along the Coastal Range between 25°S and 31°S. The strike of this belt roughly coincides with that of the Neocomian magmatic arc. The principal districts (*e.g.*, El Algarrobo-Penoso: 28°47'S; El Romeral: 29°43'S), are situated between 27°S and 30°S and their reserves (before mining) are about 200 Mt (60% Fe) with some 2000 Mt ore for the whole belt. The paragenesis of these Kiruna-type ore deposits include low Ti-magnetite, actinolite and apatite as main species, as well as minor scapolite and a late sulphide phase (pyrite, minor chalcopyrite, etc.). Both, replacement and fracture filling are observed, but replacement is dominant in the larger deposits. Volcanic and sub-volcanic andesitic rocks, intruded by dioritic bodies form the mineralized complexes, probably comagmatic with the andesitic rocks (Oyarzún and Frutos, 1984). The iron-bearing magmatic complexes present horizontal sections that were originally circular (Boquerón Chañar) or ellipsoidal (El Romeral), but were later modified by strike-slip faulting. In general, faulting is intensive and extensive in the iron belt, and dynamic schist of this origin is frequent. At a regional scale, the alignment of the major deposits coincides with a line of pre-Cretaceous crustal weakness that controlled the western border of the Neocomian Basin (the present Atacama Fault Zone).

Hydrothermal alteration is widespread and complex. However, actinolite, partly altered to chlorite, is dominant, followed by silicification and rock bleaching. K/Ar dating of the iron deposits gives ages between 128 Ma (Boquerón Chañar, Zentilli, 1974) and 110 Ma (Los Colorados and El Romeral, Munizaga *et al.*, 1985). Several age determinations at El Algarrobo (Montecinos, 1983) are also in the 128 - 111 Ma range, which coincides not only with the climax of the mafic magmatism, but also with the passage from the Mariana to the Chilean style of oceanic plate subduction (Sillitoe, 1991).

The iron belt also includes smaller iron vein-type deposits as well as a few iron skarns, such as Bandurrias, and some chalcopyrite-magnetite skarn ores, such as San Cristobal, that have been mined for their copper content. Pneumatolytic-hidrothermal fluids were considered as a satisfactory depositional mechanism for the origin of the main iron ore deposits by Ruiz *et al.* (1965), Bookstrom (1977), Oyarzún and Frutos (1984) and other authors, although there are differences of opinion over the source of the fluids. However, Nystrom and Henríquez (1994) and Travisany *et al.* (1995) have recently proposed that these deposits were formed during a magmatic phase and subsequently altered hydrothermally.

The iron deposits of the coastal belt of Peru (Soler *et al.*, 1986; Cardozo and Cedillo, 1990) are similar in mineralogy to the Cretaceous deposits of northern Chile. The principal deposit is Marcona, consisting of stratiform ore lenses, hosted in carbonate and volcaniclastic rocks. According to Atkin *et al.* (1985), their origin is related to replacement by hydrothermal fluids from Middle Jurassic subvolcanic intrusive rocks.

The iron-copper skarns deposits of the Andahuaylas-Yauri Zone in Peru are situated along a belt trending WNW between 13°30'S - 14°30'S and between 71°39'W - 73°39'W. The deposits are associated with quartz monzonite stocks dated at 34 - 33 Ma, that intrude carbonate sediments dated as Albian-Turonian (Noble *et al.*, 1984; Soler *et al.*, 1986). The ores include magnetite with some native gold as early minerals, and chalcopyrite as a later sulphide phase. According to Bellido and Montreuil (1972) they contain the highest potential ore reserves in Peru, estimated at 2000 Mt (60% Fe) by Petersen and Vidal (1996). Among the principal deposits are Huancabamba, Colquemarca, Livitaca and Tintaya. Tintaya is considered to be copper deposit.

The El Laco Kiruna-type iron ore deposits consists of several flow-like and subvolcanic intrusive magnetite-rich bodies with the same mineralogy, that also includes minor apatite. These bodies crop out across a surface of 1.8 km^2 around a Pliocene volcanic centre of northern Chile, close to the border with Argentina (Fig. 1). Pyroxene andesites are dominant in the volcanic flows but a central subvolcanic intrusive is dacitic in composition. The El Laco iron deposit contains several 100 Mt of iron ore, but has not been extensively mined. On the other hand, the peculiar characteristics of the deposits have been the subject of several studies, as well as the origin of controversies regarding their genesis (Park, 1961; Frutos and Oyarzún, 1975; Frutos *et al.*, 1990; Nystrom and Henríquez, 1994; Larson and Oreskes, 1994).

The copper province

Copper deposits occur from the northern to the southern ends of the Andean Belt, and their ages span the

Late Paleozoic to the Pleistocene. The deposits belong to a variety of types, including porphyry copper deposits, enargite-bearing vein deposits and replacement, skarn, breccia pipe, manto-type, massive sulphide, and exotic deposits. In these deposits, copper is associated with a number of metals, including Mo, Fe, Au, Ag, Zn and Pb.

Porphyry copper deposits are also present along the whole Andean Belt (Fig. 2), where they attain the world class, both in tonnage and grade. Some of these, such as the El Salvador deposit (Gustafson and Hunt, 1975), have been studied in great detail, becoming classical examples of their type. Furthermore, the distribution of the deposits along and across the Andean Belt, and the fact that they belong to a wide chronological span, present different level of erosion, and were emplaced in a variety of host rocks under distinct tectonic conditions, have enabled the construction of a number of genetic models. For example, the relationship of porphyry copper deposits and plate tectonics, as well as studies on the top and base of porphyry systems have been described by Sillitoe (1972) and Sillitoe (1973), respectively. Nevertheless, it is difficult to make a synopsis due to the abundance of important deposits and studies on these. The work edited by Camus *et al.* (1996) as well as the paper by Sillitoe (1992) are highly recommended.

Sillitoe (1988) considered six epochs of porphyry copper mineralization in the Chilean-Argentinean sector of the Andes, from the Late Carboniferous-Early Permian to middle Miocene-early Pliocene, and six epochs, from the Jurassic to middle Miocene-early Pliocene, for the Peru to Colombia Andean sector. Longitudinal belts up to 100 km wide in which other types of ore deposit also occur representing each of these epochs. However, if we consider only those porphyry copper that have been selected for high tonnage mining operations, the field is geographically restricted to the sector between 10°S and 35°S and chronologically to those deposits of Tertiary age. These alone account for about 25 to 30% of the world's reserves and current production of both copper and molybdenum. As shown in Figure 12, this sector closely coincides with the Andean segment that has a thickest continental crust. The larger porphyry copper deposits of the segment, such as Chuquicamata and El Teniente, have ore reserves (before mining) up to 50 Mt of metallic Cu (Oyarzún and Frutos, 1980).

Most porphyry copper deposits in the Andes are related to dacitic-granodioritic porphyric stocks, emplaced in volcanic rocks or in intrusive complexes. Although the stocks generally belong to the calc-alkaline series, shoshonitic or high-K calc-alkaline rocks have been found at the Farallon Negro District (Sasso and Clark, 1988). Sr isotope ratios of the porphyric stocks are low and point to a deep-seated origin. Besides which their Pb isotope ratios have a narrow range. Thus, in the case of Chuquicamata and El Salvador deposits, Pb isotope ratios are similar to those of the Southern Volcanic Zone of the Andes, the magmas of which were not affected by crustal contamination (Zentilli *et al.*, 1988). Furthermore, there is evidence to suggest that the magmas responsible for the porphyric systems, ascended rapidly through the crust, permitting a degree of contamination that was small or nihil (Maksaev and Zentilli, 1988). In general, the emplacement-alteration-mineralization process can be generalized as a subvolcanic magmatic development of a metal-rich magma, where residuals fluids mixed with meteoric waters during the late stage of its cooling, as stated by Ambrus (1988). Although the majority of the deposits fit well into the Lowell and Guilbert (1970) model, the phyllic zone is essentially absent in some of them, including El Abra and El Teniente. El Teniente has been recently reinterpreted in terms of the intrusive emplacement of a high-K, ore bearing, mafic magma (Skewes and Arévalo, 1997), but fits better into the dioritic model proposed by Hollister (1974).

Porphyry copper deposits present both spacial and chronological clusters in the Andean Belt. Thus, the Arequipa Lineament includes four important Paleocene deposits (Cerro Verde, Cuajone, Quellaveco and Toquepala) occurring along a 150 km long narrow belt in southwestern Peru (Fig. 2). Also, six major Eocene-Oligocene deposits, including Chuquicamata, El Abra and Collahuasi, are distributed along a N-S lineament extending for some 125 km line, following the important Domeyko Fault System. Other important cluster is that of the Los Bronces-Río Blanco and El Teniente deposits of Plicene age that occur about 130 km to the S of the six major Eocene-Oligocene deposits cited above.

As pointed out before, many important porphyry copper deposits in the Andes lie in or close to large fault zones (Fig. 10). However, although this structural control is evident for those deposits of the stockwork-type, such as Chuquicamata, this is not the case for porphyry deposits of the breccia pipe-type, such as Los Bronces-Río Blanco or El Teniente (Camus, 1975).

The Andean porphyry copper deposits have Mo content that ranges between 0.01% and 0.1%, and this metal follows copper in economic importance. Given the large tonnages of porphyries like Chuquicamata and El Teniente, they also rank among the major Mo deposits of the world (Ambrus, 1988). On the other hand, the gold content is rather low, with the important exception of the Farallon Negro District in Argentina, where Bajo de la Alumbrera deposit contains 780 Mt of ore, at 0.52% Cu and 0.67 g/t Au (Sasso and Clark, 1988).

Although the enargite-bearing vein and replacement Cu +/- Au, Ag, Zn, Pb deposits are better represented in Peru, they are also common in other parts of the Tertiary volcanic belts of the Andes. However, Petersen and Vidal (1996) observed that the number of large and high-grade enargite-bearing deposits is an unusual feature of the Peruvian metallogeny. The Peruvian deposits are well zoned from Cu and Au in the centre to Zn and Pb at the margins. Among the principal enargite-bearing deposits in Peru are Quiruvilca, Cerro de Pasco, Colquijirca, Huarón, Morococha, Yauricocha and Julcani (Figs. 2 and 5). The rich vein-gold deposit of El Indio, Chile (Fig. 4) also belongs to the enargite-bearing type. According to Sillitoe (1983), enargite-bearing massive sulphide deposits may represent the upper levels of porphyry copper systems.

Peru is also richly endowed with Cu +/- Fe, Au, Zn deposits related to calcic skarns, partly as a consequence of the broad distribution of Mesozoic back-arc carbonate rocks, that host Tertiary monzonitic granitoid plutons (Fig. 7). As mentioned before, some skarn deposits of the Andahuylas-Yauri Zone are also important for their magnetite content. Among the major skarn deposits in Peru, stand out Antamina, Cobriza, Ferrobamba and Tintaya (Petersen and

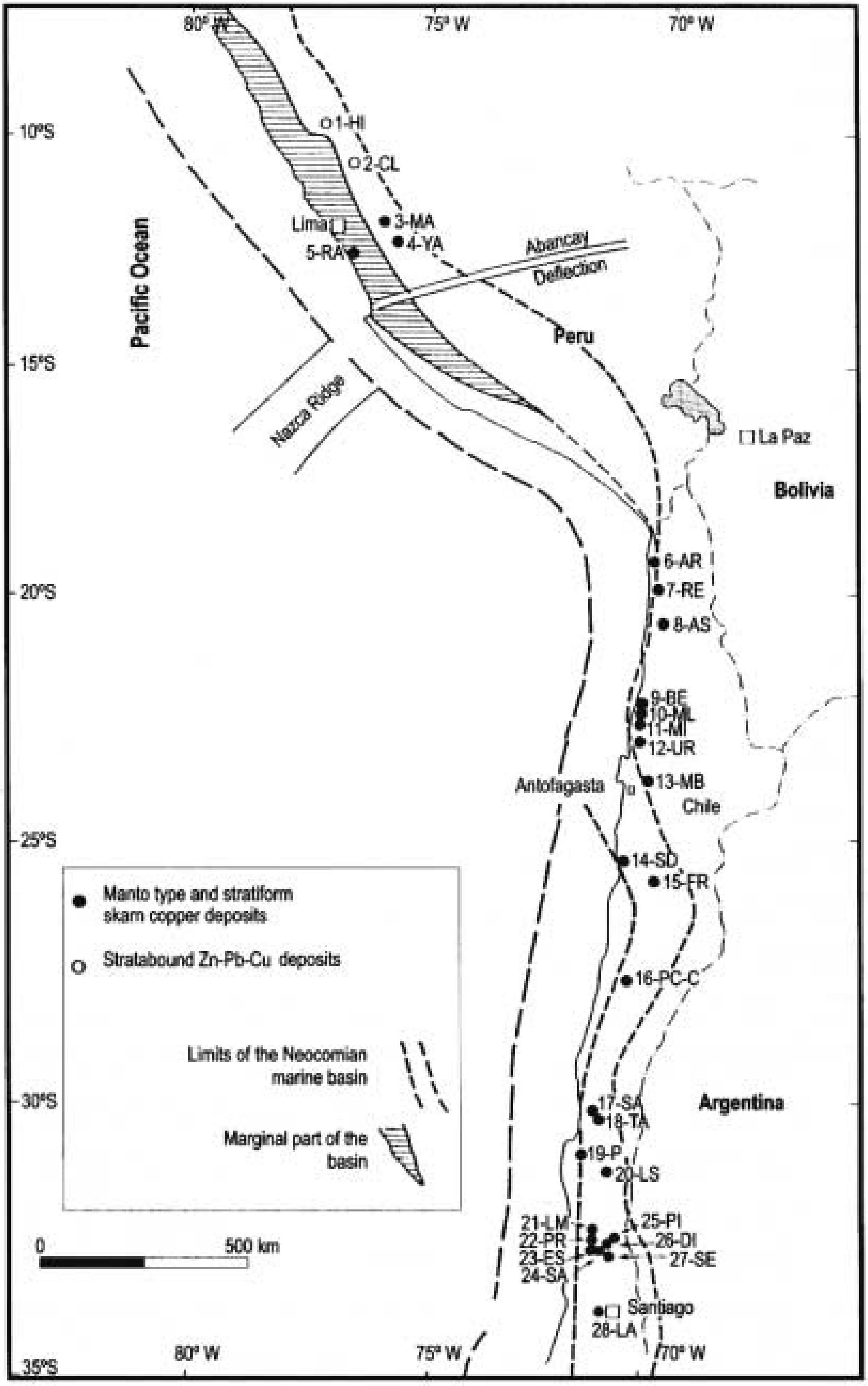


FIGURE 2 - Manto-type and stratiform skarn copper deposits in Mesozoic volcano-sedimentary rocks of Peru and Chile. List of deposits: 1- Huanzala. 2- Colquijirca. 3- Manon. 4- Yauricocha. 5- Raúl. 6- Argolla. 7- Regina. 8- Asturias. 9- Buena Esperanza. 10- Mantos de la Luna. 11- Michilla. 12- Urbina. 13- Mantos Blancos. 14- Santo Domingo. 15- Frankestein. 16- Punta del Cobre-Candelaria. 17- San Antonio. 18- Talcuna. 19- Panulcillo. 20- Los Sapos. 21- Los Maquis. 22- Peumo-Rusa. 23- El Soldado. 24- El Salado. 25- Pirquitas. 26- Diablo. 27- Sauce. 28- Lo Aguirre (modified after Bellido et al., 1972; Frutos et al., 1990; Ruiz et al., 1965).

Vidal, 1996). A second type of skarn deposit, the amphibolitic Cu +/- Fe skarn deposit (Vidal *et al.*, 1990), is represented in Peru by Raul - Condestable, and in Chile by Candelariá (Fig. 3). Both these deposits are related to the Lower Cretaceous basins, and have mineralogical analogies comparable to the Kiruna-type iron deposits of Peru and Chile.

Breccia pipe ore deposits are widespread in the Andes. Although many of them are related to porphyry copper systems, others appear as independent mineralizations, and exhibit much variety with respect to the diameter of the pipe as well as in the number of deposits in a given district. The mineralogy of the deposits is generally cupriferous (with Au) or polymetallic. A detailed description of Cu-bearing tourmaline breccia pipes in Chile was made by Sillitoe and Sawkins (1971).

Manto-type copper deposits are typically found in volcano-sedimentary formations of Mesozoic age in northern and central Chile (Espinoza *et al.*, 1996). The deposits are stratiform or stratabound, but frequently also include veins, ores in breccias, and stockworks that are probably co-genetic (Vivallo and Henríquez, 1998). Their paragenesis is rather simple and includes chalcocite, bornite, chalcopyrite, pyrite and hematite, the Cu/Fe ratio decreasing outward from the Cu-rich cores. The stratiform Cu mineralization, that also contains some g/t Ag, was deposited in the groundmass and vesicles of lava flows or in the matrix of pyroclastic rocks. The associated hydrothermal alteration is propylitic and includes albite, chlorite and calcite. Mineralization occured in the epithermal or low mesothermal range. These deposit have magnitudes up to hundred Mt ore, containing 1 - 2% Cu (El Soldado), but normally they are in the 1 - 10 Mt ore range (Camus, 1985). Some typical deposits of this group are Buena Esperanza, Carolina de Michilla, Talcuna, and Lo Aguirre (Fig. 3).

Massive sulphide deposits are not abundant in the Andean Belt, although the accreted oceanic prisms of the Northern Andes offer favourable environments for Cyprus-type deposits, and a few are known in western Colombia (Ortiz, 1990). Also, an important Fe-Cu-Zn volcanogenic massive sulphide deposit, Tambo Grande, is situated in the north-western corner of Peru, at 5°S. In Chile, the manto-type Cu deposits at Punta del Cobre (Fig. 3) and the polymetallic skarn of El Toqui, at 45 °S have been interpreted as massive sulphide deposits by Camus (1985) and by Wellmer *et al.* (1983), respectively.

Favourable climatic and tectonic conditions for the formation of exotic Cu deposits, existed in the Andes of southern Peru and northern Chile between 12°S and 27°S (Munchmayer, 1996). In Chile, twelve deposits of this type are already known. The largest deposit, Exótica, was deposited on a wide paleochannel, 2 to 4 km S of Chuquicamata, and the source of the Cu mineralization, contained some 3 - 4 Mt of metallic Cu (before mining). Similar figures (1.2 to 3.5 Mt metallic Cu) are given by Munchmayer (1996) for the Damiana deposit, on the western slope of Cerro Indio Muerto (El Salvador porphyry copper district). In general, exotic Cu deposits formed by lateral migration of supergene solutions from porphyry copper deposits. Their mineralogy (chrysocolla, atacamite, Cu-wad) was controlled by Eh - pH conditions of ground water, and their shape by the former topography around the porphyry. According to Munchmayer (1996), the exotic mineralization episodes mainly occurred during the lower Miocene.

Copper vein deposits are widespread in the Andean Belt, being difficult to present a synthesis of this subject. However, it is important to state that Cu mining in the Andes began with this type of deposit. In northern Chile, favourable climatic and tectonic conditions produced a high degree of supergene enrichment in Cu +/- Au vein deposits, allowing for the development of a highly profitable mining industry during the 19[th] century.

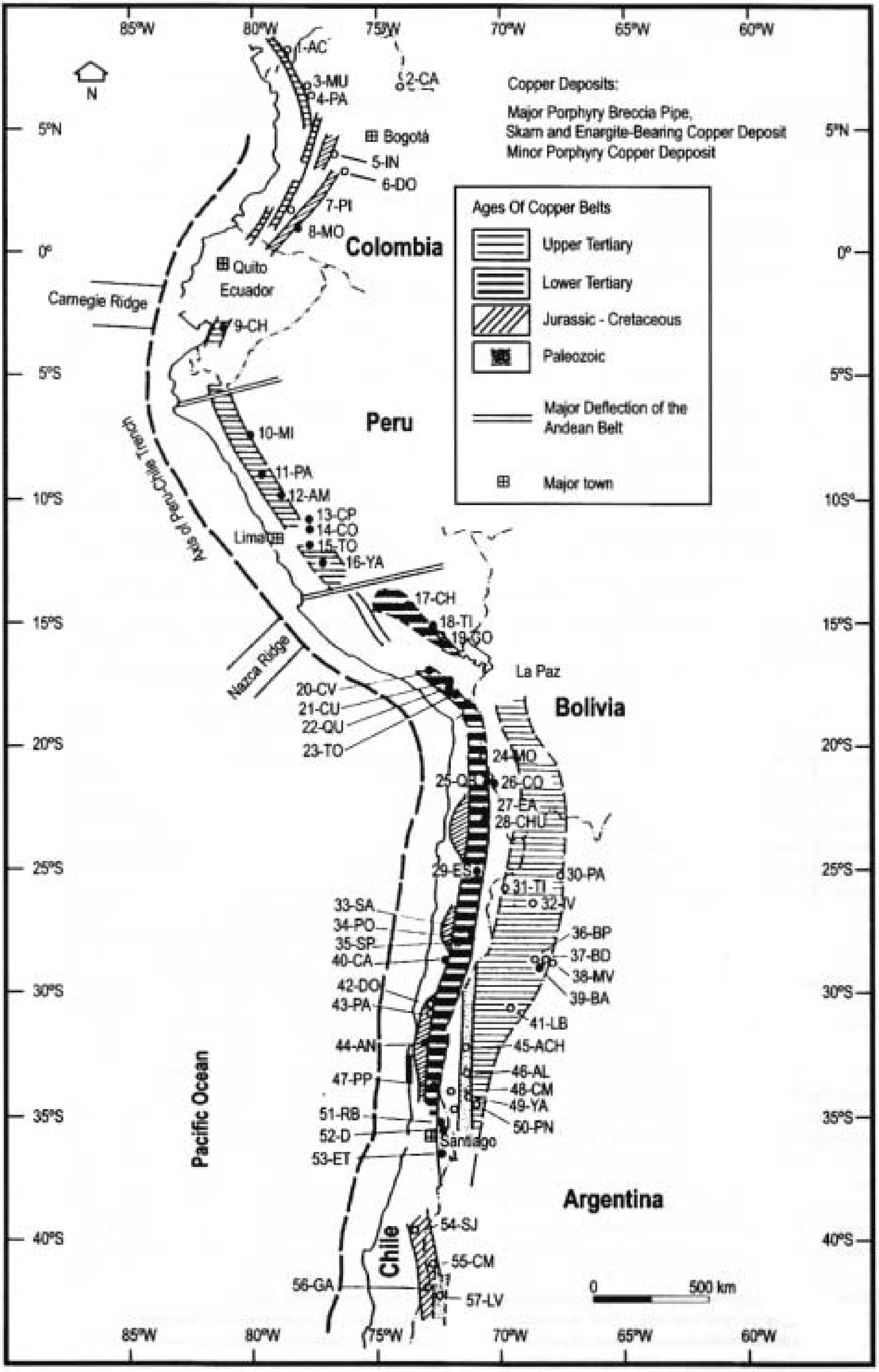
Gold and silver metallic belts

Gold and silver were main lures for the Spanish conquerors in the Andean countries, and hidden deposits of these metals, together with those of copper, are today the prime target for the exploration and mining companies.

Colombia has been an important gold producer in the northern Andes, and the first in the world in colonial times. Although Colombia's gold production comes mainly from placer and vein-type deposits, there are also several lode gold deposits such as those of California, Segovia, Frontino and Marmato (Fig. 4). Some of these lode deposits, such as California and Marmato are related to porphyry copper systems (Sillitoe *et al.*, 1982). By contrast, there are no important silver deposits in Colombia, and this metal is mined as a by-product of gold mining. It is interesting to remember that platinum was first discovered in Colombian placers, and that it was from Colombia that all the world's platimum was obtained until 1819 (Angulo, 1978).

Gold mining began in colonial times in Ecuador at the famous Portovelo Deposit (Fig. 4) as well as in many small Au-Ag veins and placer gold workings. According to Gemutz *et al.* (1992), gold deposits and prospects in Ecuador are associated with the epithermal veining (Portovelo, Pilzhum, and Molleturo), skarn-type deposits (Nambija and Pachicutza), stockwork-vein deposits (Chinapitza), intrusive breccia deposits (Gaby) and porphyry copper (Fierro Urco) deposits, in addition to placer deposits. Most of these are Tertiary in age, but a few are Jurassic (Nambija, Chinapitza). In like manner to Colombia, silver is subordinate to gold in most of the Ecuadorian precious-metal deposits.

A review of the gold deposits in Peru was presented by Noble and Vidal (1994). Peru has a long and important history as a gold and silver producer that began in pre-Hispanic times. Noble and Vidal (1994), classified the Peruvian gold deposits (Figs. 2, 5 and 7) into the following groups: 1 - Quartz veins of Paleozoic and Mesozoic age: a) Pataz-Buldibuyo Belt (Pataz, Parcoy, etc.); b) Santo Domingo-Ananea Region (Ananea, Santo Domingo, etc.); c) Nazca-Ocoña Belt (Calpa, Ishihuinca). 2 - Gold bearing systems of Cenozoic age: a) Au-bearing porphyry and skarn-type deposits (Michiquillay, Tintaya, etc.); b) Sedimentary rock-hosted gold (Yauricocha, Utupara, etc.); Polymetallic and precious metal deposits, subdivided into: Polymetallic systems (Quiruvila, Sayapullo, etc); Epithermal deposits of the adularia-sericite type Ag-Au vein systems (Cailloma, Arcata, etc.); and high-level, acid-sulphate systems (Yanacona, Ccarhuaraso, etc.). At Julcani, the acid-sulphate stage was developed between two stages of adularia-sericite


Copper Deposits:
Major Porphyry Breccia Pipe,
Skarn and Enargite-Bearing Copper Deposit
Minor Porphyry Copper Depposit
Ages Of Copper Belts
Upper Tertiary
Lower Tertiary
Jurassic - Cretaceous
Paleozoic
Major Deflection of the Andean Belt
Major town
Colombia
Ecuador
Peru
Bolivia
Argentina
Chile
Pacific Ocean
Carnegie Ridge
Nazca Ridge
Axis of Peru-Chile Trench
Bogotá
Quito
Lima
La Paz
Santiago
N
500 km

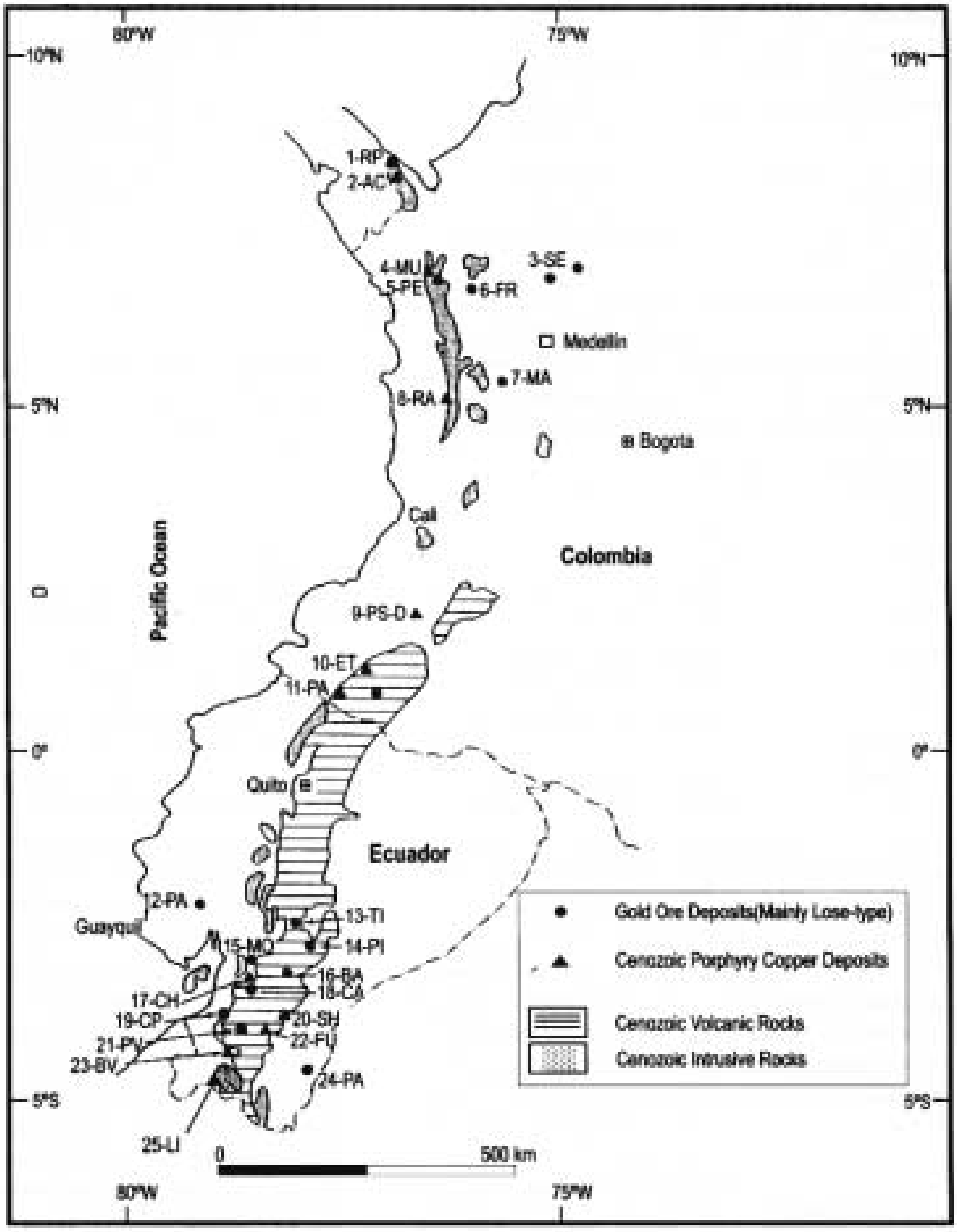


FIGURE 3 - Porphyry, breccia pipe, skarn and enargite-bearing copper deposit in the Andean Belt. List of deposits: 1- Acandí. 2- California. 3- Murindó. 4- Pantanos-Pegadorcito. 5- Infierno-Chili. 6- Dolores. 7- Piedrasentada. 8- Mocoa. 9- Chaucha. 10-Michiquillay. 11- Pashpap. 12- Antamina. 13- Cerro de Pasco. 14- Colquijirca-Huarón. 15- Toromocho. 16- Yauricocha. 17- Chalcobamba. 18- Tintaya. 19- Coroccohuayco. 20- Cerro Verde-Santa Rosa. 21- Cuajone. 22- Quellaveco. 23- Toquepala. 24- Mocha-Queen Elizabeth. 25- Quebrada Blanca. 26- Rosario-Collahuasi. 27- El Abra. 28- R. Tomic-Chuquicamata-M.M. 29- Escondida-Zaldívar. 30- Pancho Arias. 31- Taca-Taca. 32- Inca Viejo. 33- El Salvador. 34- Potrerillos. 35- San Pedro de Cachiyuyo. 36- Bajo de Pampitas. 37- Bajo del Durazno. 38- Mi Vida. 39- Bajo de la Alumbrera-Farallón Negro. 40- Cabeza de Vaca. 41- Los Bayos. 42- Domeyko. 43- Pajonales. 44- Andacollo. 45- Arroyo Chita. 46- Alcaparrosa. 47- Los Pelambres-Pachón. 48- Cerro Mercedario. 49- San Jorge-Yalguaraz. 50- Paramillos Norte. 51- Río Blanco. 52- El Bronce (Disputada). 53- El Teniente. 54- Estero San José. 55- Campana Mahuida. 56- Galletué. 57- La Voluntad (modified after Sillitoe, 1988).

FIGURE 4 - Gold and porphyry copper deposits in Colombia and Ecuador (Northern Andes). List of deposits: 1- Río Pito. 2- Acandi. 3- Segovia. 4- Murindó. 5- Pegadorcito. 6- Frontino. 7- Marmato. 8- Río Argueda. 9- Piedrasentada-Dominical. 10- El Tambo. 11- Piedrancha. 12- Pascuales. 13- Tixan. 14- Pilzhum. 15- Molleturo. 16- Baños. 17- Chaucha. 18- Cascayo. 19- Cerro Pelado. 20- Shingata. 21- Portovelo. 22- Fierro Urco. 23- Buena Vista. 24- Pachicutza district. 25- Los Linderos (modified after Barrero, 1976; Gemutz et al., 1992; Goosens, 1969; Sillitoe et al., 1982).

alteration. 3 - Bulk mineable ores (Yanacocha, Hualgayoc). 4 - Quaternary placer deposits.

Although Peru ranks third in present gold production among the Andean countries (after Chile and Colombia), this situation should change, due to the development of a number of important mining projects, such as the Barrick's Pierina Mine near Ancash. The Pierina Mine is scheduled to produce 22 t Au/year (equivalent to total gold production of Peru in 1993).

Silver is also an abundant metal in many hydrothermal deposits in the volcanic rocks of the Western Cordillera of Peru, appearing in independent primary (argentite, proustite) or in secondary (native Ag, acantite) minerals, as well as in inclusions of silver minerals or as solid solution in galena and Cu sulphasalts (tetrahedrite). However, in sediment-hosted deposits in the western and eastern cordilleras silver is commonly found only in solid solution or as inclusions in galena and sulphasalts (Bellido and Montreuil, 1972). The principal Ag-rich deposits are polymetallic Quiruvilca deposit (Ag/Au = 100); the ephithermal San Juan de Lucanas deposit (Ag/Au = 160); the deposits of the María Luz-Huachacolpa District (Ag/Au = 450); and the Julcani deposit (Ag/Au = 65) (Noble and Vidal, 1994).

The Miocene sub-volcanic deposits of the central and southern part of the Cordillera Real, W of the Altiplano Region of Bolivia, are best known for their Sn-Ag veins as well as for the Sb vein deposits. However, they are also related to polymetallic veins and stockworks in the boundary zone with the Altiplano Region. Of the polymetallic districts, La Joya (Long *et al.*, 1992) has shown an important potential as a gold and silver producer with reserves before mining of 10 Mt of oxide ore, containing 1.65 g/t Au and 20 g/t Ag. The ores are present in fractures of an intrusive dacitic body and were deposited at high temperatures (300 - 550° C) from highly saline fluids. They are distributed in four hills: Kori Kollo, Llallagua, Quiviri and La Joya.

Although there are important Au-Ag deposits in Chile, most of them linked to sub-volcanic magmatic activity of Miocene age in the high Andes, Au is associated preferentially with Cu and Fe in many of them. In addition to the Au deposits, there are many important Ag deposits of the Guanajuato type that are almost devoid of Au content.

Gold production in Chile attained a peak in 1938 with some 11 t Au, mostly coming from placer deposits. Thereafter, production then gradually decreased to 2 - 3 t/year between 1955 and 1970. A rapid increase to 6 t began in 1980, passing to 12 t in 1981, and to 18 t in 1982, as a result of the discovery and development of the high grade enargite-bearing gold deposit of El Indio. This was the first of a series of discoveries in the Tertiary Volcanic Belt of the Andes between 26°S and 31°S (Cuadra and Dunkerley, 1991; Sillitoe, 1991). In 1996, gold production was 53 t.

Chilean hydrothermal gold deposits are Jurassic to upper Miocene in age, and the mineralization occurs in hydrothermal breccia, veins, stockworks and disseminations (Sillitoe, 1991). Although most of the Au +/- Cu deposits correspond to Mesozoic pluton-related veins, only two districts, Los Mantos de Punitaqui and El Bronce (Fig. 5) had Au content over 10 t. The rest of the deposits over 10 t Au were classified by Sillitoe (1991) in four types: 1 - High sulphidization, epithermal (Choquelimpie, Guanaco, El Hueso, La Coipa, La Pepa, Nevada/Pascua and El Indio-Tambo). 2 - Low sulphidization, epithermal (Faride, San Cristobal, Fachinal). 3 - Porphyry-type (Marte, Lobo, Refugio). 4 - Distal contact, metasomatic (Andacollo). With the important exception of the Andacollo Deposit of Cretaceous age (Reyes, 1991; Oyarzún *et al.*, 1996), all are Tertiary. As pointed out by Gemutz *et al.* (1992) as well as by other authors, El Indio, Nevada and the Maricunga District (La Coipa, La Pepa, Marte, Lobo, Refugio, etc.) are situated over an Andean segment having a flat subduction zone that also includes the Au-rich porphyries of the Farallón Negro District in Argentina (Sasso and Clark, 1988).

Of those deposits containing more than 10 t Au listed before, only six deposits have Ag/Au ratios over 10 (Choquelimpie, Faride, San Cristóbal, El Guanaco, La Coipa and Fachinal). La Coipa (Ag/Au = 98), is properly an Ag-Au deposit.

Chile was an important silver producer in the 19th century (300 t in 1873, 15% of the total world production). Among the principal silver districts are those of Huantajaya, Caracoles, Tres Puntas, Chañarcillo and Agua Amarga (Fig. 6). They are epithermal, low sulfidation vein-type deposits, hosted by stratified rocks belonging to the volcano-sedimentary transitional facies of the Jurassic and Cretaceous back-arc marine basins. Silver mineralization includes a variety of sulphide species (argentite, proustite, pyrargirite), and supergene processes are responsible for the deposition of secondary minerals (native Ag, cerargyrite) in very rich oxidation zones (Ruiz *et al.*, 1965).

A review of precious and base metal deposits in Argentina by Gemuts *et al.* (1996) mentions the Paramillos, (Mendoza) silver deposit and the Gualilán gold deposit as the older mines in Argentina (Gualilán dates from the 17th century). Modern exploration pre-1960 was centred in high-grade precious and base metal deposits such as Mina Angela (Ag-Pb-Zn-Au vein), Farallón Negro (Mn-Ag-Au vein) and El Aguilar, a SEDEX massive sulphide deposit in the Province of Jujuy. After 1960, a series of porphyry copper prospect were found and drilled. Several of them are now the basis for an important mining industry in Argentina. Also, after the discovery of El Indio, the Argentinean border of the Andes was rapidly explored and a number of precious metal deposits were discovered (some of them very close to the Chilean ones, *e.g.*, Lama in the vecinity of the Pascua deposit). Besides which, a group of government geologists discovered the Cerro Vanguardia and El Dorado districts in the El Deseado Massif (Province of Santa Cruz) in southeastern Argentina. Both district include epithermal, low sulphidization, Au-Ag and polymetallic veins. They are related to a phase of Middle Jurassic acid magmatic in the pre-rift tectonic conditions associated with the break-up of Gondwana (Giacosa *et al.*, 1988; Echavarría and Etcheverry, 1998).

The polymetallic province

The polymetallic province (Fig. 11) is present along the entire Andean Belt, although the principal deposits are located in the Peruvian segment, which also has thick and widespread carbonate sedimentary strata. Although Paleozoic deposits are known, some of them important like the Zn-Pb-Cu deposit

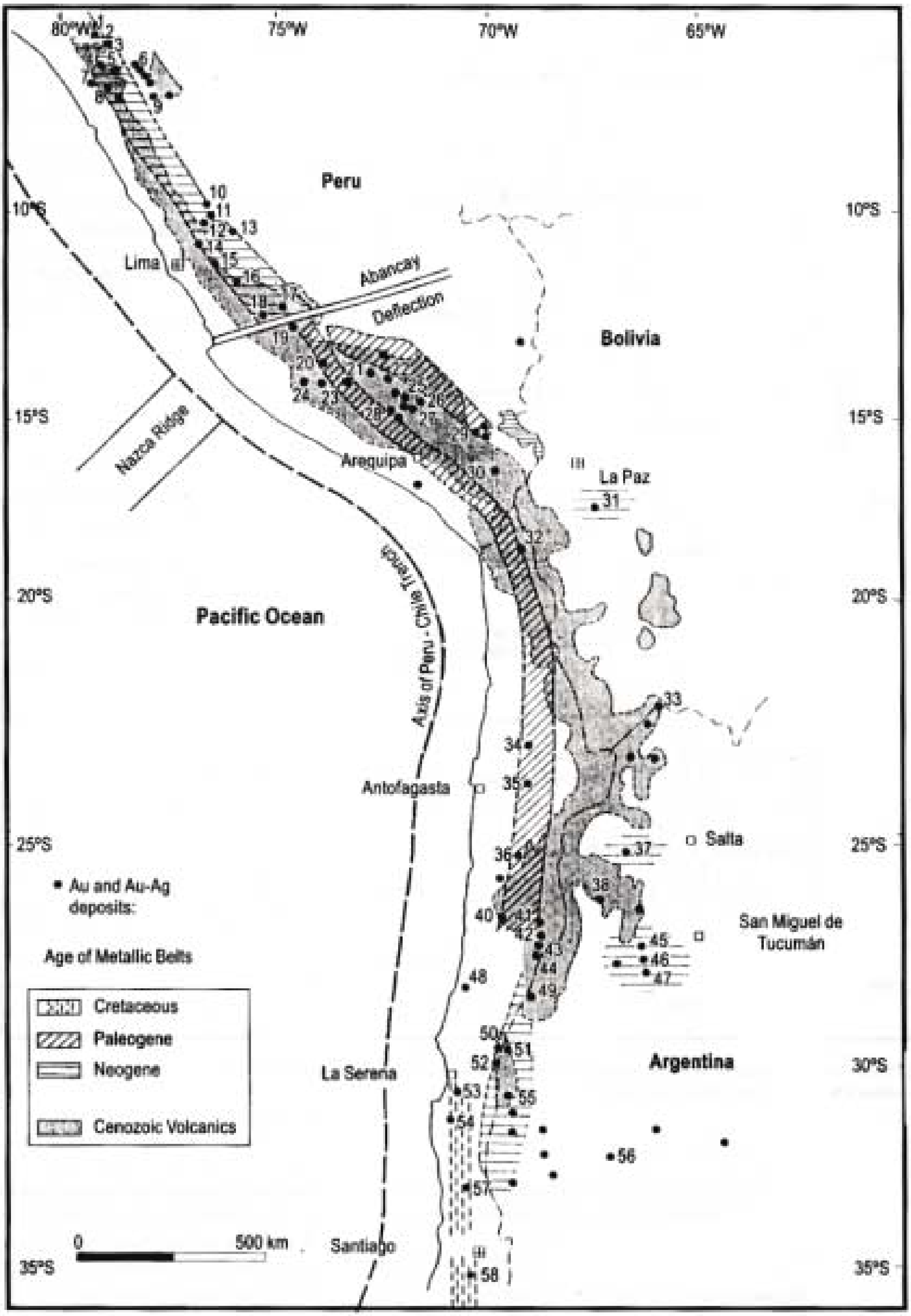


FIGURE 5 - Hydrothermal gold deposits in Peru, Bolivia, Chile and Argentina. List of deposits: 1- Hualgayoc. 2- Yanacocha. 3- Minas Conga. 4- Sayapullo. 5- Algamarca. 6- Pataz. 7- Milluachaqui. 8- Quiruvilca. 9- Parcoy. 10- Cerro de Pasco. 11- Colquijirca. 12- Huarón. 13- Chanchamina. 14- Colqui. 15- Millotingo. 16- Yauricocha. 17- Julcani. 18- Huachacolpa. 19- Antasulla. 20- Ccarhuaraso. 21- Utupara. 22- Cerro Millo. 23- Palla-Palla. 24- San Juan de Lucanas. 25- Arcata. 26- Sucuitambo. 27- Caílloma. 28- Orcopampa. 29- Santa Lucía. 30- Cucachara. 31- Kari-Kollo. 32- Choquelimpie. 33- Pirquitas. 34- Faride. 35- San Cristóbal. 36- Guanaco. 37- Diablillos. 38- Sierra Overa. 39- Incahuasi. 40- Inca de Oro. 41- La Coipa. 42- Marte-Lobo. 43- La Pepa. 44- Refugio. 45- Farallón Negro. 46- Bajo de la Alumbrera. 47- Capillitas. 48- Capote. 49- Laguna Verde. 50- Pascua. 51- Lama. 52- El Indio-Tambo. 53- Andacollo. 54- Punitaqui. 55- Río Frío. 56- Cerro Blanco. 57- El Bronce de Petorce. 58- Alhué (modified after Arenas, 1988; Gemutz et al., 1996; Noble and Vidal, 1994; Sillitoe, 1988, 1991).

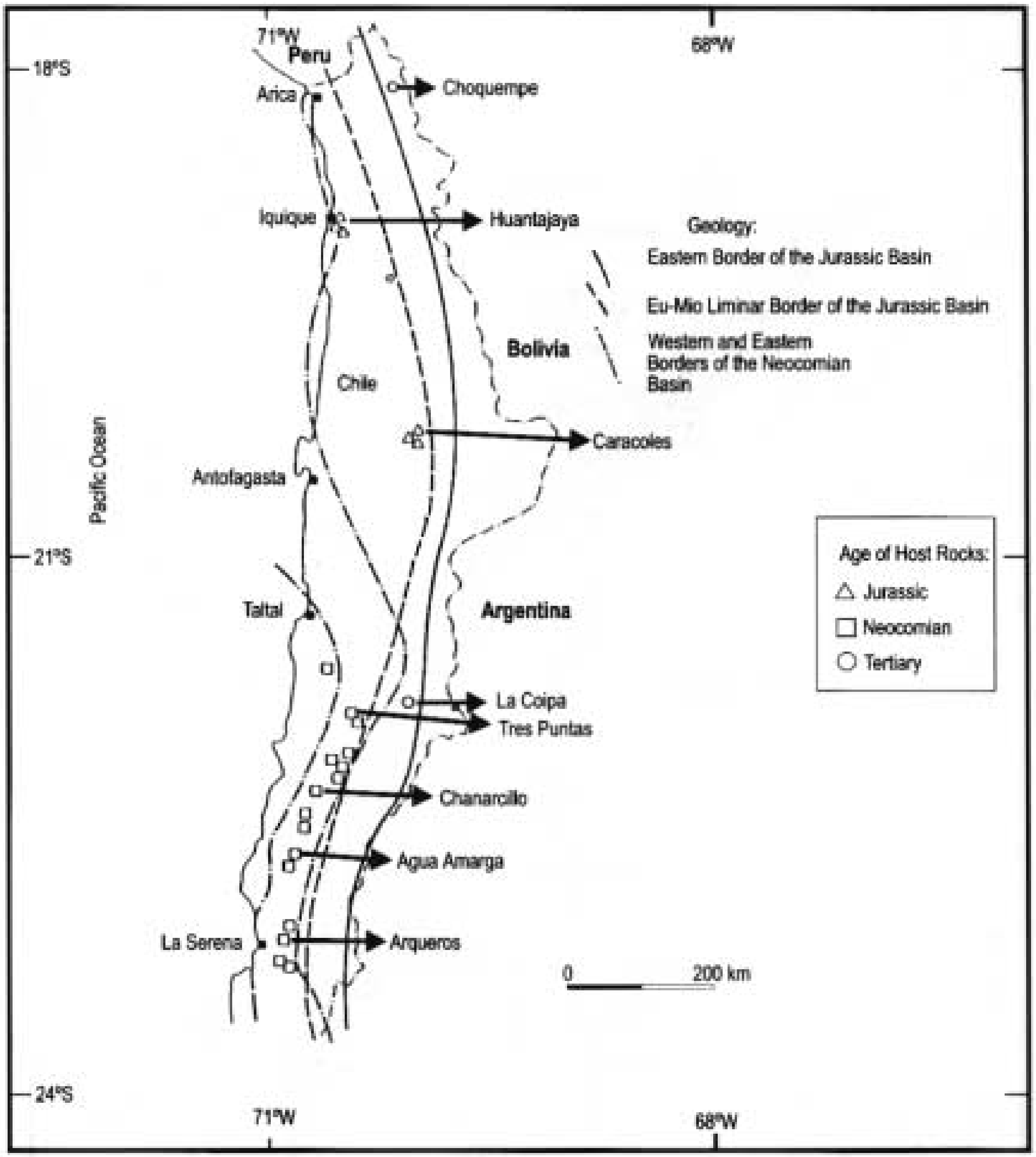


FIGURE 6 - The distribution of silver ore deposits in Chile and some major paleogeographic features (modified after Frutos, 1975; Ruiz et al., 1965).

of Los Bailadores, in the Sierra Nevada, Venezuela (Carlson, 1977) or El Aguilar in northwestern Argentina, most of the deposits are Mesozoic or Cenozoic in age.

El Aguilar (23°13'S / 65°42'W); a Pb-Zn-Ag SEDEX deposit in Ordovician quartzite beds, represents the largest Paleozoic Pb-Zn concentration in South America (Sureda and Martin, 1990), with some 30 Mt ore (12% Pb+Zn; 100 g/t Ag). The fact that a Cretaceous plutonic intrusion thermally modified the original deposit and some skarn-type ore bodies were formed obscured the genesis of the deposit, now well established as being of the SEDEX-type. Other Pb-Zn-Ba ores in Ordovician clastic sediments are those of Pumahuasi (22°17'S / 65°33'W). They are part of a belt that continues for some 500 km northwards to the Sucre Zone in Bolivia (Sureda *et al.*, 1986).

Although Mesozoic and Cenozoic polymetallic deposits are present in the Northern Andes (Colombia and Peru), most of these vein-type deposits have been mined for silver. Polymetallic deposits are poorly represented in Chile, except for the Patagonian Cordillera, between 46°00'S and 47°20'S. Thus, the clastic-carbonate rocks that interfingered with andesitic volcanics of the Jurassic and Lower Cretaceous back-arc basin, mainly host epithermal silver veins or skarn-type Cu or Fe deposits. On the other hand, a rich polymetallic province developed in Peru that may be partly explained by the widespread distribution of Mesozoic marine sediments, including abundant carbonate facies (Fig. 7).

During the Late Triassic, the sea advanced from the N, and reached 13 °S (Audebaud *et al.*, 1973), covering the domain of the Pucará Basin, a NW-SE trending belt between 76°W - 77°W at 9°S and 72°W - 74°W at 14°S, where clastic and carbonate sediments were deposited. To the W, the basin also received andesitic lavas. The marine sedimentation continued during the Lias, when the basin was divided in two sectors (N and S). These sectors were united in the Dogger, and separated again during the Malm by a major NW-SE trending positive block. During the Malm and the Early Cretaceous, marine sedimentation continued in association with andesitic volcanics only in the southwestern part of the basin. However, a new marine transgression during the Albian, the sea coming this time from the S, covered the area now underlain by the western and eastern cordilleras of Peru, where the sea remained until the Late Cretaceous (Senonian). Thus, paleogeographic conditions were favourable for the deposition of carbonate rocks in Peru. On the other hand, contemporary basins in Bolivia received only clastic sediments, except for some carbonates of Campanian-Maastrichtian age (Pareja *et al.*, 1978).

Rich stratiform polymetallic deposits, with very high Zn grades, are found in the sedimentary rocks of the Triassic-Liassic platform of the Pucará Basin (Amstutz and Fontboté, 1987; Cardozo and Cedillo, 1990). These are, in part, of the Mississippi Valley-type, such as San Vicente, situated in the eastern part of the basin, and Shalipayko in the western part. In the western part there are also some deposits that show volcanic affinities, *e.g.*, Carahuacra, San Vicente. The largest Zn producers in Peru have been sediments developed on tidal flats, in lagoons, and in carbonate reefs. The Cercapuquio Pb-Zn stratiform deposit in central Peru (Cedillo, 1990), hosted by lagoonal sediments of Late Jurassic age, also exhibits strong similarities to the Mississippi Valley-type deposits.

About 80 stratabound Zn-Pb (Ag-Cu) ore deposits and prospects are known in the Santa Formation (Valanginian to Aptian), deposited in an ephemeral basin (Cardozo and Cedillo, 1990). Among the principal deposits are Huanzala (Fig. 7) and El Extraño (9°09'S / 78°05'W). Several features of these ore deposits indicate a syn-diagenetic origin, *e.g.*, the presence of rhytmites involving the ore minerals (Samaniego, 1980). However, there is evidence of *hydrothermal activity and contact metamorphism affecting* the deposits.

The stratabound ore deposits of the Casma Formation (Middle Albian) are rich in sphalerite and barite and have minor Cu, Pb and Ag content. The principal deposit of this group, Leonila Graciela (Vidal, 1987), at 11°51'S / 76°37'W, is hosted by altered volcano-sedimentary rocks.

Lead-zinc (silver) stratabound deposits are hosted by Upper Cretaceous carbonate rocks at Hualgayoc (Fig. 7), in the Western Cordillera of northern Peru (Cardozo and Cedillo, 1990). Many of the deposits are in the Chulec Formation (*e.g.*, Carolina, Porica), as well as in the Pulluicana Formation (*e.g.*, Yanacancha, Quijote). Although mined since colonial times for their silver ores, the deposits of the Hualgayoc District were later mined for their polymetallic ores (Zn, Pb, Cu, Ag) beneath the oxidation and supergene enrichment zones. As pointed out by Canchaya (1990), the origin of the stratabound deposits of the district remains obscure, in spite of the large number of geological studies that have been carried out.

In northwestern Argentina, there are a number of polymetallic (Cu, Pb, Zn) stratabound sulphide ore deposits in carbonate rocks of Late Cretaceous-Early Tertiary age (Sureda *et al.*, 1986). The deposits are dispersed along a narrow N-S belt about 150 km long between 24°10'S / 64°23'W and 25°15'S / 65°06'W. However, both their tonnage and grade are low.

The major enargite-bearing stratabound Cu-Pb-Zn-Ag deposit of Colquijirca (Fig. 7) some 8 km S of Cerro de Pasco is hosted by the La Calera Series (Tertiary), consisting of clastic and carbonate sediments, with chert and tuffaceous intercalations. Although this deposit has been traditionally classified as a hydrothermal replacement deposit (Mc Kinstry, 1936), Lehne (1990) proposed a syngenetic origin, considering the bedding and other sedimentary features of the ores. The thickness of the ores beds is normally less than 2 m and they are separated from each other by shale beds.

Most of the hydrothermal polymetallic deposits in Peru (Soler *et al.*, 1986; Cardozo and Cedillo, 1990) are associated *with subvolcanic intrusives of Miocene age in the northern* and central part of the country. Although it is possible that some of the deposits considered as Miocene, such as Uchucchacua are late Eocene-early Oligocene in age (Soler and Bonhomme, 1988, *apud* Cardozo and Cedillo, 1990), the Miocene remains as a principal metallogenetic period for this and other types of ore deposits. Cardozo and Cedillo (1990) classify the hydrothermal polymetallic deposits of Miocene age in four groups: 1 - Complex deposits, including both replacement and veins. They are normally zoned and rich in Cu-As sulphosalts. Cerro de Pasco, Huarón, and Morococha are included in this group. 2 - Skarn bodies, some of which are associated with veins, such as Santander and Milpo-Atacocha. 3 - Veins, hosted by Mesozoic sedimentary rocks and Oligocene-Miocene volcanics, *e.g.*, Colqui and Casapalca. 4 - Irregular bodies, skarns, veins and disseminations related to the Cordillera Blanca batholith. This group includes the polymetallic skarns of Magistral, Antamina and Contonga, as well as the polymetallic veins with silver and tungsten of Pusajirca.

The Miocene belt of polymetallic ore deposits in Bolivia is situated W of the Sn-Ag province, and represents a southern extension of the Peruvian Miocene belt. Its geological context (Miocene sub-volcanic intrusives hosted by Paleozoic clastic rocks) is similar to that of the tin belt. The larger deposits include Laurani, San Andreas, Berenguela, Carangas, Negrillos and Garcí Mendoza. Laurani, the principal deposit, is a zoned deposit, associated with an andesitic-dacitic complex, cross-cut by a rhyolitic stock and by dykes, directly related to the mineralization (Ahlfeld, 1967; Routhier, 1980).

An extension of the Miocene polymetallic belt to the S is represented by Pb-Zn-Ag (Cu, Bi) veins in northwestern Argentina (Salta and Jujuy provinces). The main districts, Pan de Azúcar (22°43'S / 66°06'W), La Esperanza (24°14'S / 66°34'W) and La Concordia (24°10'S/66°24'W), are associated with dacite domes, and the ore mineralogy includes galena, sphalerite, chalcopyrite, pyrite, and tetrahedrite. The main deposits of this group have before mining reserves up to 0.26 Mt ore, containing 5 - 11% Pb, 1 - 6% Zn and 200 - 500 g/t Ag (Sureda *et al.*, 1986).

In the Patagonian Cordillera of Argentina and Chile, between 46° and 52°S, there are numerous polymetallic deposits hosted by Paleozoic, Mesozoic and Cenozoic rocks

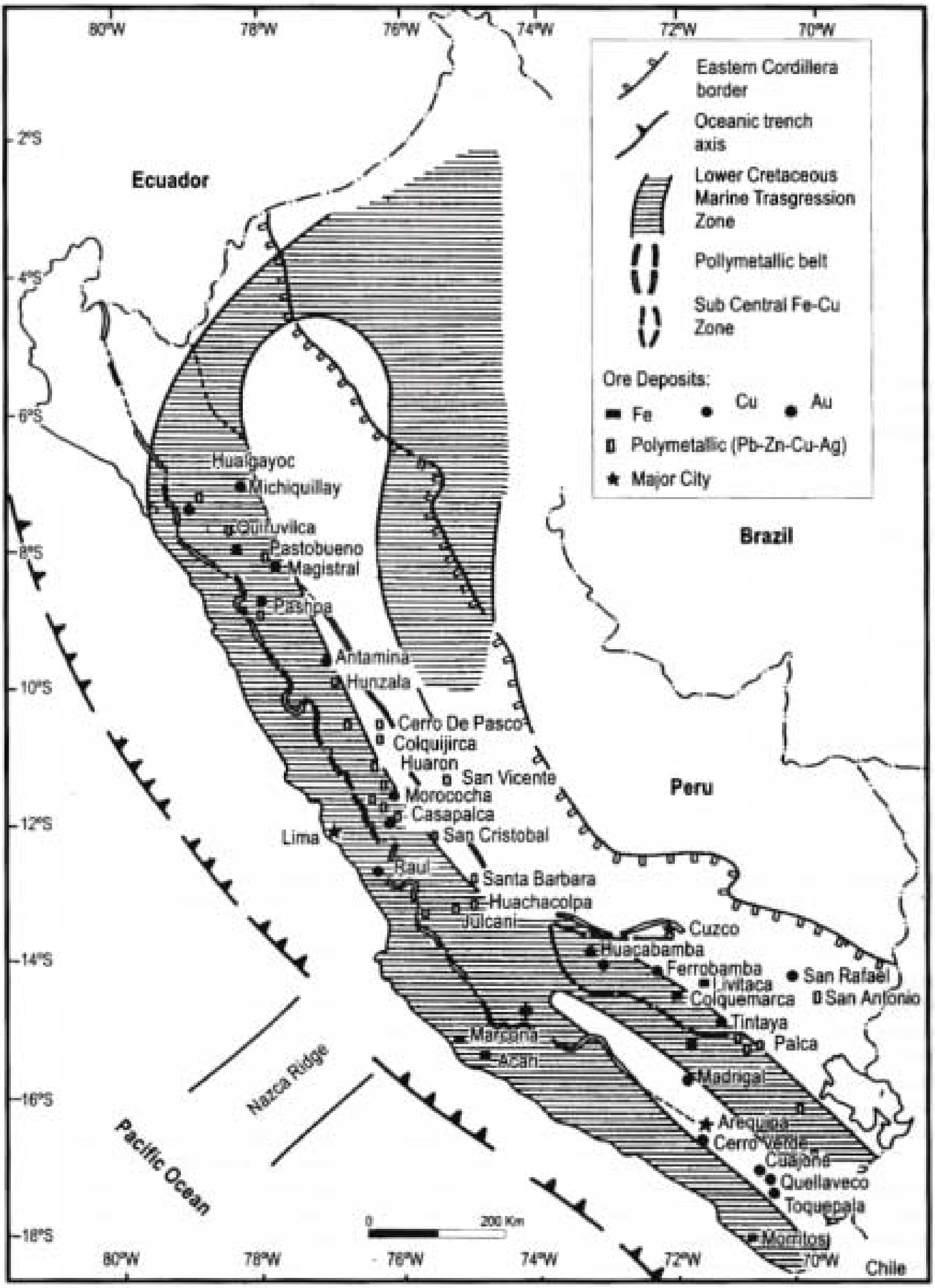


FIGURE 7 - The distribution of major metallic ore deposits in Peru and some principal geological features (modified after Bellido et al., 1972; Corvalán, 1981; Zeil, 1979).

of different types. Márquez (1988) describes a general zoning pattern, with Mo, W in or around central granitic intrusive rocks and Pb-Zn, Cu, Au and Ag at the periphery. According to this author, the granitic rocks responsible for the mineralization are Miocene in age.

At least in the case of the Chilean polymetallic deposits of the Patagonian Cordillera, it is possible that they belong to different ages of mineralization although these ages remain uncertain. Thus, Pb-Zn-Ag-(Cu) deposits occur between 46°00'S and 47°20'S, hosted by Paleozoic metamorphic rocks (phyllite and marble of marine origin) intruded by post-Paleozoic granitoid plutons (Ruiz and Peebles, 1988; Schneider and Toloza, 1990). The main deposit, Mina Silva (46°33'S / 72°24'W) consists of high grade Pb-Zn (Ag) ores, with minor copper content, that form lenticular bodies hosted by metamorphic limestone. Although Ruiz and Peebles (1988) interpreted the deposit as singenetic mineralization of Paleozoic age. Schneider and Toloza (1990) argue that all ore deposits of the district (which also include stratabound and non-stratabound deposits in Jurassic rocks) are related to calc-alkaline magmatism developed in a Mesozoic back-arc setting.

The other important district in this belt is El Toqui, situated at 45°00'S / 71°58'W, and described by Wellmer *et al.* (1983) and Wellmer and Reeve (1990). The district, which covers some 25 km², contains several Early Cretaceous deposits consisting of siliceous volcanic rocks and clastic and carbonate marine sediments, intruded by quartz-bearing porphyries. The basal volcanic unit is cross-cut by Zn-Pb-Ag veins, and is overlain by andesitic-rhyolitic flows and clastic-carbonate sediments, that host the statiform sulphide ore bodies. They occur in the interfingering zone of carbonate rocks with black shale or pyroclastic zones, and contain Zn-Pb-Cu or just Zn as principal economic metals, while Ag is recovered as a sub-product. The largest ore body, San Antonio, overlies a quartz-bearing porphyric sill, partially altered and mineralized. Some cross-cutting mineralizing feeders, and basal hydrothermal alteration and mineralization, have been recognized in the district. Wellmer and Reeve (1990) interpreted the genesis of the deposits of the El Toqui District in terms of massive sulphide mineralization. This was related to the submarine volcanic environment of an aborted back-arc system, at a time close to the Jurassic-Cretaceous boundary.

The tin province

Of the different Andean metallic provinces, the tin belt presents the highest degree of definition and specification. Thus, all the major deposits are in Bolivia, along a NW to NS belt, up to 500 km long, extending from the western border of the continent. So far, no tin ores have been found in Chile. Besides which, the tin province is situated in the central part of the Andean Belt, where the present continental crust attains its maximum thickness (Figs. 8, 9 and 11).

Although the principal deposits of the tin province have a Tertiary or Lower Mesozoic age and are situated in the Cordillera Real of Bolivia. Tin deposits of Paleozoic age are known in Argentina. It is also possible that some minor tin deposits in the Caraballa Cordillera of Peru, close to the Bolivian border, be related to Permian granitoid plutons (Clark *et al.*, 1983).

The Argentinean Paleozoic tin deposits occur in two areas of the Pampean Ranges (Fig. 12). Those of the northern area are vein or greisen-type; their age is Cambrian to Silurian, and their ores include cassiterite, wolframite and sulphide minerals. The deposits of the southern area are pegmatitic, and have a Cambrian to Ordovician age (Malvicine, 1975). Their interest is more scientific than strictly economic.

The tin belt of Bolivia (Turneaure, 1971), may be divided into two segments. Northwards from 18 °S, the belt trends NW and most of the deposits have a Late Triassic-Early Jurassic age. By comparison, in the southern segment, as well as in the central part of the belt, Miocene Sn-Ag deposits are dominant. Whereas the ore deposits of Early Mesozoic age are related to granitic rocks, those of Miocene age are associated with acid subvolcanic bodies. The strong hydrothermal alteration associated with both types of deposit obscures the original composition of the mineralizing igneous rocks. However, their high potassic, peraluminous character, is recognized, as well as the likely participation of crustal material in the generation of their magmas. This participation is coherent with the larger distance of the tin province to the possible position of the paleo-subduction zones (during the Triassic-Jurassic and the Miocene, respectively). It is interesting to note that the southern part of the tin belt coincides with a rich Sb sub-province (Bolivia was the world's third Sb producer and has some 200 deposits of this metal, Routhier, 1980). To the E of the tin province, there are several polymetallic deposits (mainly rich in Ag). This fact raises the question as to whether the southern part of the tin province is superimposed on the larger polymetallic one, or whether it is situated to the E.

The host rocks for both the igneous bodies and the tin deposits of the belt are Paleozoic clastic metasedimentary rocks, the product of detrital sedimentation that began as early as the Cambrian, in a shallow but persistent intercratonic marine basin (Zeil, 1979). This continued till the Middle Devonian, when conditions changed from marine to continental, but the subsidence of the basin, and the sedimentation, persisted up to the Mesozoic. The outcrops of this monotonous series of shale and sandstone beds, some 10 000 to 20 000 m thick, make up a major part of the present Cordillera Real, where most of the Bolivian tin deposits of all types and ages are found.

Two types of tin deposits of Late Triassic-Early Jurassic age are known. The most abundant corresponding to Sn-W veins associated to greisen-type alteration, within small batholiths (*e.g.*, Yani, Sorata) or in the contact metamorphic zones imprinted by the batholiths in the Paleozoic sedimentary host rocks. The age of the emplacement of the batholiths is in the 257 to 150 Ma range (Grant *et al.*, 1980). Among the principal districts are those of Sayaquira, Caracoles and Araca. None of these attain the magnitude of the Tertiary Sn-Ag deposits.

The other type of Upper Triassic-Lower Jurassic tin deposit, which is found along a NW belt, N of 19 °S, shows stratabound control of the ore. Although this type of tin deposit is not economical under present tin price conditions, its origin (syngenetic or epigenetic) presents an interesting problem (Schneider and Lehmann, 1977). As stated by

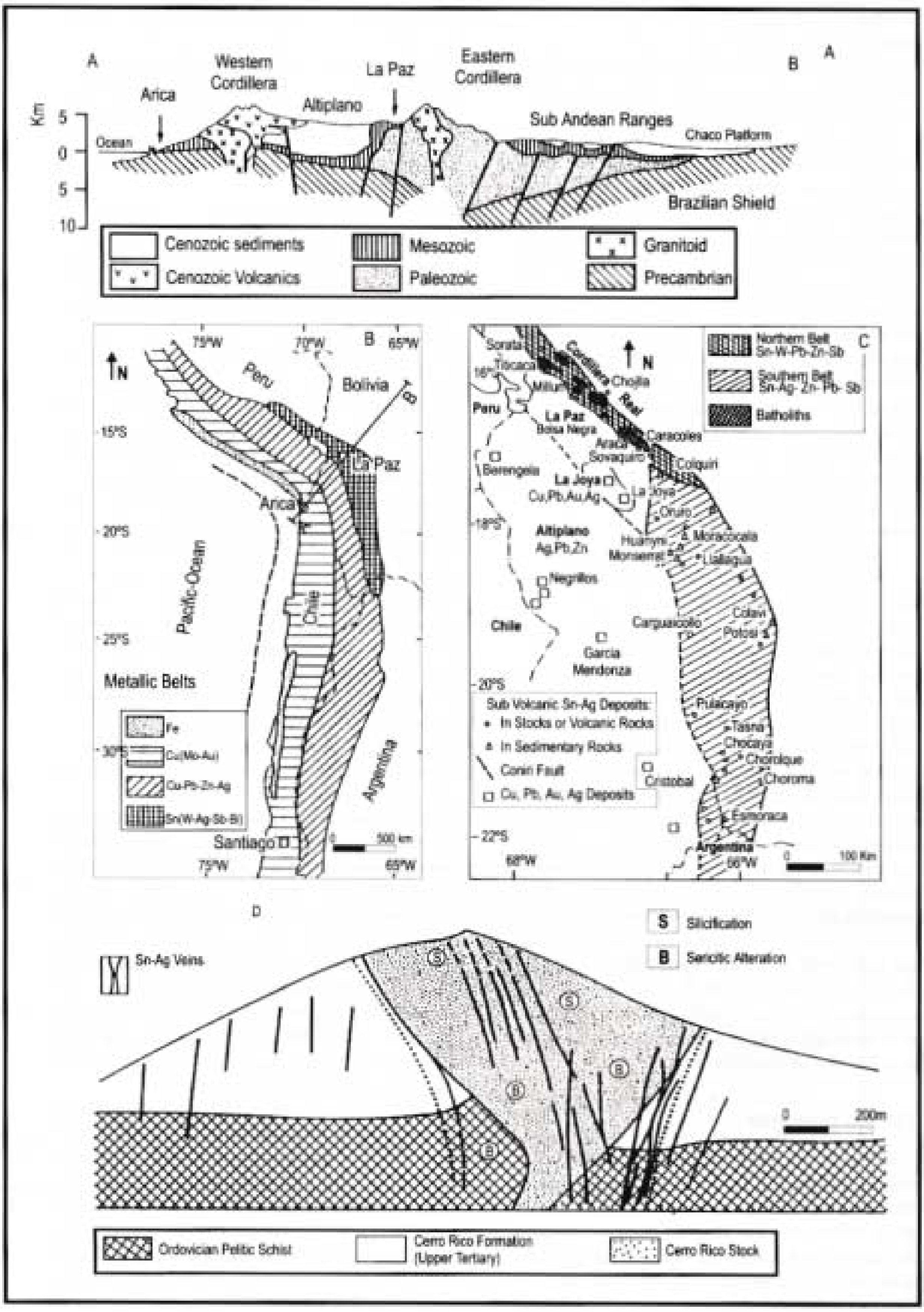


FIGURE 8 - The tin belt of Bolivia and the Cerro Rico tin-silver deposit (modified after James, 1971, A; Sillitoe, 1972, 1976, B; Turneaure, 1960, C; Turneaure, 1971, D).

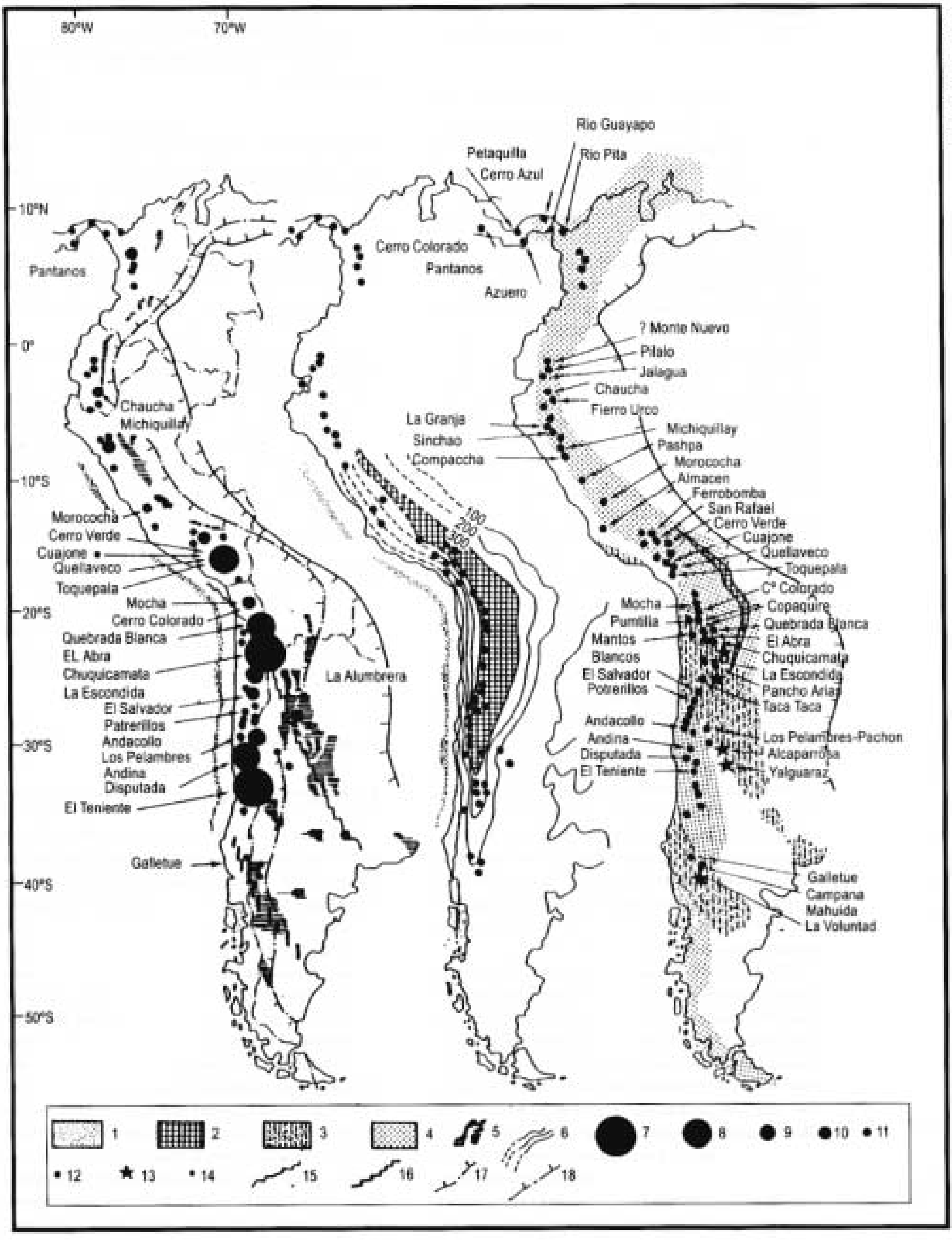


FIGURE 9 - Porphyry copper deposits and major geological features of the Andean Belt. 1- Ocean trench. 2- Thick continental crust. 3- Paleozoic tecto-magmatic belts. 4- Andean magmatism. 5- Precambrian massifs. 6- Gravimetric contours. 7 to 12- Order of magnitude of ore deposits, expressed in metallic Cu content; 7: 50 Mt, 8: 20 Mt, 9: 5-10 Mt, 10: 2.5-5 Mt, 11: 1-2.5 Mt, 12: < 1 Mt 13- Paleozoic ore deposits. 14- Mesozoic and Cenozoic ore deposits. 15- Pampean Massif. 16- Patagonian Massif. 17- Limits of the Andean geosynclinal basin. 18- Limits of the Paleozoic geosynclinal basin (modified after Oyarzún and Frutos, 1980).

Lehmann (1985, 1990), the host rocks for the stratabound tin deposits are Lower Paleozoic metasedimentary rocks, which are intruded by granite and granodiorite.

Kellhuani, one of the three principal stratabound-type tin deposits (Lehmann, 1985; 1990) is situated some 15 km north of La Paz. About 10 Mt ore, containing 0.5% Sn, are distributed in the quartzitic units of the Catavi Formation (Silurian), whereas no ores are present in the interlayered black shale strata. The ore has been mined from several quartzitic zones (mantos) in a district 16 x 4 km, trending NW-SE. Although the stratabound character of the mineralization (that is both concordant and discordant) favours a syngenetic sedimentary origin, studies by Lehmann (1985) presented important evidence in favour of the epigenetic-hydrothermal deposition of the ores. The ore contain about 25 mineral species, several of which are sulphide and sulphosalts, in addition to cassiterite, the economic mineral. The principal evidence is the zoned character of the geochemical halos and the hydrothermal alteration of the district around the Chacaltaya granite porphyry dated at 213 ± 5 Ma, a distribution in which cassiterite occupies a distal position. Lehmann (1990) considered the fact that sericite alteration, associated with cassiterite ore, produced a similar age to that of the granitic intrusion, as a further confirmation for the epigenetic origin of this Upper Triassic district.

The Tertiary tin deposits (Sillitoe *et al.*, 1975; Grant *et al.*, 1976, 1980; Francis *et al.*, 1981) are related to sub-volcanic intrusive bodies, partly brecciated, at a high emplacement level, that cross-cuts the Paleozoic clastic formations. Grant *et al.* (1979), distinguished two chronological groups. The first is formed by 26 to 20 Ma old intrusive rocks that crops out between latitudes 16°30'S and 19°50'S, and are associated with several important deposits, such as Catavi and Llallagua. However, the richest Sn-Ag deposits (Cerro Rico, Chorolque; Fig. 8), are related to a second, younger (17-12 Ma) group of sub-volcanic bodies that crop out between latitude 19 °S and northern Argentina. The association of these acid intrusions to ignimbritic deposits occurs frequently. Thus, the Potosí district is associated with a large ignimbritic source in the Karikari resurgent caldera. Grant *et al.* (1980) distinguished two types of tin deposits in this belt, that they denominated porphyric and non-porphyric.

The first group includes such important deposits as Llallagua, Cerro Rico and Chorolque. Taken together their principal economic mineralization is of the vein-type. They also contain, as a whole, some 80 Mt of disseminated ore grading 0,3% Sn, which is still far from being of economic interest, but represents an important reserve for the future. Five principal geological-mineralogical features are common to the deposits of this group: 1 - The mineralization is centred on small (1 - 2 km^2) porphyric stocks, emplaced under or within volcanic pipes. 2 - Several intrusive pulses and breccification are observed. Some stocks are converted to breccia pipes. 3 - The stocks and their host rocks have suffered intense and penetrative feldspar-destructive hydrothermal alteration, in which sericite and tourmaline predominate. 4 - The mineralization is very complex. The main sulphides that accompany the cassiterite are pyrite, stannite, chalcopyrite, sphalerite, and arsenopyrite. 5 - The disseminated mineralization is earlier than the high-grade vein-type. The radiometric dating by Sr isotopy have yielded Miocene ages of 20 Ma at Llallagua, 15 - 14 Ma at Cerro Rico, and 17 - 12 Ma at Chorolque (Grant *et al.*, 1980).

The magmas related to tin mineralization usually have a much differentiated petrological evolution (Lehmann, 1990). Although some magmas related to the Bolivian tin porphyries are evolved, such as at Karikari and Potosí, where peraluminous, high initial Sr isotope ratios (0.707 - 0.716) magmas, evolved from andesite to toscanite (Grant *et al.*, 1980). In general, tin porphyries are associated with only moderately fractionated subvolcanic rocks of rhyodacitic composition. However, the recent paper by Dietrich *et al.* (1999) provided analytical evidence (melt inclusions data) for the origin of the Bolivian tin porphyry magmas by mixing of highly evolved siliceous melts, containing quartz phenocryts with andesitic to basaltic melt fractions, in an upper crustal reservoir.

In the group of non-porphyric deposits are included vein-type Sn mineralization, hosted in Paleozoic clastic sediments that are not related to outcropping intrusive bodies (except dykes). Among them are the Colquiri (fluorite-sphalerite-cassiterite); Huanuni, Santa Fe, and Morocala (cassiterite) and Tasna (cassiterite, with Bi and Cu in the sulphide phase) deposits (Grant *et al.*, 1980)

Tin-silver veins in north-western Argentina (Sureda *et al.*, 1986) represent the southwestern extension of the Bolivian tin belt. The major deposit, Pirquitas (22°44'S / 66°27'W) is hosted by strongly folded, clastic Paleozoic rocks. Its paragenesis includes high T (pyrrhotite, cassiterite, arsenopyrite) and low T (sphalerite, galena, sulphosalts) phases, both crystallized at shallow sub-volcanic levels. The average grade of the deposit is 1.1% Sn and 500 g/t Ag.

Andean Metallogenesis

Andean magmas and ore deposits

Magmatic rocks are dominant in the Andean Belt and most ore deposits are directly or indirectly associated to magmatic activity. A major part of the extrusive and intrusive rocks of Paleozoic to Cenozoic age belong to the calc-alkaline series, although tholeiitic rocks are present in the accreted oceanic prisms of the northern Andes, and both shoshonitic and alkaline rocks are associated to the calc-alkaline series. Except for the tholeiitic rocks, the chemical and isotope composition of Andean igneous rocks suggest that their magmas originated from common though variable sources and mechanisms. This point is illustrated, by the strong similarities in chemical and isotope composition of rocks from such different setting and ages as the Paleozoic granitoids of the Cordillera Frontal in Argentina ($^{87}Sr/^{86}Sr$ (i) = 0.7053 - 0.7070; Caminos *et al.*, 1979) and the Plio-Quaternary andesites of the Central Andes ($^{87}Sr/^{86}Sr$ (i) = 0.7051 - 0.7077; Pichler and Zeil, 1972). The general model (López-Escobar *et al.*, 1977, 1979, 1995; Thorpe and Francis, 1979) considers that the Andean magmas originated in the Upper Mantle zone between the subducted oceanic plate and the continental crust. The model also considers the participation of melts and fluids from the upper layers of the subducted plate, as a trigger mechanism for partial melting in the mantle, a contribution that has been sustained by Be-10 isotopy (Morris *et al.*, 1985). The final composition of

Andean magmas are then explained in term of different contribution from the oceanic plate, variable degrees of partial melting of mantle materials, different fractional crystallization processes during the rise of magmas and possible contamination during their passage through the continental crust. An alternative source proposed for Andean magmas generated in zones with a thick continental crust, are the lower crustal levels (*e.g.*, Pichler and Zeil, 1972; Mc Kee *et al.*, 1994). The participation of mantle melts interacting with crust derived melts in deep reservoir, has also been considered and supported by Sr isotopy (*e.g.*, Deruelle and Moorbath, 1993, for lava from the south-central Andes).

The incorporation of crustal igneous and sedimentary materials to the magmas during its passage through the crust is well established as a mechanism for emplacement of the Coastal Batholith of Peru (described in the important book by Pitcher *et al.*, eds., 1985, and considered as a model for batholith emplacement in the Andes). Although this process involves the continuous (since 102 Ma to 60 Ma) canibalistic assimilation of stratified rocks, the fact that they were mainly volcanics, with similar chemical and isotope composition to the batholith's magma, implies that no detectable compositional change occurred.

However, it is possible that crustal materials contribute to the magma enrichment in LIL-type (*e.g.*, K, Rb, Ba) and incompatible (*e.g.*, Cu, Mo, Pb) elements, by partial assimilation of crustal materials. Thus, normal high-K and shoshonitic, intermediate to mafic, Mesozoic volcanic rocks in central-northern Chile, differ only in their K, Rb and Ba content, the non LIL-elements remaining almost constant (Oyarzún *et al.*, 1993).

In consequence, there are several possible sources of the metals and metaloids to the Andean ore deposits related to magmatic processes, and the isotope data are relevant to assess their relative importance. Two elements are most relevant in terms of their isotope ratios to evaluate possible ore sources. They are the Pb isotope ratios for the metals and the S isotope ratios for the metaloids. However, Pb has a strong tendency to accumulate in the crust and the interpretation of their isotope ratios in term of source for the ores does not necessarily apply to other metals such as Cu, Zn or Mo. Besides which, where the country rocks are volcanic or volcaniclastic with a similar age and composition to that of the intrusive rocks, Pb isotope ratios cannot be used to discriminate between the metal provided by the magma from the metal scavenged from the country rocks by hydrothermal or metamorphic fluids. This situation is fairly common for metallic ore deposits in Mesozoic and Cenozoic rocks of the Andean Belt.

There are numerous studies on Pb isotope ratios in Andean igneous rocks and ore deposits. In general, they conclude that different sources participate to variable degrees according to the tectonic settings of the rocks and the ore deposits. Thus, Puig (1988, 1990) drew attention to the relatively narrow range of Pb isotope ratios in Andean ore deposits, interpreted by this author in terms of reservoir mixing processes during the Andean evolution. However, he also established some relationships between the Pb isotope ratios and the tectonic setting of the deposits. Thus, polymetallic ores in volcano-sedimentary rocks of the tectonically extensional Lower Cretaceous basin in Chile, are less radiogenic than those found in similar Jurassic rocks. These results are consistent with the conclusion of Fontboté *et al.* (1990) for stratabound deposits in the Andes: those related to mafic or intermediate rocks have Pb isotope ratios pointing to a mantle source, while those deposits related to felsic igneous rocks or to sediments present isotope ratios according to an orogenic (recycled lower and upper crust) or to an upper crustal source (San Vicente). They also noted that the Pb isotopes of the ores are more radiogenic in those deposits situated in the E. Petersen *et al.* (1993), enlarging on the previous study by Macfarlane *et al.* (1990), proposed four Pb isotope provinces for the Central Andes, from W to E: Coastal region of Peru and northern Chile; High Andes (Peru, Chile, Bolivia, Argentina); Eastern Andes (Peru, Bolivia, Argentina); and Eastern Andean Foothills. A deep source is suggested for the first two provinces; a shale-bed source for the Eastern Andes ore deposits; and a cratonic source for those deposits in the Eastern Andean Foothills.

Regarding ^{32}S / ^{34}S isotopy, the different studies are coincident in terms of the magmatic origin of sulphur in most of the sulphide metallic deposits of the Andean Belt. In the case of porphyry copper systems, $\delta^{34}S$ in sulphide minerals is very close to the meteoritic standard (*e.g.*, - 3 ‰ at El Salvador; Field and Gustafson, 1976; - 1.4 ‰ at Chuquicamata, - 2.1‰ at Río Blanco, and - 3.1 ‰ at El Teniente; Sasaki *et al.*, 1984). This is also the case for sulphides in magnetite ore deposits (*e.g.*, minor pyrite at El Laco; Vivallo *et al.*, 1994). Concerning the stratabound sulphide deposits, those emplaced in volcanic, volcaniclastic or coarse detrital sedimentary rocks have $\delta^{34}S$ close to the meteoritic standard. On the other hand, those deposits hosted in sedimentary rocks, including black shale generally have $\delta^{34}S$ in the - 10 to - 40 ‰, suggesting the effect of bacterial activity over sulphate ions of magmatic origin (Spiro and Puig, 1988). An important exception is the San Vicente Mississippi Valley type Zn-Pb deposit in Peru, that presents positive and homogeneous $\delta^{34}S$ values between + 6.9 ‰ and + 13 ‰, which are interpreted in terms of bacterial reduction of ^{34}S-enriched sedimentary sulfate (Gorzawski *et al.*, 1990).

Though the close relationship between magmas and Andean ore deposits is well established, many aspects of this association remain poorly understood or are just beginning to clarify. In the following paragraphs, some aspects of the problem will be briefly considered.

Porphyry copper deposits are the best-studied deposits in the Andean Belt and possibly in the world. They have low ^{87}Sr / ^{86}Sr (i) ratios, very low $\delta^{34}S$ indices, and, at least in the case of those of the Eocene-Oligocene age in northern Chile, have Pb isotope ratios that are much narrower than those of all other types of ore deposit or intrusive or volcanic rocks of all ages in the Central Volcanic Zone (Zentilli *et al.*, 1988). Generally accepted models (*e.g.*, Sillitoe, 1973) situate their porphyric intrusives over the cupola of calc-alkaline batholites. However, as pointed by Maksaev and Zentilli (1988), the Eocene-Oligocene porphyries formed during the last important period of magmatic activity recorded in the Domeyko Range, before a 50 to 150 km eastward shift of the magmatic belt, display a somewhat anomalous trait of calc-alkaline magmatism. The porphyry magmas had to cross a thickened crust, a consequence of the late Eocene Incaica Compresive Stage (a condition in common with the Paleocene porphyries of south-western Peru and with the Pliocene porphyries of central Chile). However, they exhibit the lowest possible degree of crustal contamination. Therefore, a rapid, diapiric-style of ascent through the crust of deeply generated magmas has been proposed by Maksaev and Zentilli (1988) for the porphyries. This model is not consistent with the simple relationships of porphyry copper deposits to normal batholiths that the model proposed by Sillitoe (1973) implicates.

Several studies (*e.g.*, Baldwin and Pearce, 1982; López-Escobar and Vergara, 1982) have attempted to find some significant relationship between the chemical composition of slightly altered intrusive rocks associated with porphyry copper deposits and their productivity in terms of porphyric mineralization. However, no significant difference was found with respect to non-productive contemporary intrusive rocks. The only difference was a lower Y and Mn content observed by Baldwin and Pearce (1982) in the productive porphyries of the El Salvador District of northern Chile.

However, the possibility that porphyry copper systems were not related to normal calc-alkaline batholiths, but rather to magnetite-rich, mafic bodies of batholithic magnitude, was suggested recently by Behn and Camus (1997). These authors considered the presence of large ENE and NWN magnetic anomalies that exhibit spatial coincidence with Eocene-Oligocene porphyry copper deposits between 18 °S and 27 °S, in terms of mafic magmatic reservoirs from which porphyry copper systems were possibly derived.

Although calc-alkaline magmatism has been assumed as the source for porphyry copper systems, it is well known that the principal mineralization is closely associated to potassium metasomatism. Skewes and Arévalo (1997) have proposed a daring alternative interpretation for the case of El Teniente, where the Cu (Mo) ore is in K-rich biotitic andesites, that hosts quartz dioritic and dacitic porphyties. Instead of the traditional interpretation (that is, the andesite was hydothermally altered by the porphyries), they proposed that the andesite represent an ore-rich, high-K, intrusive magma. Considering the chemical analysis published by Camus (1975), these andesites, if interpreted as primary rocks, should be classified as absarokite (shoshonitic basalt) according to the Peccerillo and Taylor (1976) diagram. It is an interesting fact that the presence of high-K or shoshonitic magmas has been established at the Farallón Negro Complex (Sasso and Clark, 1988), related to porphyric Cu (Au) mineralization.

Besides which, the model proposed by Skewes and Arévalo (1997) resembles the ore-magma concept, which has been applied in Chile, with a variable degree of acceptance, to explain the origin of Kiruna-type iron deposits since 1931. Although objections to this theory (*e.g.*, Nynstrom and Henríquez, 1994) have been raised on the on grounds of the mineralogical and physico-chemical data, the concept is now making a comeback.

The fact that the Tertiary igneous rocks related to Sn-Ag mineralization in southern Bolivia have a peraluminous character and high-Sr (i) isotope ratios (0.707 - 0.716), suggest a significant participation of the continental crust in their petrogenesis (Schneider, 1987). However, the recent paper by Dietrich *et al.* (1999) presented analytical data that also supported the participation of andesitic to basaltic melts (mixed with highly evolved rhyolitic melts in upper crustal reservoirs) in the genesis of Tertiary Bolivian tin porphyries. Therefore, mafic magmas may have played a more important part in the genesis of Andean ore deposits than has yet been recognized. Also, the mechanism of magma mixing proposed by Drietich *et al.* (1999) may be useful to explain the genesis of other types of Andean deposits, such as the Kiruna-type Cretaceous iron deposits of northern Chile, where there is evidence for the involvement of both mafic and alkaline, F, Cl rich magmas.

Finally, although most of the Andean ore deposits are associated with magmatic activity, which has been almost permanent in the belt, metallogenesis seems to have been somewhat discontinuous, and related to significant tectonic activity that abruptly displaced the magmatic belts. Therefore, favourable conditions for the mixing of different types of magma may have occurred during these disruptive episodes that will be discussed in the next section.

Andean tectonics and ore deposits

Although magmatic activity provides the direct source and mechanisms for the generation of ore deposits in the Andean Belt, tectonics control not only the production and emplacement of magma, but also the channels for the ore bearing fluids. Besides which, although the association between plutonic and coeval volcanic rocks is a normal condition of the Andean magmatism, the ratio of the volume of intrusive and extrusive magmas has been very variable, the volcanism being favoured during the stretching stages and the plutonism increasing with the compressive tectonic pulses.

Both the geological and the metallogenetic evolution of the Andean Belt during the Mesozoic-Cenozoic span, can be consistently explained in terms of the interactions of the continental and oceanic lithosphere plates. Among the main consequences of this interaction is the continuous production of calc-alkaline magma; the accretion to the continent of oceanic prisms; the development of back-arc basins; the occurrence of several orogenic episodes; the formation of mega-fault zones and the generation of ore deposits.

Post-Paleozoic accretion of oceanic prisms occurred during Tertiary times in the northern Andes (Colombia and Ecuador), when a Mesozoic mafic igneous-marine sedimentary complex was incorporated to the western border of the continent. Except for some peridotitic podiform Cr ores and for some massive sulphide bodies (Ortiz, 1990) this episode had little direct metallogenetic importance.

Two subduction styles have been recognized for the tectonic evolution of the central and south-central Andes (Uyeda and Kanamori, 1979): a low-stress Mariana-type, for the Jurassic-Early Cretaceous span, and the compressive Chilean-type of subduction operative since the Late Cretaceous. The passage between both regimes (related to the westward shift of South America after the break-up of Gondwana), which implies a shallower angle for the subducted slab, occurred in the Chilean segment between 108 and 100 Ma (Sillitoe, 1991), in close coincidence with the ages of a number of deposits in the Neocomian back-arc basin domain. This basin, which reached an aborted marginal basin stage in Chile (Levi and Aguirre, 1981) and a straight marginal character in Peru (Atherton and Webb, 1989), attained maximum subsidence in the Albian, receiving several thousands metres of mafic lava and marine sediments. About 110 Ma ago, numerous Kiruna-type Fe deposits and stratabound and skarn-type Cu deposits (several of them rich in magnetite), were formed in volcanic or sedimentary rocks of the basin. Among them are the Fe-Cu skarns of Eliana (112 Ma), Monterrosas (110 Ma) and Hierro Acari (109 Ma) in Peru (Petersen and Vidal, 1996); the Kiruna-type Fe deposits of the coastal belt in Chile: Los Colorados (110 Ma), El Algarrobo (128 - 111 Ma; Montecinos, 1983); and several stratabound Cu deposits in central Chile, such as El Soldado (110 Ma) and Lo Aguirre (113 Ma; Munizaga *et al.*, 1988). Shortly afterwards (112 - 105 Ma) the Andacollo porphyry copper was also emplaced (Sillitoe, 1988). This coincidence is amazing, considering that extensional conditions still prevailed in the Huarmey Basin of Peru at that time.

As pointed out by Sillitoe (1988, 1991), the eastward shift of magmatism in the Chilean-Argentinian Andes from the Jurassic to Miocene times, has produced several belts of ore deposits trending N-S, coincident with the position of the contemporaneous magmatic belt. They include porphyry copper deposits from the Albian. Although the eastward shift has been interpreted in terms of a lower angle of the subducted slab, due to acceleration of the convergence rate of the tectonic plates, the mechanism is not completely understood. Thus, as stated by Sasso and Clark (1988) for the middle Miocene stage, the arc therefore did not merely shift eastwards (Davidson and Mpodozis, 1991) but, within the limits of error of the $^{40}Ar/^{39}Ar$ dating technique, instantaneously broadened in the middle Miocene. Other example of sudden horizontal eastward magmatic and metallogenetic displacement, is that of the Andahuaylas-Yauri Cu-Fe skarn belt, linked by Noble *et al.* (1984) to a change in the subduction geometry due to the Incaica Orogeny.

As explained by Scheuber and Reutter (1992), the stress component normal to the plate boundary produces structures of crustal shortening or extension, whereas the component parallel to the plate boundary (in case of oblique convergence) causes longitudinal wrenching. Two important fault zones in the Andes of northern Chile are interpreted in terms of oblique subduction. These are the Atacama and the Domeyko fault zones, to which many high tonnage ore deposits are associated (Fig. 10). The Atacama Fault Zone (AFZ) represent an ancient zone of crustal weakness that was reactivated in the Early Cretaceous, as a consequence of a N20°E plate convergence, the oceanic Aluk Plate coming from the NNW (Pardo-Casas and Molnar, 1987). The oblique plate convergence generated regional shearing traduced in dominant sinistral strike-slip movements, up to several tens of kilometres (Bonson *et al.*, 1997). During the Early Cretaceous, magmas and their derivative fluids, responsible for Kiruna-type Fe and Cu-Fe deposits such as Manto Verde, were focused into dilation sites and fault intersections at the AFZ (Thiele and Pincheira, 1987; Bonson *et al.*, 1997).

The Domeyko Fault Zone is also interpreted in terms of an oblique convergence, this time the oceanic plate (Farallón) coming from the SW with a convergence rate of 12 cm/year. This fault zone is also considered as an early structure, along which there occurred deep readjustment of the crust (Perry, 1953, in Maksaev and Zentilli, 1988). However, this time the process followed an important compressive pulse (the Inca Compression). During a short span (10 Ma) in the Eocene-Oligocene transition, several of the most important porphyry copper deposits of the Andes (Fig. 10) were emplaced along the fault zone, which had a dominant dextral displacement.

An important wrench fault in Peru is the Huara Fault System (Petersen and Vidal, 1996) that has a N to NE strike, and occurs in the brittle environment of the Coastal Batholith, along a Lima-Cerro de Pasco course. Several volcanogenic massive sulphide deposits as well as important polymetallic districts (*e.g.*, Casapalca, San Cristobal, Colqui) may be related to this fault zone (Petersen and Vidal, 1996).

As pointed out by Maksaev and Zentilli (1988), mega fault zones have complex relationships with both magmas and ore deposits. They probably represent major weakness zones within the crust that have some control on the paths of the rising magmas. However, those magmas also contribute to the weakness of the zone, affecting the rheological properties of the rocks. In consequence, the wrenching process due to the parallel stress component (Scheuber and Reutter, 1992) is enhanced. On the other hand, although most of the stockwork-type porphyry copper deposits of the Andes (*e.g.*, Chaucha in Ecuador; Goosens and Hollister, 1973) are related to important faults, other major deposits, such as those of the Arequipa Lineament (Hollister, 1974) or El Teniente (Camus, 1975), do not present evident structural controls (although their alignment points to deep-seated controls).

Thus, the genesis the major Andean deposits, although controlled by the position of the magmatic arc and favoured by structures like the wrench fault zones, should be related to deep seated disturbances, affecting the geometrical and physico-chemical relationships between the subducted oceanic plate, the asthenosphere and the mantle-crust boundary. This concept, illustrated by the Sasso and Clark (1988) model for the middle Miocene broadening of the magmatic arc and the genesis of porphyry Cu (Au) deposits in Argentina, may explain why the larger Andean deposits were formed during such short pulsative spans in like manner to the Kiruna-type deposits in northern Chile (Oyarzún and Frutos, 1984) and the porphyry copper deposits along the whole Andean Belt (Sillitoe, 1988).

The metallogenetical zoning and evolution of the Andean Belt

Three main subjects will be discussed in this section: the tectonic segmentation of the Andes; the distribution of the different metallic provinces; and the metallogenetic evolution of the belt. As with many key aspects of Andean metallogenesis, the implications of the tectonic segmentations of the Andes in terms of magmatism and ore deposits were first rised by Sillitoe (1974), who proposed 16 tectonic boundaries between 0° (Carnegie Ridge) and 44°S (Chile Ridge). Some of these boundaries, which were proposed on the basis of main structures, seismic and volcanic activity, main morphological units, old terrane outcrops and the intersections with oceanic ridges, are coincident with the longitudinal limits of the metallic belts.

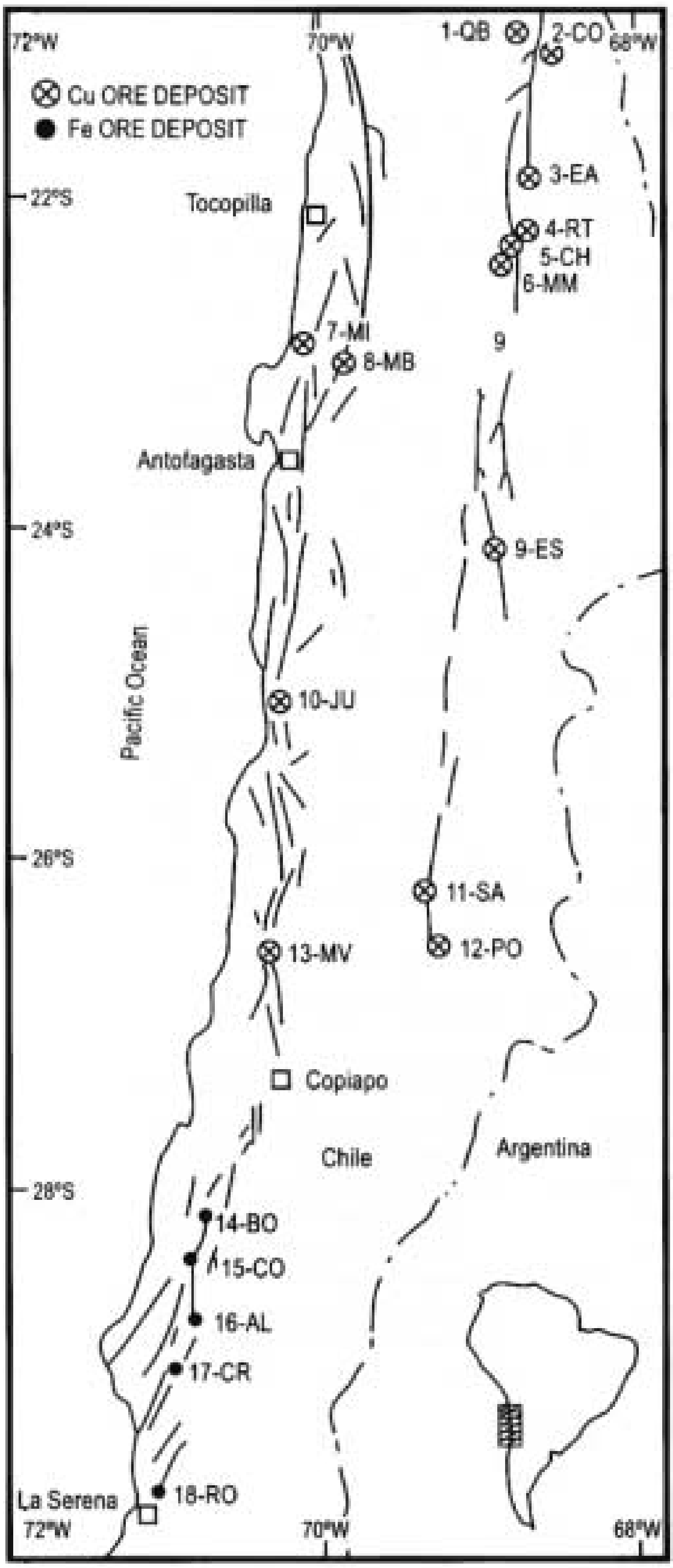


FIGURE 10 - Major iron and copper deposits structurally related to the Atacama and the Domeyko fault zones in Northern Chile (modified after Sillitoe et al., *1996; Vila* et al., *1996).*

Thus, the tin belt is restricted to three segments, enclosed by boundaries 5 (northern limits of the belt of recent central Andes volcanoes and of the Altiplano-Puna Block) and 8 (northern limit of the Domeyko Cordillera and westward step in the longitudinal belt of recent volcanoes).

The Andean tectonic segmentation is the result of a number of heterogeneities along the belt, which is made up of old, and young terranes and tectonic blocks. Among the former is the Precambrian Arequipa Massif, in southwestern Peru (Petford and Atherton, 1995), whereas the Western Cordillera of Colombia consists of a Cretaceous oceanic prism accreted to the continent during Tertiary times. If one considers the heterogeneities of the continental crust, the geometry of the continent, the complexities in the oceanic plates (*e.g.*, the ridges) and the variation in speed and angle of convergence between the plates (and their consequences in the subduction zone), longitudinal segmentation is a natural consequence. However, the relationships between tectonic boundaries and metallic belts are rather uncertain in terms of cause and effect. Thus, the tin province may be, in part, a consequence of the thicker continental crust between boundaries 5 and 8, that could have favoured the magma mixing process proposed by Dietrich *et al.* (1999). On the other hand, the break in the iron belt north of boundary 9 may be interpreted in terms of the higher degree of erosion that affects the Lower Cretaceous rocks, resulting in the unroofing of the batholithic levels. In general, erosion levels have been considered an important factor for explaining the distribution of metallic belts in the Andes (Petersen, 1970; Goossens, 1972b). This factor may be important at regional and local levels, *e.g.*, the deeper erosion levels of the flank of the western Peruvian Andes may favour the exposure of porphyry copper deposits (Petersen, 1970). Also, different erosion levels in the tin belt of Bolivia expose Triassic to Jurassic Sn-W deposits related to deep-seated plutonic rocks in the northern part of the belt, and Tertiary Sn-Ag deposits associated with shallow subvolcanic complexes in the southern segment.

Besides the erosion levels, several other factors have been considered to explain the longitudinal discontinuities of Andean metallic provinces (Oyarzún, 1985, 1990). Thus, Mesozoic paleogeographical conditions in central Peru were favourable to the abundant deposition of carbonate sediments, a factor considering favourable for the development of the rich polymetallic province in this country. On the other hand, this province is less developed in Bolivia, where most sedimentary sequences have a clastic composition, a fact that seems to confirm this hypothesis. However, Mesozoic carbonate sediments in Chile host copper or silver deposits, and Pb-Zn ores are poorly represented (except in the Patagonian Cordillera). In consequence, the presence of carbonate-rich sedimentary rocks appears to be a contributing factor, but not a decisive one.

The presence of metallic domains (Routhier, 1980), defined as volumes of continental crust that are endowed with a special metalliferous potential over geological time, is not a satisfactory explanation for the longitudinal Andean metallic segmentation. In fact, although Paleozoic and post-Paleozoic Andean metallic provinces are similar in nature, their different geographical distribution is not consistent with the concept of metallic domains. Thus,

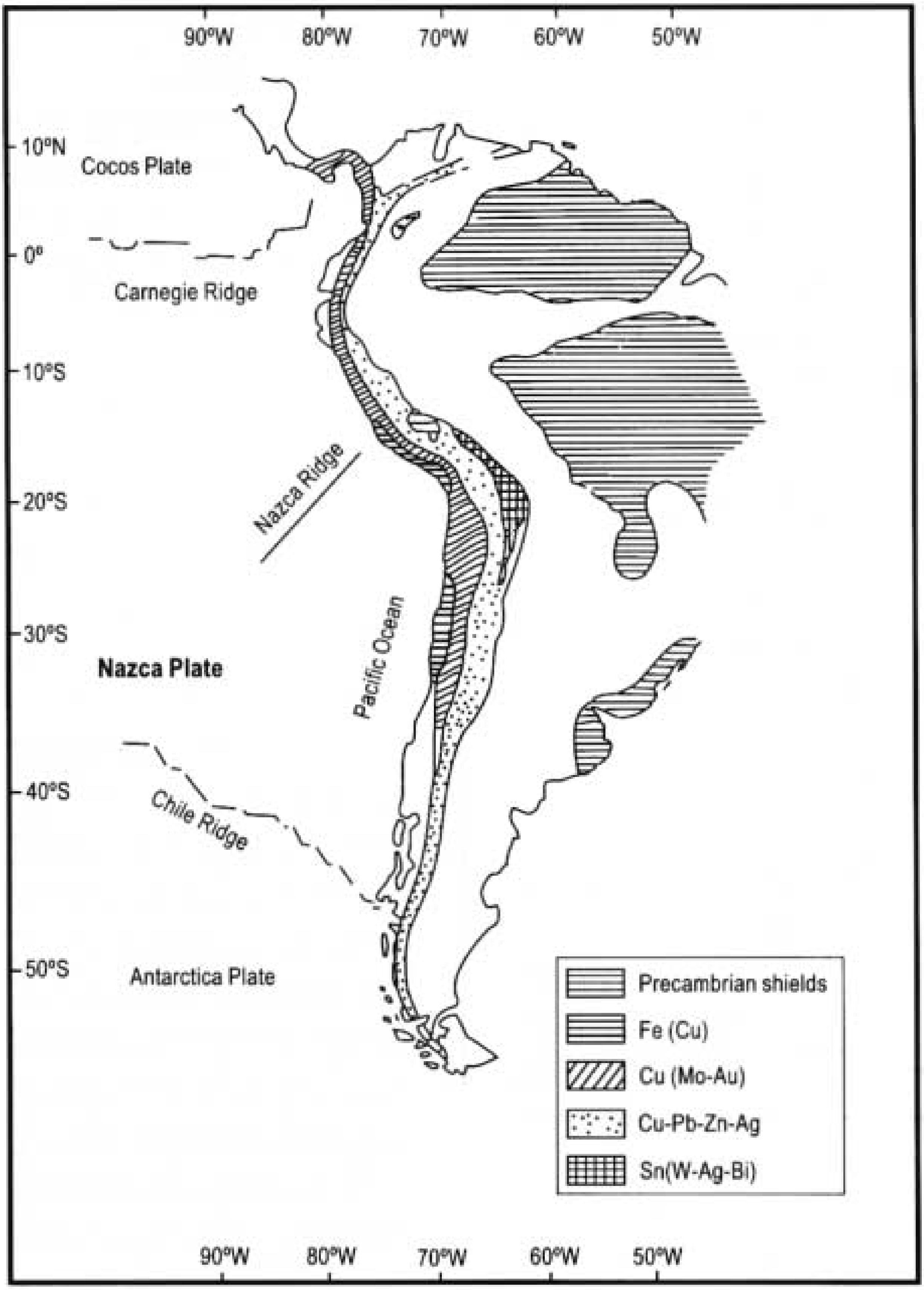


FIGURE 11 -Mesozoic and Cenozoic Andean metallic belts (modified after Ericksen, 1976; Malvicini and Llambias, 1982; Petersen, 1970; Sillitoe, 1976).

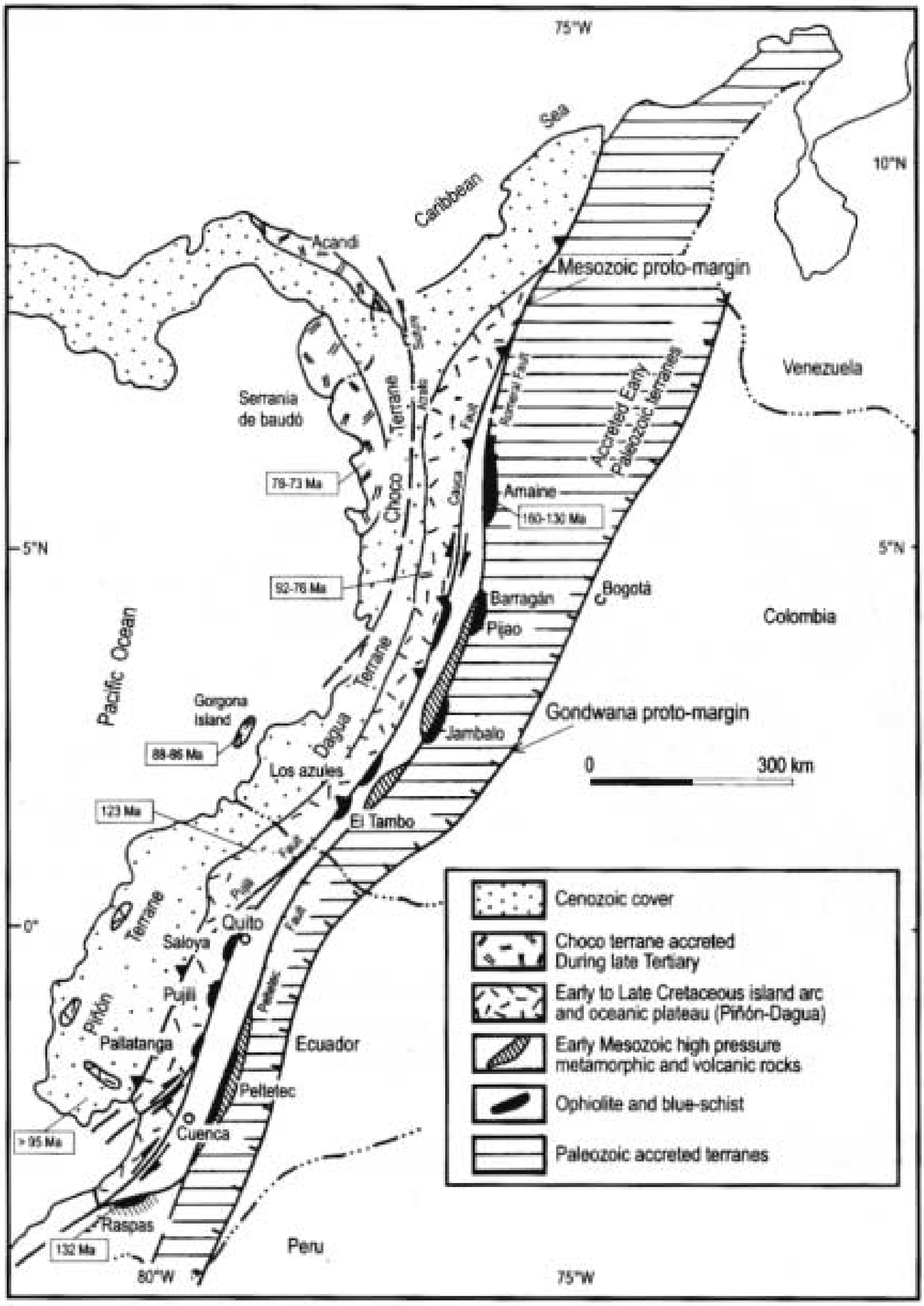


FIGURE 11: Accreted oceanic terranes of the Northern Andes with indication of main ophiolites and high pressure metamorphic rocks exposed along the Cauca Valley of Colombia and the Interandean Depression of Ecuador (modified after McCourt et al., 1984; Millward et al., 1984; Van Thornout et al., 1992). Radiometric ages correspond to the oceanic basement (modified after Reynaud et al., 1999).

(Ishihara and Ulriksen, 1980). Besides which, these relationships are similar to those reported by Ishihara (1977, 1978) for eastern Asia. There, the Sn province in the continental border is associated to ilmenitic granitoids and the island-arc sulfophile province (Cu, Mo, Pb, Zn) to magmatic rocks of the magnetite series. The participation of the subducted oceanic plate in the process is sustained by a precise ratio established in Japan between the convergence speed rate of the plates and the productivity of different arc segment in terms of volcanic sulphur (Ishihara, 1981).

Although the importance of plate tectonics in terms of Andean metallogenesis is well recognized, it is also certain that the tectonic and magmatic evolution of some Andean segments include periods when the subduction process was dormant or exhibited little activity. This is the case of the Lower Cretaceous basin in Peru (Atherton and Webb, 1989) and Chile (Levi and Aguirre, 1981). It is possible that under these circumstances, more complex mechanisms participate. An example is that proposed by Márquez *et al.* (1999) for the Mexican Volcanic Belt, involving both an asthenosphere plume and subduction-related process; or the model proposed by Sasso and Clark (1988) for the Andean segment between 26°30'S and 30°30'S, already mentioned in this review.

The comparison of the post-Paleozoic metallogenetic evolution of the Andean belt with that of the island arcs, *e.g.*, the Fidji arc, reveals interesting similarities, specially in terms of increase in both the number of different types of ore deposits and the magnitude attained by the larger ones. For the case of the island arcs, this evolution is parallel to the development of a dioritic tonalitic crust. Thus, at Fidji (Colley and Greenbaum, 1980), this crust was developed during the Tertiary, following a phase of tholeiitic and andesitic volcanism and compression. Not only the number and magnitude of sulphide deposits greatly increased, but also the number of metals involved and the number of types of metallic deposits (from one, massif sulphides, to four, including porphyry copper deposits).

The number of important deposits of Tertiary age in the Andean Belt is amazing, as well as their distribution in or around the central part of the Andes (10 °S to 35 °S), where the continental crust attains its maximum thickness. This is the case for all the metallic provinces, except for the iron belt (though the important Pliocene magnetite deposit of El Laco is in the high Andes at 23°49'S). Certainly, the possible effect of erosion levels should be considered a contributing factor, as the Tertiary hypabysal or subvolcanic intrusive rocks are normally eroded at a level that is favourable both for the exposure and preservation of most types of hydrothermal deposits. However, none of the well preserved pre-Tertiary porphyry copper deposits in the Andes attains the order of magnitude of the larger Tertiary ones, and the same is true for other types of deposits, such as those of the Bolivian tin belt.

In metallogenetic terms, an evolved crust implies a higher degree of structural complexity, better opportunities for magma mixing, contributions from sedimentary strata with different chemical composition etc. Also, a number of geological levels, from the asthenosphere to the sedimentary strata may participate in the generation and differentiation of magmas and in the genesis of the ore deposits resulting in their emplacement and the interaction with the host rocks and fluids in the upper levels of the crust.

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REFERENCES

Aberg, G., Aguirre, L., Levi, B. and Nystrom, J. (1983). Spreading subsidence and generation of ensialic marginal basins: an example from the Early Cretaceous of central Chile. *Journal of the Geological Society, London, Special issue*, **16**, 185-193.

Aguirre, L., Herve, F. and Godoy, E. (1972). Distribution of metamorphic facies in Chile-an outline. *Krystalinikum, Prague*, **9**, 7-19.

Ahlfeld, F. (1967). Metallogenetic epochs and provinces of Bolivia. *Mineralium Deposita*, **2**, 291-311.

Ambrus, J. (1988). Chilean molybdenum resources. In: *International molybdenum encyclopaedia*, ed. Sutulov, A., pp. 87-93. Intermet Publications, Santiago.

Amstutz, G.C. and Fontboté, L. (1987). Yacimientos estratoligados en el sector central del borde móvil andino. Elementos para una síntesis. In: *Investigaciones alemanas recientes en Latinoamérica. Geología.* ed. Miller, H., pp. 123-136. Bonn.

Angelelli, V., Fernández, J.C., Herrera, A. and Aristarain, L. (1970). *Descripción del Mapa Metalogénico de la República Argentina.* Dirección Nacional de Geología y Minería, Buenos Aires, 172 p.

Angulo, R. (1978). *Recursos minerales de Colombia.* Instituto Nacional de Investigaciones Geológico-Mineras, Bogotá. 544 p.

Arenas, M.J. (1988). Geological, mineralogical and chemical characteristics of epithermal precious-metal deposits in southern Perú (Abstract). *Revista Geológica de Chile, Santiago*, **15**, 198-200.

Atherton, M.P. and Webb, S. (1989). Volcanic facies, structure, and geochemistry of the marginal basin rocks of central Perú. *Journal of South American Earth Sciences*, **2**, 241-261.

Atkin, B.P., Injoque, J.L. and Harvey, P.K. (1985). Cu-Fe amphibole mineralization in the Arequipa segment. In: *Magmatism at a plate edge: the Peruvian Andes*, eds. Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D., pp. 261-270. Blackie, Glasgow.

Auboin, J., Borello, A.V., Cecioni, G., Charrier, R., Chotin, P., Frutos, J., Thiele, R. and Vicente, J.C. (1973). Esquisse paléogéographique et structurale des Andes Méridionales. *Revue de Géographique Physique et de Géologie Dynamique*,

15, 11-72.
Audebaud, E., Capdevilla, R., Dalmayrac, B., Debelmas, J., Laubacher, G., Lefevre, C., Marocco, R., Martínez, C., Mattauer, M., Mégard, F., Paredes, J. and Tomasi, P. (1973). Les traits géologiques essentiels des Andes Centrales (Pérou-Bolivie). *Revue de Géographie Physique et de Géologie Dynamique*, 15, 73-114.
Baldwin, J. A. and Pearce, J.A. (1982). Discrimination of productive and non productive porphyric intrusions in the Chilean Andes. 1982. *Economic Geology*, 77, 664-674.
Barrero, D. (1976). *Mapa metalogénico de Colombia (1:5.000.000)*. Instituto Nacional de Investigaciones Geológio-Mineras. INGEOMIN, Bogotá.
Behn, R. and Camus, F. (1997). Transanomalías magnéticas: imagen geofísica de metalotectos y clusters de pórfidos cupríferos andinos. *Proceedings VIII Congreso Geológico Chileno, Antofagasta*, 2, 852-856.
Bellido, E. and Montreuil, L. (1972). *Aspectos generales de la metalogenia del Perú*. Servicio de Geología y Minería, Lima, 149 p.
Bellido, E., Gerard, D. and Paredes, J. (1972). *Mapa Metalogénico del Perú. 1:2.500.000*. Servicio de Geología y Minería, Lima.
Bonson, C.G., Grocott, J. and Rankin, H. (1997). A structural model for the development of Fe-Cu mineralization, Coastal Cordillera, northern Chile (25°15' S - 27°15' S). *Proceedings VIII Congreso Geológico Chileno, Antofagasta*, 3, 1608-1612.
Bookstrom, A.A. (1977). The magnetite deposits of El Romeral, Chile. *Economic Geology*, 72, 1101-1130.
Bossi, G.E. and Viramonte, J.G. (1975). Contribución al conocimiento de la petrología de los yacimientos ferríferos sedimentarios de Zapla y Unchimé (provincias de Jujuy y Salta, República Argentina). *Proceedings II Congreso Iberoamericano de Geología Económica, Buenos Aires*, 5, 181-202.
Burnham, C.E. and Ohmoto, H. (1980). Late stage processes of felsic magmatism. *In: Granitic magmatism and related mineralization*, eds. Ishihara, S. and Takenouchi, S. *Mining Geology, Tokyo, Special Issue*, 8, 1-12.
Caminos, R., Cordani, U. and Linares, E. (1979). Geología y geocronología de las rocas metamórficas y eruptivas de la Precordillera y Cordillera Frontal de Mendoza, República Argentina. *Proceedings II Congreso Geológico Chileno, Arica*, 1, F43-F60.
Camus, F. (1975). Geology of El Teniente orebody with emphasis on wall-rock alternation. *Economic Geology*, 70, 1341-1372.
Camus, F. (1985). Los yacimientos estratoligados de Cu, Pb-Zn y Ag de Chile. In: *Geología y recursos minerales de Chile*, eds. Frutos, J., Oyarzún, R. and Pincheira, M.). *Editorial Universidad de Concepción, Concepción*, 2, 547-635.
Camus, F., Sillitoe, R.H. and Petersen, R., eds. (1996). Andean copper deposits: new discoveries, mineralization styles and metallogeny. *Society of Economic Geologists Special Publication*, 5, 198 p.
Canchaya, S. 1990. Stratabound ore deposits of Hualgayoc, Catamarca, Perú. In: *Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 569-582. Springer Verlag, Berlin.
Cardozo, M. and Cedillo, E. (1990). Geologic-metallogenetic evolution of the Peruvian Andes. *In: Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 38-60. Springer Virlag, Berlin.
Carlson, G.C. (1977). Geology of the Bailadores, Venezuela, massive sulphide deposit. *Economic Geology*, 72, 1131-1141.
Cedillo, E. (1990). Stratabound lead-zinc deposits in the Jurassic Chaucha Formation, Central Perú. In: *Stratabound deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 537-553. Springer Verlag (Berlin).
Clark, A.H., Palma, V.V., Archivald, D.A., Farrar, E., Arenas, M.J. and Robertson, R.C. (1983). Occurence and age of tin mineralization in the Cordillera Oriental, southern Perú. *Economic Geology*, 78, 514-520.
Colley, H. and Greenbaum, D. (1980). The mineral deposits and metallogenesis of the Fidji platform. *Economic Geology*, 75, 807-829.
Coney, P.J. (1970). The geotectonic cycle and the new global tectonics. *Bulletin of the Geological Society of America*, 81, 739-748.
Corvalán, J. (1981). *Plate-tectonic map of the Circum-Pacific region-Southeast quadrant (1:10.000.000)*. Circum-Pacific Council for Energy and Mineral Resources and the American Association of Petroleum Geologists.
Cuadra, W.A. and Dunkerley, P.M. (1991). A history of gold in Chile. *Economic Geology*, 86, 1155-1173.
Davidson, J. and Mpodozis, C. (1991). Regional geologic setting of epithermal gold deposits, Chile. *Economic Geology*, 86, 1174-1186.
Deruelle, B. and Moorbath, S. (1993). A similar magma source for ignimbrites and non-ignimbritic lavas from South-Central Andes. In: Andean Geodynamics. *Proceedings of the ORSTOM/Oxford University International Symposium Oxford*, 351-354.
Dietrich, A., Lehmann, B., Wallianos, A., Traxel, K. and Palacios, C. (1999). Magma mixing in Bolivian tin porphyries. *Naturwissenschaften, Springer Verlag, Berlin*, 86, 40-43.
Di Marco, A. and Mutti, D. (1996). Los depósitos de Cu-Fe en anfibolitas de la Sierra de Cunuputo, Córdoba. *Instituto de Recursos Minerales, Universidad Nacional de La Plata, Argentina, Publicación*, 5, 311-316.
Echavarría, L.E. and Etcheverry, R.O. (1998). Características geoquímicas de la mineralización epitermal del area El Dorado-Monserrat, Provincia de Santa Cruz, Argentina. *Revista Geológica de Chile*, 25, 1, 69-83.
Ericksen, G.E. (1976). Metallogenic provinces of the southeastern Pacific region. *American Association of Petroleoum Geologists, Memoir*, 25, 527-538.
Espinoza, S., Véliz, H., Esquivel, J., Arias, J. and Moraga, A. (1996). The cupriferous province of the Coastal Range, northern Chile. In: *Andean copper deposits: new discoveries mineralization styles and metallogeny*, eds. Camus, F., Sillitoe, R.H. and Petersen, R. *Society of Economic Geologists Special Publication*, 5, 19-32.
Field, C. W. and Gustafson, L.B. (1976). Sulfur isotopes in the porphyry copper deposit at El Salvador, Chile. *Economic Geology*, 71, 1533-1548.
Fontboté, L., Gunnesch, K.A. and Baumann, A. (1990). Metal sources in stratabound ore deposits in the Andes (Andean Cycle) - Lead isotope constraints. *In: Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 759-773. Springer Verlag, Berlin.
Francis, P.W. and Baker, M.C.W. (1978). Sources of two large ignimbrites in the central Andes: some Landsat evidence. *Journal of Volcanology and Geothermal Research*, 2, 73-111.
Francis, P.W., Baker, M.C.W. and Halls, C. (1981). The Karikari Caldera, Bolivia and the Cerro Rico stock. *Journal of Volcanology and Geothermal Research*, 10, 113-124.
Frutos, J. (1975). Porphyry copper-type mineralization and geosynclinal tectonic evolution in the Chilean Andes. *Annales de la Societé Géologique de Belgique*, 98, 5-15.
Frutos, J. and Oyarzún, J. (1975). Tectonic and geochemical evidence concerning the genesis of El Laco magnetite lava flows deposits, Chile. *Economic Geology*, 70, 988-990.
Frutos, J., Fontboté, L., and Amstutz, G.C. (1990). Map of stratabound ore deposits of the Andes. *In: Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 759-773. Springer Verlag, Berlin.
Frutos, J., Oyarzún, J., Shiga, Y. and Alfaro, G. (1990). The El Laco magnetite lava flow deposits, northern Chile: an up-to-date review and new data. In: *Stratabound ore deposits in the Andes*,

eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 681-690. Springer Verlag, Berlin.

Gemutz, I., López, G. and Jiménez, F. (1992). Gold deposits of Southern Ecuador. *SEG Newsletter*, **11**, 1 and 13-16.

Gemutz, I., Little, M. and Giudicci, J. (1996). Precious and base metal deposits in Argentina. *SEG Newsletter*, **25**, 1 and 7-14.

Giacosa, R., Márquez, M. and Pezzuchi, H. (1988). Actualización metalogénica de la región patagónica al sur del paralelo 42° S, República Argentina. *Proceedings Tercer Congreso Nacional de Geología Económica, San Juan*, **3**, A1-A20.

Goosens, P.J. (1969). *Mapa índice mineralógico. República del Ecuador (1:1.000.000)*. Servicio Nacional de Geología y Minería, Quito.

Goossens, P. (1972a). An exhalative volcanic iron sulphide stratabound deposit near San Fernando, Azuay province, Ecuador. *Economic Geology*, **67**, 469-480.

Goossens, P. (1972b). Metallogeny in Ecuadorian Andes. *Economic Geology*, **67**, 458-468.

Goossens, P. and Hollister, V.F. (1973). Structural control and hydrothermal alteration pattern at Chaucha porphyry copper, Ecuador. *Mineralium Deposita*, **8**, 321-331.

Gorzawski, H., Fontboté, L., Field, C.W. and Tejada, R. (1990). Sulfur isotope studies in the zinc-lead mine San Vicente, central Perú. In: *Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 305-312. Springer Verlag, Berlin.

Grant, J.N., Halls, C., Avila, W. and Avila, G. (1976). Igneous geology and the evolution of hydrothermal systems in some sub-volcanic tin deposits of Bolivia. *Proceedings of the Annual Meeting of the Geological Society of London and the Institution of Mining and Metallurgy*, 117-126.

Grant, J.N., Halls, C., Avila, W. and Snelling, N.J. (1979). Potassium-argon ages of igneous rocks and mineralization in part of the Bolivian tin belt. *Economic Geology*, **74**, 838-851.

Grant, J.N., Halls, C., Sheppard, S.M.F. and Avila, W. (1980). Evolution of the porphyry tin deposits of Bolivia. In: *Granitic magmatism and related mineralization*, eds. Ishihara, S. and Takenouchi, S. *Mining Geology, Tokyo, Special Issue*, **8**, 151-173.

Grez, E., Aguilar, A., Henríquez, F. and Nystrom, J.O. (1991). Magnetita Pedernales: A new magmatic iron deposit in Northern Chile. *Economic Geology* **86**(6), 1346-1349.

Gustafson, L.B. and Hunt, J.P. (1975). The porphyry copper deposit at El Salvador, Chile. *Economic Geology*, **70**, 857-912.

Hollister, V.F. (1974). Regional characteristics of porphyry copper deposits of South America. *Transactions of the American Institution of Mining Engineers*, p. 213. NÚMERO?

Ishihara, S. (1977). The magnetite-series and ilmenite-series granitic rocks. *Mining Geology, Tokyo*, **27**, 293-305.

Ishihara, S. (1978). Metallogenesis in the Japanese island-arc system. *Journal of the Geological Society, London*, **135**, 389-406.

Ishihara, S. (1981). The granitoids series and mineralization. In: *Economic Geology 75 Anniversary volume*, ed. Skinner, B.J., 458-484.

Ishihara, S. and Ulriksen, C. (1980). The magnetite-series and ilmenite-series granitoids in Chile. *Mining Geology, Tokyo*, **30**, 183-190.

James, D.E. (1971). Plate tectonic model for the evolution of the Central Andes. *Bulletin of the Geological Society of America*, **82**, 3325-3346.

Larson, A. and Oreskes, N. (1994). The magnetite "lava flows"?, El Laco, Chile: new evidence for formation by water transport. *Proceedings VII Congreso Geológico Chileno, Concepción*, **2**, 1501-1505.

Lehmann, B. (1985). Formation of the strata-bound Kellhuani tin deposits, Bolivia. *Mineralium Deposita*, **20**, 169-176.

Lehmann, B. (1990). The stratabound Kellhuani tin deposits, Bolivia. In: *Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 147-160. Springer Verlag, Berlin.

Lehne, R.W. (1990). The polymetallic ore deposits of Colquijirca, central Perú. In: *Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 615-620. Springer Verlag, Berlin.

Levi, B. and Aguirre, L. (1981). Ensialic spreading-subsidence in the Mesozoic and Paleogene Andes of central Chile. *Journal of the Geological Society, London*, **138**, 75-81.

Levi, B., Nystrom, J.O., Thiele, R. and Aberg, G. (1988). Geochemical trends in Mesozoic-Tertiary volcanic rocks from the Andes in central Chile, and tectonic implications. *Journal of South American Earth Sciences*, **1**(1), 63-74.

Long, K., Ludington, S., du Bray, E., Ramos, O. and Mc Kee, E.H. (1992). Geology and mineral deposits of La Joya district, Bolivia. *SEG Newsletter*, **10**, 1 and 13-16.

López-Escobar, L. and Vergara, M. (1982). Geoquímica y petrogénesis de rocas granodioríticas asociadas con el yacimiento cuprífero de Río Blanco-Los Bronces. *Revista Geológica de Chile, Santiago*, **15**, 59-70.

López-Escobar, L., Frey, F.A. and Vergara, M. (1977). Andesites and high-alumina basalts from the central-south Chile high Andes. Geochemical evidence bearing on their petrogenesis. *Contributions to Mineralogy and Petrology*, **63**, 199-228.

López-Escobar, L., Frey, F.A. and Oyarzún, J. (1979). Geochemical characteristics of central Chile (33°-34° S) granitoids. *Contributions to Mineralogy and Petrology*, **70**, 439-450.

López-Escobar, L., Cembrano, J. and Moreno, H. (1995). Geochemistry and tectonics of the Chilean Southern Andes basaltic Quaternary volcanism (37-46° S). *Revista Geológica de Chile*, **22**(2), 219-234.

Lowell, J.D. and Guilbert, J.M. (1970). Lateral and vertical alteration-mineralization zoning in porphyry ore deposits. *Economic Geology*, **65**, 373-408.

Maksaev, V. and Zentilli, M. (1988). Marco metalogénico regional de los megadepósitos de tipo pórfido cuprífero del Norte Grande de Chile. *Proceedings V Congreso Geológico Chileno, Santiago*, **1**, B181-B212.

Malvicini, L. (1975). La continuación del cinturón occidental de estaño y wolframio de América del Sur en Argentina. *Proceedings V Congreso Latinoamericano de Geología, Buenos Aires*, **2**, 384-404.

Malvicini, L. and Llambías, E. (1982). El magmatismo Mioceno y las manifestaciones metalíferas asociadas en Argentina. *Proceedings V Congreso Latinoamericano de Geología, Buenos Aires*, **3**, 547-566.

Márquez, M.J. (1988). Mineralización polimetálica asociada al magmatismo mioceno en la Cordillera Patagónica austral, provincia de Santa Cruz, Argentina. *Proceedings V Congreso Geológico Chileno, Santiago*, **1**, B237-B256.

Márquez, A., Oyarzún, R., Doblas, M. and Verma, S.P. (1999). Alkalic (ocean-island basalt type) and calc-alkalic volcanism in the Mexican volcanic belt: A case for plume-related magmatism and propagating rifting at an active margin? *Geology*, **27**(1), 51-54.

Macfarlane, A.W., Marcet, P., Le Huray, A.P. and Petersen, U. (1990). Lead isotope provinces of the Central Andes inferred from ores and crustal rocks. *Economic Geology*, **85**, 1857-1880.

Mc Kee, E.H., Robinson, A.C., Rybuta, J.J., Cutiño, L. and Moscoso, R.D. (1994). Age and Sr isotope composition of volcanic rocks in the Maricunga belt, Chile: implications for magma sources. *Journal of South American Earth Sciences*, **7**(2), 167-177.

Mc Kinstry, H.E. (1936). Geology of the silver deposit at Colquijirca, Perú. *Economic Geology*, **31**, 618-635.

Montecinos, P. (1983). *Pétrologie des roches intrusives associées au gisement de fer El Algarrobo (Chile)*. Dr. Thesis, Université de

Paris-Sud, 191 p.

Morris, J., Tera, F., Hammond, R., López-Escobar, L., Klein, J. and Middleton, R. (1985). Be-10 in lavas from the Andean Southern Volcanic Zone (35°40' S): Evidence for sediment subduction. *Comunicaciones, Santiago,* **35**, 147-155.

Munchmayer, C. (1996). Exotic deposits-products of lateral migration of supergene solutions from porphyry copper deposits. In: Andean copper deposits: new discoveries, mineralization styles and metallogeny, eds. Camus, F., Sillitoe, R.H. and Petersen, R. *Society of Economic Geologists Special Publication,* **5**, 43-58.

Munizaga, F., Huete, C. and Hervé, F. (1985). Geocronología K-Ar y razones iniciales $^{87}Sr/^{86}Sr$ de la "Faja Pacífica de Desarrollos Hidrotermales". *Proceedings IV Congreso Geológico Chileno, Antofagasta,* **3**, 4/357-4/479.

Munizaga, F., Holmgren, C., Huete, C. and Kawashita, K. (1988). Geocronología de los yacimientos de cobre El Soldado y Lo Aguirre, Chile central. *Proceedings V Congreso Geológico Chileno, Santiago,* **3**, G177-G193.

Noble, D.C. and Vidal, C.E. (1994). Gold in Perú. *SEG Newsletter,* 17, 1 and 7-17.

Noble, D.C., Mc Kee, E.H., Eyzaguirre, R. and Marocco, R. (1984). Age and regional tectonic and metallogenetic implications of igneous activity and mineralization in the Andahuaylas-Yauri belt of southern Perú. *Economic Geology,* **79**, 172-176.

Nystrom, J. O. and Henríquez, F. (1994). Magmatic features of iron ores of the Kiruna-type in Chile and Sweden: ore textures and magnetite geochemistry. *Economic Geology,* **89**, 820-839.

Ortiz, F. (1990). Massive sulphides in Colombia. In: *Stratabound ore deposits in the Andes,* eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 379-387. Springer Verlag, Berlin.

Oyarzún, J. (1985). *La métallogenie andine: cadre géologique, petrologique et géochemique, et essai d'interpretation.* Sc.D. Thesis, Université de Paris-Sud. 864 p.

Oyarzún, J. (1990). El desarrollo geológico y metalogénico de la cadena andina. *In: Yacimientos Minerales,* eds. Lunar, R. and Oyarzún, R., pp. 695-807. Editorial R. Areces, Madrid.

Oyarzún, J. and Frutos, J. (1974). Porphyry copper and tin-bearing porphyries. A discussion of genetic models. *Physics of the Earth and Planetary Interiors,* **9**, 259-263.

Oyarzún, J. and Frutos, J. (1980). Metallogenesis and porphyry deposits of the Andes (Southeastern Pacific region). XXVI International Geological Congress (Paris). *Memoire du B.R.G.M.,* **106**, 50-62.

Oyarzún, J. and Frutos, J. (1984). Tectonic and petrological frame of the Cretaceous iron deposits of northern Chile. *Mining Geology, Tokyo,* **34**(1), 21-31.

Oyarzún, J., Levi, B. and Nystrom, J.O. (1993). A within-plate geochemical signature and continental margin setting for the Mesozoic-Cenozoic lavas of central Chile. In: Andean Geodynamics. *Proceedings of the ORSTOM/Oxford University International Symposium, Oxford,* 419-422.

Oyarzún, R., Clemmey, H. and Collao, S. (1984). Chemical characteristics of the Nahuelbuta mountains banded iron formations, southern central Chile. *Journal of the Japanese Association of Mineralogists, Petrologists and Economic Geologists,* **79**, 146-156.

Oyarzún, R., Ortega, L., Sierra, J., Lunar, R. and Oyarzún, J. (1996). The manto-type gold deposits of Andacollo (Chile) revisited: a new model based on fluid inclusions and geologic evidence. *Economic Geology,* **91**, 1298-1309.

Palacios, C. and Oyarzún, R. (1975). Relationships between depth to Benioff zone and K and Sr concentrations in volcanic rocks of Chile. *Geology,* **3**, 595-596.

Pareja, J., Vargas, C., Suarez, R., Ballon, A., Carrasco, R. and Villarroel, A. (1978). Mapa geológico de Bolivia. Memoria explicativa. *Y.P.F.B. and Servicio Geológico de Bolivia, La Paz,* 27 p.

Pardo-Casas, F. and Molnar, P. (1987). Relative motion of the Nazca (Farallón) and South American plates since Late Cretaceous time. *Tectonics,* **6**, 233-248.

Park, C.F. (1961). A magnetite "flow" in northern Chile. *Economic Geology,* **56**, 431-436.

Peccerillo, A. and Taylor, S.R. (1976). Geochemistry of Eocene calc-alkaline volcanic rocks from the Kastamonu area, northern Turkey. *Contributions to Mineralogy and Petrology,* **58**, 63-81.

Petersen, U. (1970). Metallogenetic provinces in South America. *Geologie Rundschau,* 59(3), 834-897.

Petersen, U. (1977). Introduction to "An issue devoted to mineral deposits in the South American Cordillera". *Economic Geology,* **72**, 887-892.

Petersen, U. and Vidal, C. (1996.) Magmatic and tectonic controls on the nature and distribution of copper deposits in Perú. In: Andean copper deposits: new discoveries, mineralization styles and metallogeny, eds. Camus, F., Sillitoe, R.H. and Petersen, R. *Society of Economic Geologists Special Publication,* **5**, 1-18.

Petersen, U., Macfarlane, A.W. and Danielson, A. (1993). Lead isotope provinces in Perú, Bolivia and Northern Chile. In: Andean Geodynamics. *Proceedings of the ORSTOM/Oxford University International Symposium, Oxford,* 477-480.

Petford, N. and Atherton, M.P. (1995). Crustal segmentation and isotope significance of the Abancay Deflection: Northern Central Andes (9-20° S). *Revista Geológica de Chile, Santiago,* **22**(2), 235-243.

Pichler, H. and Zeil, W. (1972). The Cenozoic rhyolite-andesite association of the Chilean Andes. *Bulletin Volcanologique,* **35**, 424-452.

Pitcher, W.S., Atherton, M.P., Cobbing, E.J. and Beckinsale, R.D., eds. (1985). *Magmatism at a plate edge: the Peruvian Andes.* Blackie (Glasgow), 328 p.

Puig, A. (1988). Geologic and metallogenetic significance of the isotope composition of lead in galenas of the Chilean Andes. *Economic Geology,* **83**, 843-858.

Puig, A. (1990). Lead isotopes in the Chilean ores. In: Stratabound ore deposits in the Andes, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J. Springer Verlag, Berlin, 749-758.

Reyes, M. (1991). The Andacollo strata-bound gold deposit, Chile, and its position in a porphyry copper-gold system. *Economic Geology,* **86**, 1301-1316.

Routhier, P. (1980). Où sont les métaux por l'avenir?-Les provinces métalliques. Essai de metallogénie globale. *Memoire du BRGM,* Orleans, 105, 410 p.

Ruiz, C. and Peebles, F. (1988). *Geología y yacimientos metalíferos de Chile.* Editorial Universitaria, Santiago, 334 p.

Ruiz, C., Aguirre, L., Corvalán, J., Klohn, C., Klohn, E. and Levi, B. (1965). *Geología y yacimientos metalíferos de Chile.* Instituto de Investigaciones Geológicas, Santiago, 305 p. and maps.

Samaniego, A. (1980). *Strata-bound Pb-Zn (Ag-Cu) ore occurences in Early Cretaceous sediments of north and central Perú-a contribution to their metallogenesis.* Doctoral thesis, Heidelberg University, 214 p.

Sasaki, A., Ulriksen, C., Sato, K. and Ishihara, S. (1984). Sulfur isotope reconnaissance of porphyry copper and manto-type deposits in Chile. In: Research on calc-alkaline magmatism and related mineralization in Chile. *Report N° 7911, ITIT, Geological Survey of Japan-Sernageomin, Chile, Tokyo,* 48-56.

Sasso, A.M. and Clak, A.H. (1988). The Farallón Negro group, northwest Argentina: magmatic, hydrothermal and tectonic evolution and implications for Cu-Au metallogeny in the Andean back-arc. *SEG Newsletter,* **34**, 1 and 8-18.

Scheuber, E. and Reutter, K.J. (1992). Magmatic arc tectonics in the central Andes between 21° and 25° S. *Tectonophysics,* **205**, 127-140.

Schneider, A. (1987). Eruptive processes, mineralization and isotope evolution of the Los Frailes-Karikari region, Bolivia. *Revista Geológica de Chile, Santiago*, **30**, 27-33.

Schneider, H.J. and Lehmann, B. (1977). Contribution to a new genetical concept on the Bolivian tin province. *In: Time and strata-bound ore deposits*, eds. Klemm, D.D. and Schneider, H.J., pp. 153-168. Springer Verlag, Berlin.

Schneider, A. and Toloza, R. (1990). The minig district of the General Carrera lake and the Rosillo manto deposit, Aysen province, southern Chile. In: *Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 485-494. Springer Verlag, Berlin.

Skewes, A. and Arévalo, A. (1997). Andesitas de la mina El Teniente. *Proceedings VIII Congreso Geológico Chileno, Antofagasta*, **2**, 1162-1164.

Sillitoe, R. H. (1972). Relation of metal provinces in western America to subduction of oceanic lithosphere. *Bulletin of the Geological Society of America*, **83**, 813-818.

Sillitoe, R.H. (1973). The tops and bottoms of prophyry copper deposits. *Economic Geology*, **68**, 799-815.

Sillitoe, R.H. (1974). Tectonic segmentation of the Andes: implications for magmatism and metallogeny. *Nature*, **250**, 542-545.

Sillitoe, R.H. (1976). Metallic mineralization affiliated to sub-aereal volcanism: a review. *Procedings Annual Meeting of the Geological Society, London-Institution of Mining and Metallurgy*, 99-116.

Sillitoe, R.H. (1977). Permo-Carboniferous, Upper Cretaceous and Miocene porphyry copper type mineralizations in the Argentinean Andes. *Economic Geology*, **72**, 99-103.

Sillitoe, R.H. (1983). Enargite-bearing massive sulphide deposits high in porphyry copper systems. *Economic Geology*, **78**, 348-352.

Sillitoe, R.H. (1988). Epochs of intrusion-related copper mineralization in the Andes. *Journal of South American Earth Sciences*, **1**(1), 89-108.

Sillitoe, R.H. (1991). Gold metallogeny of Chile-an Introduction. *Economic Geology*, **86**, 1187-1205.

Sillitoe, R.H. (1992). Gold and copper metallogeny in the Central Andes-past, present and future exploration objectives. *SEG Distinguished Lecture. Economic Geology*, **87**, 2205-2216.

Sillitoe, R.H. and Sawkins, F.J. (1971). Geologic, mineralogic and fluid inclusion studies relating to the origin of copper-bearing tourmaline breccia pipes, Chile. *Economic Geology*, **66**, 1208-1041.

Sillitoe, R.H., Halls, C. and Grant, J.N. (1975). Porphyry tin deposits in Bolivia. *Economic Geology*, **70**, 913-927.

Sillitoe, R.H., Jaramillo, L., Damon, P., Shafiqullah, M. and Escovar, R. (1982). Setting, characteristics and age of the Andean porphyry copper belt in Colombia. *Economic Geology*, **77**, 1837-1850.

Sillitoe, R.H., Marquardt, J.C., Ramírez, F., Becerra, H. and Gómez, M. (1996). Geology of the concealed MM porphyry copper deposit, Chuquicamata District, northern Chile. In: *Andean copper deposits: New discoveries, mineralization styles and metallogeny*, eds. Camus, F., Sillitoe, R.H. and Petersen, R. *Society of Economic geologists. Special Publication*, **5**, 59-69.

Soler, P. and Bonhomme, M.G. (1988). New K-Ar age determinations of intrusive rocks from the Cordillera Occidental and Altiplanó of central Perú: identification of magmatic pulses and episodes of mineralization. *Journal of Soth American Earth Sciences*, **1**(2), 169-177.

Soler, P., Grandin, G. and Fornari, M. (1986.) Essai de synthese sur la métallogenie du Pérou. *Géodynamique*, **1**(1), 33-68.

Spiro, B. and Puig, A. (1988). The source of sulphur in polymetallic deposits in the Cretaceous magmatic arc, Chilean Andes. *Journal of South American Earth Sciences*, **1**(3), 261-266.

Stoll, W.C. (1975). Yacimientos stratabound de scheelita en el basamento cristalino de Argentina (Abstract). *Proceedings II Congreso Iberoamericano de Geología Económica, Buenos Aires*, 3, 76.

Sureda, R.J. and Martin, J.L. (1990). El Aguilar mine: an Ordovician sediment-hosted stratabound lead-zinc deposit in the Central Andes. *In: Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 161-174, Springer Verlag, Berlin.

Sureda, R., Galliski, M., Argañaraz, P. and Daroca, J. (1986). Aspectos metalogénicos del noroeste argentino (provincias de Salta y Jujuy). *Capricornio (Salta)*, **1**, 39-96

Thiele, R. and Pincheira, M. (1987). Tectónica transpresiva y movimiento de desgarre en el segmento sur de la Zona de Falla Atacama, Chile. *Revista Geológica de Chile, Santiago*, **31**, 77-94.

Thorpe, R.S. and Francis, P.W. (1979). Petrogenetic relationships of volcanic and intrusive rocks of the Andes. *In: Origin of granitic batholithes. Geochemical evidence*, eds. Atherton, M.P. and Tarney, J., pp. 65-75. Shiva Publications Kent, U.K.

Turneaure, F.S. (1960). A comparative study of major ore deposits of central Bolivia. *Economic Geology* **55**, 217-254 and 574-606.

Turneaure, F.S. (1971). The Bolivian tin-silver province. *Economic Geology*, **66**, 215-225.

Travisany, V., Henríquez, F. and Nystrom, J.O. (1995). Magnetite lava flows in the Pleito-Melón district of the Chilean iron belt. *Economic Geology*, **90**, 438-444.

Uyeda, S. and Kanamori, H. (1979). Back-arc opening and mode of subduction. *Journal of Geophysic Research*, **84**, 1049-1061.

Vila, T., Lindsay, N. and Zamora, R. (1996). Geology of the Manto Verde copper deposit, northern Chile: A specularite-rich, hydrothermal-tectonic breccia related to the Atacama Fault Zone. In: *Andean copper deposits. New discoveries, mineralization styles and metallogeny*, eds. Camus, F., Sillitoe, R.H. and Petersen, R. *Society of Economic Geologists. Special Publication*, **5**, 157-169.

Vidal, C. (1987). Kuroko-type deposits in the Middle Cretaceous marginal basin of Central Perú. *Economic Geology*, **82**, 1409-1430.

Vidal, C., Injoque-Espinoza, J., Sedder, G.B. and Mukasa, S.B. (1990). Amphibollitic Cu-Fe skarn deposits in the central coast of Perú. *Economic Geology*, **85**, 1447-1461.

Vivallo, W. and Henríquez, F. (1998). Génesis común de los depósitos estratoligados y vetiformes de cobre del Jurásico Medio a Superior en la Cordillera de la Costa, Región de Antofagasta, Chile. *Revista Geológica de Chile*, **25**, 199-228.

Vivallo, W., Henríquez, F. and Espinoza, S. (1994). Oxygen and sulfur isotopes in hydrothermally altered rocks and gypsum deposits at El Laco mining district, nothern Chile. *Comunicaciones, Santiago*, 93-100.

Wellmer, F.W. and Reeve, E.J. (1990). The Toqui zinc-lead-copper-silver deposits, Aysen Province, Chile. In: *Stratabound ore deposits in the Andes*, eds. Fontboté, L., Amstutz, G.C., Cardozo, M., Cedillo, E. and Frutos, J., pp. 473-484. Springer Verlag, Berlin.

Wellmer, F.W., Reeve, E.J., Wentzlau, E. and Westenberger, H. (1983). Geology and ore deposits of El Toqui district, Aysén, Chile. *Economic Geology*, **78**, 1119-1143.

Zeil, W. (1979). *The Andes. A geological review*. Gebruder Bontraeger, Berlin, 260 p.

Zentilli, M. (1974). *Geological evolution and matallogenetic relationships in the Andes of northern Chile between 26° and 29° S*. Ph.D. Thesis, Queen University, Ontario, 394 p.

Zentilli, M., Doe, B.R., Hedge, C.E., Alvarez, O., Tidy, E. and Daroca, J.A. (1988). Isótopos de plomo en yacimientos de tipo pórfido cuprífero comparados con otros depósitos metalíferos en los Andes de Chile y Argentina. *Proceedings V Congreso Geológico Chileno, Santiago*, **1**, B331-B369.

The Metallogenesis of the South American Platform

Marcel Auguste Dardenne and Carlos Schobbenhaus

The main objective of this review is to describe the most pertinent aspects of the metallogeny of the South American Platform, aiming to define the most important metallogenic units that developed during its evolution in the way of economically significant mineral provinces, and illustrating these with selected examples. The South American Platform represents, together with the Patagonia Platform and the Andean Belt, one of the principal tectonic divisions of the South American Continent. The consolidation of the South American Platform was completed by the end of the Neoproterozoic (Almeida *et al.*, 1976) (Figs. 1 and 2).

The constraints imposed by the space in which to adequately treat the far-reaching nature and diversity of the subject have necessitated that the selected models be described only in synthesis. Thus the opportunity for a more profound scientific discussion was reduced. For similar reasons the references cited are restricted to those considered to be wider in scope and in which additional references on the specific matter are cited.

The geochronological divisions adopted follow the recommendations of the International Commission on Stratigraphy (ICS/IUGS). However, for the Proterozoic we have preferred to adopt a three-part division, with boundaries at 1.8 Ga and 1.0 Ga, in line with the Tectonic Map of South America (Almeida, 1978). On the other hand the Proterozoic-Cambrian boundary is placed at 544 Ma, following the suggestion of Bowring *et al.* (1993).

The Geotectonic Framework of the South American Platform

The South American Platform forms the nucleus or core of South America. It covers an area of about 15 M km^2, some 40% of which is exposed in three Precambrian shields: Guiana, Central Brazil (or Guaporé) and Atlantic. About 34% of the continental crust exposed in these shields was formed in the Archean, 80% was formed during the late Paleoproterozoic by the end of the Transamazonian Cycle, and about 98% at the end of the Neoproterozoic Brasiliano cycle (Cordani *et al.*, 1988;Cordani and Sato, 1999). At the end of the Neoproterozoic the South American Platform consisted of several plates or independent cratonic nuclei, most of which were still aggregated to their African counterparts. Between *c.* 650 and 540 Ma, the final amalgamation of these terranes (Fig. 3) was performed by a series of collisions during the Brasiliano (PanAfrican) Orogenic Cycle. The most important Brasiliano cratons of the South American Platform are the Amazonian, São Francisco, and Rio de La Plata cratons, in addition to smaller continental fragments. The basement of these cratons consists essentially of medium and high-grade metamorphic rocks, including associations of the granite-greenstone belt type and numerous granitoid plutons. Fragments of Archean medium to high-grade metamorphites occur as inliers in the Proterozoic mobile belts.

Geochronological data show that the evolution of the Amazonian Craton involved the addition of juvenile material performed by a number of tectonic events during the Archean, the Paleoproterozoic and the Mesoproterozoic, as well as the reworking of older continental crust. In the high-grade terranes (granulite and gneiss) some Archean protoliths occur in the Imataca Complex, Venezuela with U/Pb and Sm/Nd ages between 3.7 and 3.4 Ga (Sidder and Mendoza, 1995; L.A. Bizzi, pers. com.).

In the State of Amapá, Brazil, tonalitic rocks gave U/Pb and Sm/Nd ages between 3.1 and 2.94 Ga, whereas granulites showed Rb/Sr ages between 3.35 and 2.45 Ga (Lafon et al., 1998). In the Province of Carajás, Brazil, it seems that the principal time for the formation of continental crust is constrained between 3.0 and 2.8 Ga. However, zircon crystals aged up to 3.7 Ga have been reported in Paleoproterozoic granite and in sedimentary rocks. (Rio Maria granite-greenstone terrane). (Tassinari and Macambira, 1999). The oldest rocks found so far in South America occur in the Gavião Block of the São Francisco Craton, yielding Sm/Nd T_{DM} model ages up to 3.7 Ga (Cordani and Sato, 1999). In general, the majority of the radiometric results from this craton are of Neo-Archean age, between 2.9 and 2.5 Ga, and they are most prevalent in granite-greenstone terranes.

Archean granite-greenstone terranes or similar sequences form extensive areas in the interior of ancient cratons of the South American Platform. These include; (a) the Province of Carajás (Central Brazil Shield); (b) the Gavião Block and the Quadrilátero Ferrífero of the São Francisco Craton and; (c) the Crixás area of the Central Goiás Massif. Some of these granite-greenstone associations show Paleoproterozoic ages of about 2.2-2.1 Ga, such as the Rio Itapicuru region (São Francisco Craton) and an extensive belt in the NNE of the Guiana Shield. Banded iron formation units (BIF) of the Superior Province type (Paleoproterozoic), the Carajás type (Archean), and the Algoma-type (Archean) as well as sediments of the Witwatersrand type are also found in some of the cratonic areas.

With the close of the Brasiliano Orogenic Cycle, a network of mobile belts may be distinguished separating the cratonic areas. On the Atlantic side these mobile belts are especially well represented by the Brasília, Araçuai,

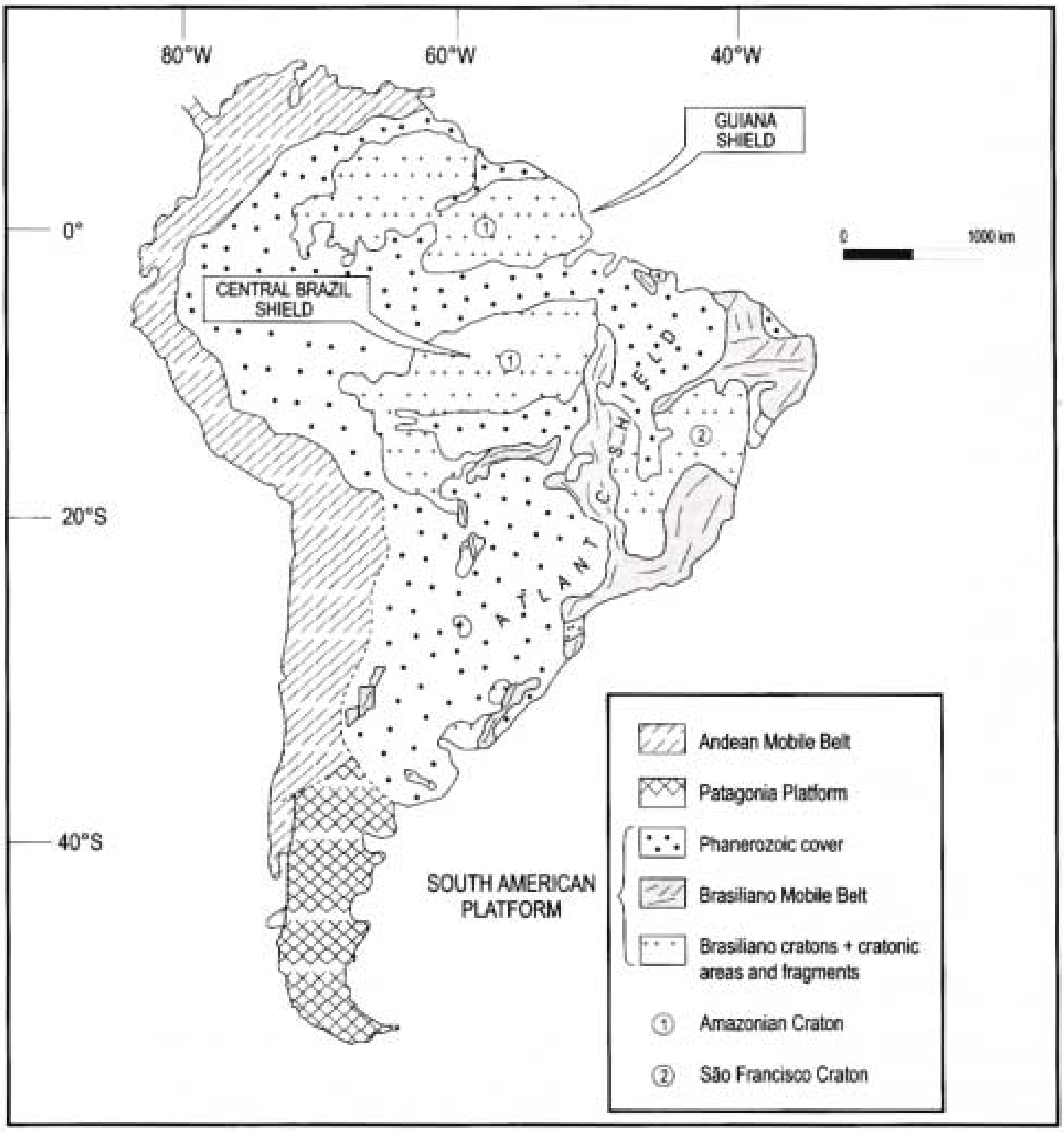


FIGURE 1 - Tectonic division of the South American Platform (modified after Almeida et al., 1978)

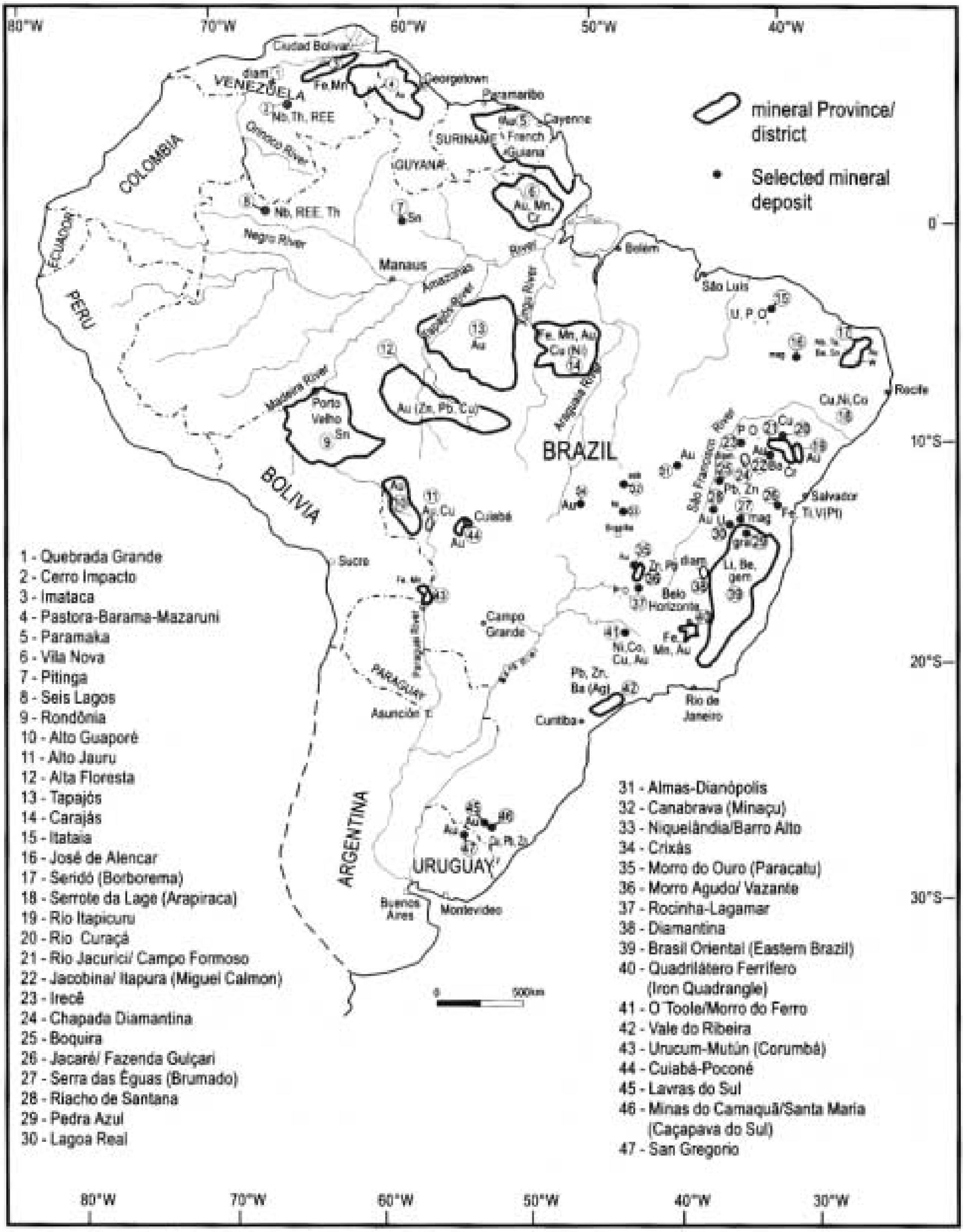


FIGURE 2 - Distribution of main Precambrian mineral provinces and selected mineral deposits of the South American Platform. Sources are cited in text. Abbreviations: asb - asbestos; diam. - diamond; gem - gemstones; gra. - graphite; mag - magnesite; REE - Rare Earths.

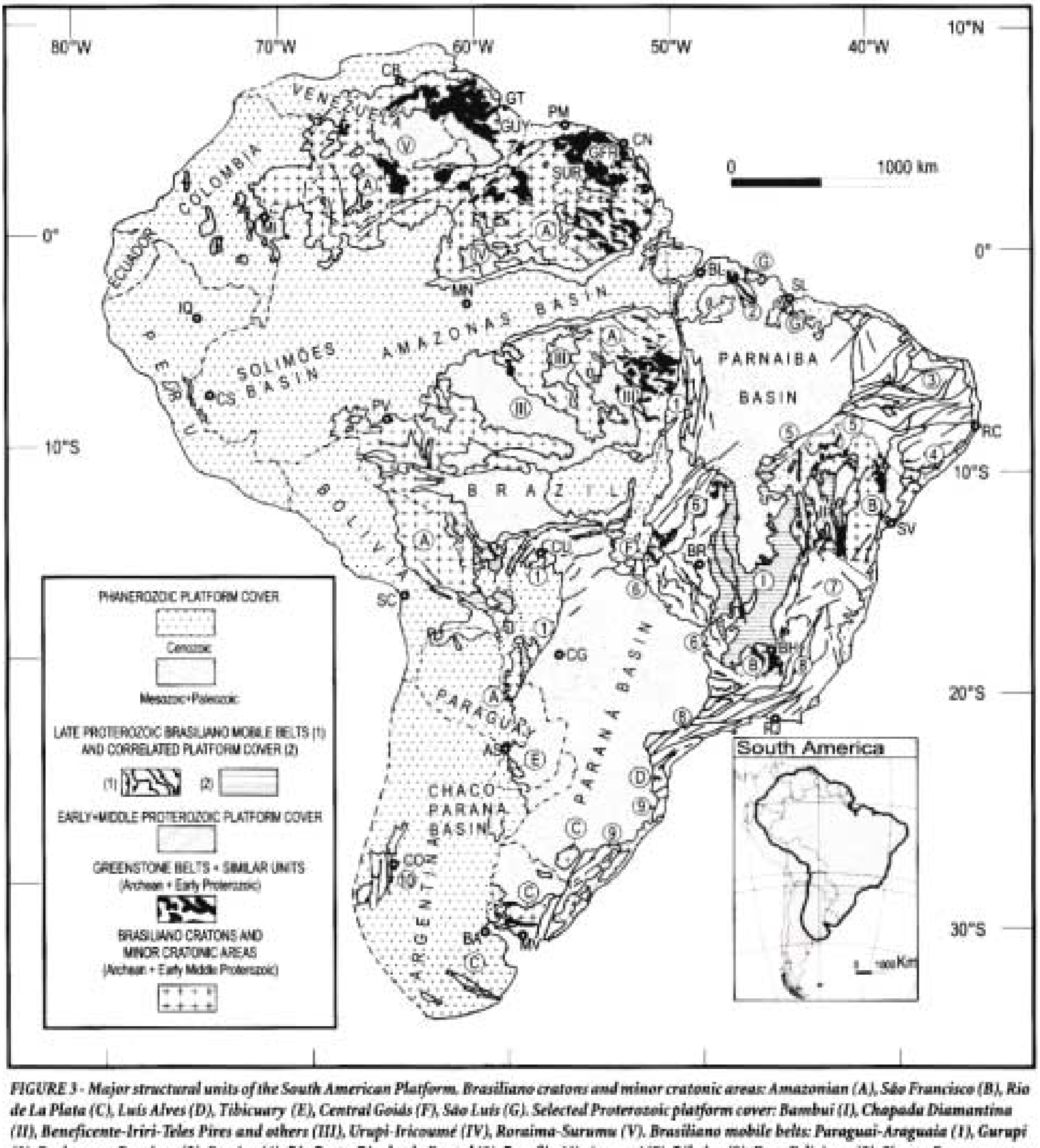


FIGURE 3 - Major structural units of the South American Platform. Brasiliano cratons and minor cratonic areas: Amazonian (A), São Francisco (B), Rio de La Plata (C), Luís Alves (D), Tibicuary (E), Central Goiás (F), São Luís (G). Selected Proterozoic platform cover: Bambuí (I), Chapada Diamantina (II), Beneficente-Iriri-Teles Pires and others (III), Urupi-Iricoumé (IV), Roraima-Surumu (V). Brasiliano mobile belts: Paraguai-Araguaia (1), Gurupi (2), Borborema Province (3), Sergipe (4), Rio Preto-Riacho do Pontal (5), Brasília (6), Araçuaí (7), Ribeira (8), Dom Feliciano (9), Sierras Pampeanas Orientales (10). Modified after Schobbenhaus and Campos (1984), Almeida et al. (1978) and others sources cited in text.
AS: Asunción, BA: Buenos Aires, BH: Belo Horizonte, BL: Belém, BR: Brasília, CB: Cuidad Bolivar, CN: Cayenne, CO: Córdoba, CS: Cruzeiro do Sul, CU: Cuiabá, GFR: French Guiana, GUY: Guyana, GT: Georgetown, IQ: Iquitos, MI: Mitu, MN: Manaus, MV: Montevideo, PM: Paramaribo, PV: Porto Velho, RC: Recife, RJ: Rio de Janeiro, SC: Santa Cruz de la Sierra, SL: São Luís, SUR: Suriname, SV: Salvador.

Ribeira, Dom Feliciano belts and by the Borborema Province. In central Brazil, the Paraguay-Araguaia Belt, over 2500 km long, extends into Bolivia and Paraguay. The relationship between the Paraguai-Araguaia Belt with the Sierra Pampeanas Orientales Belt, situated in the southwestern extremity of the Platform in the interior of Argentina, is not clear. Along the western margin of the Rio de la Plata Craton, there occurred during the Brasiliano Cycle, in early Cambrian times, a collision with the Pampia Terrane, that resulted in the Sierras Pampeanas Orientales mobile belt (Ramos, 1988). The Brasiliano mobile belts usually contain metasediments and metavolcanic rocks of low to medium metamorphic grade, and high metamorphic grade, locally. Many different types of granitoid pluton intrude these rocks. In part, these belts include older reworked rocks. The Borborema Province is different to the other mobile belts. It consists of branching system of orogens developed around a number of small cratonic nuclei, mostly Paleoproterozoic terranes, or rarely, Archean terranes.

The characteristic feature of the Amazonian and São Francisco cratons is the extensive platform cover consisting of sediments and volcanic rocks, mainly of Mesoproterozoic or Neoproterozoic age. These rocks have undergone little or no deformation, and generally contain well-preserved primary structures. This is probably the largest exposure of this type of cover in the world. In the Amazonian Craton this platform cover was deposited between *c.* 1.95 and 1.0 Ga, and was subsequently intruded by anorogenic granitoid plutons. The most important phase of magmatism (*c.* 1.95-1.8 Ga) is represented by calc-alkaline acid to intermediate volcanism of the Uatumã type, the rocks of which are overlain by mature sediments typical of continental and shallow-water marine deposition (Roraima/Beneficente type). Over the São Francisco Craton there are observed large exposures of platform cover consisting of terrigenous clastic and carbonate rocks of Mesoproterozoic and Neoproterozoic age (Espinhaço, Chapada Diamantina, Arai and Bambuí types). Anorogenic magmatism, associated with the opening of diverse continental rifts (Espinhaço-Arai) occurred in this craton at the beginning of the Mesoproterozoic between 1.77 and 1.70 Ga.

Fragmentation of the cratons and the development of Brasiliano oceanic basins started to occur at the beginning of the Neoproterozoic between 1.10 Ga and 950 Ma, involving extensive areas of the proto-South American Platform. The closure of these basins during the Brasiliano Orogenic Event resulted in the amalgamation of several cratons and smaller cratonic areas or terranes, a process that continued up to the beginning of the Paleozoic, and terminated with the consolidation of the present-day South American Platform.

The Precambrian basement of the South American Platform is partially covered by: (a) five large Paleozoic intracratonic basins: Solimões, Amazonas, Parnaiba, Paraná and Chaco-Paraná, the last-named being covered by extensive (Milani and Zalán, 1999) Cenozoic deposits; (b) several smaller Mesozoic/Cenozoic basins situated along the Atlantic coast (Cainelli and Mohriak, 1999) and; (c) the subandean basins in the extensive Andean foredeep (Llanos, Beni, Chaco, Pampas) lying along the margin of the Andean Cordillera, and almost totally covered by Cenozoic sediments.

From the end of the Brasiliano Orogenic Cycle this extensive Precambrian landmass was almost unaffected by tectonic events. Platform reactivation only occurred in the Mesozoic with the opening of the South Atlantic (South Atlantic Event) and continued up to the beginning of the Tertiary. Mainly during the Cretaceous there occurred the eruption of enormous volumes of basaltic lava and the intrusion of a number of alkaline-carbonatite complexes and kimberlitic pipes. The generation of sedimentary rift basins along the Atlantic coast is likewise related to this geotectonic event (Schobbenhaus and Brito Neves, 1996).

The Amazonian Craton

The Amazonian Craton is one of the largest cratonic areas in the world, and has a surface of about 4.3 M km^2. It stabilized at the end of the Mesoproterozoic (Almeida *et al.*, 1976; Cordani *et al.*, 1988). Geographically, the Amazonian Craton is separated into two blocks by the Solimões-Amazonas basins (Fig. 3): the Guiana Shield and the Central Brazil (or Guaporé) Shield (Fig. 3).

According to the model proposed by Cordani and Brito Neves (1982), Lima (1984), Teixeira *et al.* (1989), Tassinari (1996), Tassinari and Macambira (1999), Cordani and Sato (1999), the geotectonic evolution that led to the cratonization of the Amazonian region resulted from a process of progressive crustal accretion from a more ancient nucleus that stabilized at the end of the Archean at about 2.5 Ga (Macambira and Lafon, 1995). Mobile and/or geochronological belts that succeeded each other in time and space surrounded this more ancient nucleus. In accordance with this concept, the Amazonian Craton was divided into six geochronological provinces (Fig. 4): Central Amazonia (>2.3 Ga); Maroni-Itacaiúnas (2.2-1.95 Ga); Ventuari-Tapajós (1.95-1.8 Ga); Rio Negro-Juruena (1.8-1.55 Ga); Rondônia-San Ignácio Belt (1.55-1.3 Ga) and Sunsas Belt (1.3-1.0 Ga).

Each of these provinces consists of specific plutonic, volcanic and sedimentary associations having distinct lithological, geochronological, geochemical and isotopic characteristics. The progressive stabilization of the Amazon region occurred during the Proterozoic through successive manifestations and ruptile intracratonic tectonics as can be seen through the development of rift mechanisms with associated volcanism, continental and marine sedimentation and the migration of anorogenic plutonism. These events terminated at the end of the Mesoproterozoic at about 1.0 Ga, at which time the Amazonian Craton was juxtaposed along its southern and eastern margins to the Paraguai-Araguaia Fold Belt.

The Guiana Shield

The Guiana Shield extends from the Atlantic Ocean at the northern and northeastern ends of the South American Platform to the Amazon sedimentary basin to the S (Fig. 5). In its western part this shield consists of the basement of the Andean foreland basins. Geological maps at adequate scales and/or systematic geochronological studies are

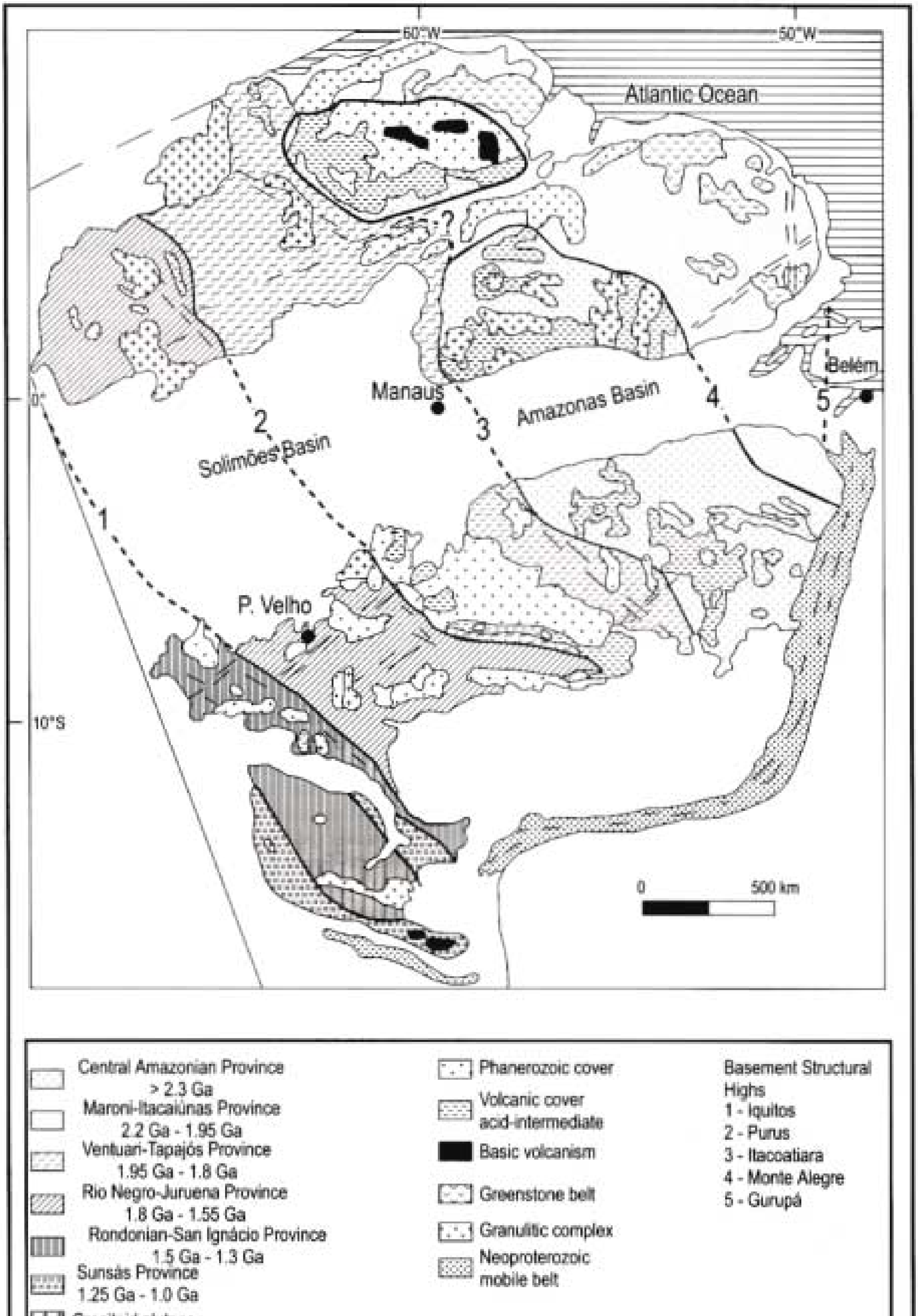


FIGURE 4 - Schematic map of Amazonian Craton, showing the distribution of geochronological provinces (modified after Teixeira et al., 1989; Tassinari, 1996; Tassinari and Macambira, 1999).

restricted to specific areas and are practically non-existent in the largest part of the shield.

The Guiana Shield may be divided into (a) a terrane consisting of granulite and gneiss in the western region of Venezuela as an Archean protolith (Imataca Complex); (b) a Paleoproterozoic granite-greenstone terrane some 300 to 400 km wide, lying along the Atlantic margin; (c) a non-differentiated terrain consisting of granite and gneiss and; (d) a central and western part with extensive cover of Paleoproterozoic felsic volcanic rocks and Mesoproterozoic continental sediments. Additionally, there occur Mesoproterozoic or younger dykes, sills and mafic flows besides alkaline complexes, also of Mesoproterozoic age. Paleoproterozoic events occurring at about 2.0 Ga (Transamazonian Orogenic Event) involved metamorphism, deformation and granitic magmatism, manifest in the Imataca Complex, the granite-greenstone terranes and to some extent the non-differentiated terranes consisting of granitic rocks (Gibbs and Barron, 1983).

In the extensive regions of the Guiana Shield underlain by non-differentiated granite and gneiss terranes and by granite-greenstone terranes, Tassinari and Macambira (1999) defined several geochronological provinces such as those referred to in the previous item (Fig. 4).

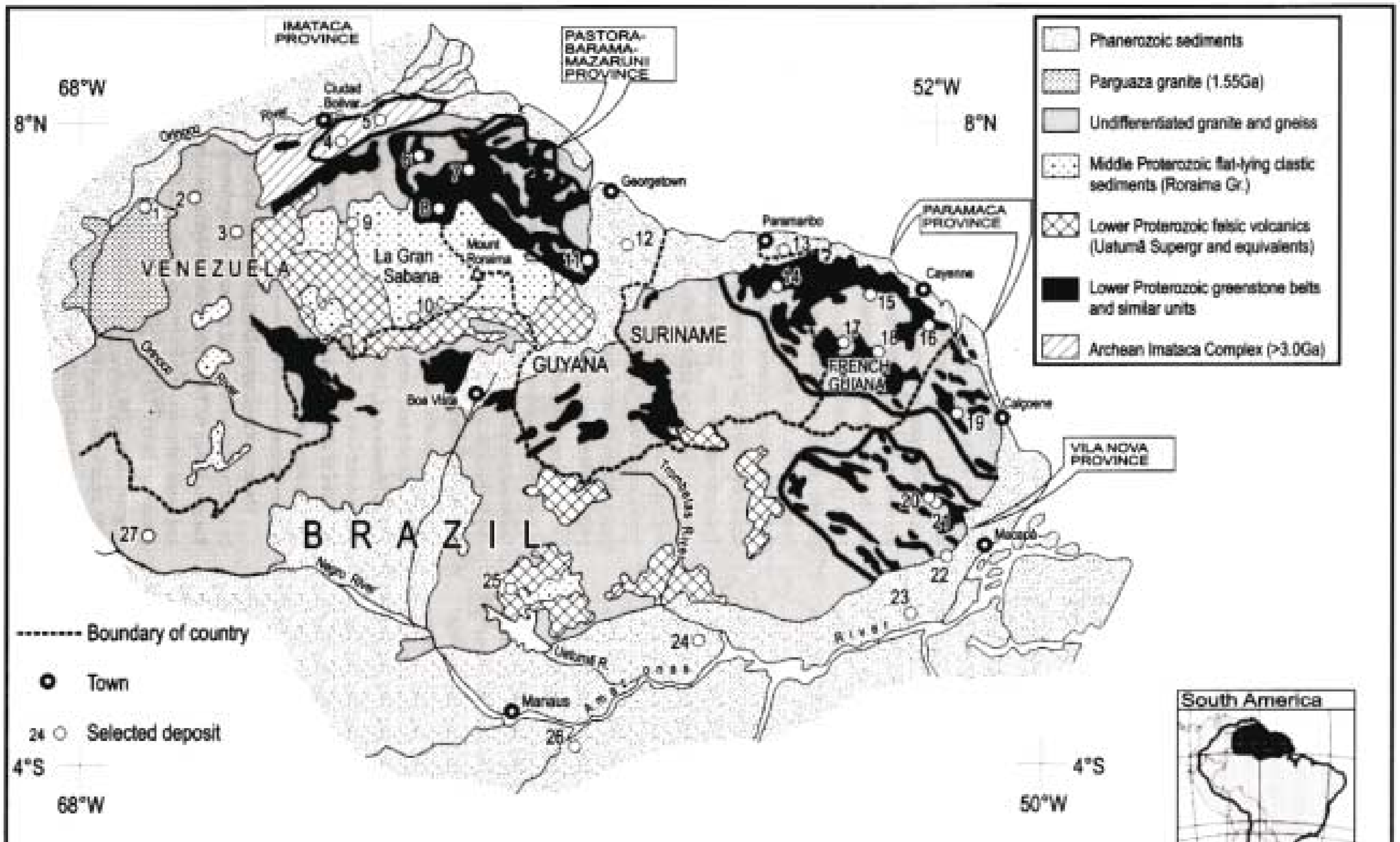
Terranes underlain by rocks of defined Archean age are rare in the Guiana Shield. The area having the largest exposures is that occupied by the Imataca Complex with ages between 3.7-3.4 Ga (U/Pb, Sm/Nd). The granulite and gneiss of the Imataca Complex are the oldest rocks of the shield, and display faulted contacts with the surrounding Precambrian rocks. At about 3.4 Ga this complex formed a stabilized continental nucleus. Between the various rock-types, of note are the intercalations of iron formation units, to which the origin of an important metallogenic province is related. Intrusion of granitic rocks and development of migmatite (Cerro La Ceiba migmatite) define a tectonomagmatic event between 2.8 Ga (zircon age/SHRIMP) and 2.7 Ga (Rb/Sr WR age) in the Imataca Complex (Sidder and Mendoza, 1995; Gibbs and Barron, 1993; L.A. Bizzi, personal communication). This event is known as the Aroense Event (Martin-Bellizzia, 1972). An extensive granite-greenstone province of Paleoproterozoic age and predominantly auriferous, lies along the Atlantic margin of the shield from Venezuela to Brazil. These units also include supracrustal formations that differ from the greenstone belt sequences in their relative paucity of volcanic rocks, and in some cases, by their younger stratigraphic position. Geochronological data, including that for U/Pb isochrons in zircon, Sm/Nd and Rb/Sr, suggest that the volcanic rocks of greenstone belt sequences and granite associated with the provinces under discussion were formed between about 2.25 and 2.10 Ga. In French Guiana, recent dating showed ages between 2.14 and 2.09 Ga for the Paramaka Volcanics (Carte Geologique de la Guyane Française, 1:500 000, in press). In general, the greenstone belt rocks of the Guiana Shield show the same typical characteristics as their Archean counterparts in other parts of the world. However, they contain smaller amounts of ultramafic rocks and larger amounts of clastic sediments. The granite-greenstone terranes represent the most important metallogenic unit of the shield, principally on account of the associated gold mineralization. Three different provinces may be distinguished: (a) Pastora-Barama-Mazaruni, in Venezuela and Guyana; (b) Paramaka, in Suriname, French Guiana and Brazil (State of Amapá) and; (c) Vila Nova, on the divide between the states of Pará and Amapá in Brazil.

In the central part of the shield in Brazil, Guyana and Suriname, excluding the areas referred to above, there occur exposures of volcano-sedimentary supracrustal rocks that underwent medium to high-grade metamorphism during the Transamazonian Event. In Brazil, these supracrustal rocks include the Cauarane Group.

Late Paleoproterozoic calc-alkaline continental volcanic rocks of acid to intermediate composition, locally associated with clastic sediments, can be observed over much of the central and western regions of the shield. These volcanic rocks are associated, in part, with granitic to granodioritic intrusions (*c.* 1.85 Ga), locally of sub-volcanic nature. This volcano-plutonism, to which the collective name of Uatumã has been assigned (Gibbs and Barron, 1983, 1993) is widely distributed throughout the Guiana Shield. However, Reis and Fraga (1996) consider that the cogenetic character of the Uatumã plutonic and volcanic rocks is inconsistent, taking in account the clear geochemical incompatibility of both. Indeed, the Uatumã volcanic rocks are more compatible with those observed in the post-orogenic suites related to the Transamazonian Event. The volcanic rocks of the Uatumã type occur at shallow crustal levels, suggesting that these might have been affected by the Transamazonian Event, and for this reason were interpreted as being related to a late or post-collision phase of this tectonic event (Bosma *et al.*, 1983; Reis and Fraga, 1996).

Several regional names are applied to the volcanic rocks of the Uatumã Supergroup: Surumu and Iricoumé in Brazil, Cuchivero in Venezuela, Burro-Burro and Kuyuwini in Guyana, and Dalbana in Suriname. U/Pb determinations in zircon from rhyolite from the Surumu and Iricoumé formations gave ages of 1.96 Ga, and samples of ash flow from the Cuchivero Group collected near Santa Helena de Uairén were dated at 1.98 Ga (Brooks *et al.*, 1995). A granite intrusive assigned to the Iricoumé Group at Pitinga situated to the NE of Manaus was dated at about 1.8 Ga by U/Pb (Lenharo, 1998). This type of granite hosts important tin mineralization in addition to other elements.

The units of the Uatumã Event are overlain discordantly by continental sediments of the early-Mesoproterozoic Roraima Group (*c.* 1.8 Ga), intruded by dykes, basic sills and small subordinate intrusives of continental tholeiite (Avanavero Suite). The most precise date for this tholeiitic magmatism is 1.789 Ga (U/Pb in baddeleyite) (Norcross, 1998; *apud* Santos *et al.*, 1999). The Roraima platform cover extends over much of the western part of the shield. It represents the most important unit of continental sedimentary rocks in the Guiana Shield that form tablelands (*tepuis*) typically represented by the Mount Roraima, at the triple divide between Brazil, Guiana and Venezuela, where this group attains a thickness of about 2500 m. The Roraima Group consists of orthoquartzite, arkose, conglomerate and smaller amounts of shale and tuff with jasper, deposited in fluvio-deltaic and lacustrine environments. It is believed that the conglomerate units



1. Los Pijiguaos (Al), 2. Quebrada Grande (Diamond), 3. Cerro Impacto (Th, Nb, REE), 4. San Isidro Iron Quadrilateral/Cerro Bolivar (Fe), 5. El Pao (Fe), 6. El Callao/ Lo Increible(Au), 7. Botanamo (Au), 8. Las Cristinas (Au), 9. San Salvador de Paúl (Diamond), 10. Icabarú (Diamond), 11. Omai (Au), 12. Berbice and Demerara (Al), 13. Moengo/Onverdacht/Paranam (Al), 14. Gross Rosebel (Au), 15. St Elie (Au), 16. Camp Caiman (Au), 17. Yaou (Au), 18. Dorlin (Au), 19. Laborrie Siboa/Salamangone (Au), 20. Amapari (Au), 21. Serra do Navio (Mn), 22. Igarapé do Breu (Cr), 23. Almeirim/Mazagão (Al), 24. Oriximiná/Porto Trombetas (Al), 25. Pitinga (Sn, Zr, Nb-Ta, REE), 26. Fazendinha (K), 27. Seis Lagos (Nb, REE)

FIGURE 5 - Schematic geological map of the Guiana Shield (modified after Gibbs and Barron, 1993; Bellizzia et al., 1981; Bardoux et al., 1998; Carvalho et al., 1995; Faraco and Carvalho, 1994; Gray et al., 1993; Brooks et al., 1995; Sidder, 1995)

intercalated in the sediments of the Roraima Group are responsible for the extensive alluvial and eluvial diamond mineralization that occurs regionally (Fleischer, 1998). The youngest of the Mesoproterozoic manifestations of granitic intrusion in the Guiana Shield is the Parguaza Batholith (1.55 Ga), situated near the margin of the Orinoco River in Venezuela (Fig. 5). Most of the Parguaza-type granites show rapakivi textures, anorogenic characteristics, and bear tin mineralizations. A good example example is the Surucucus Granite (1.551 Ga, U/Pb; Santos *et al.*, 1999), situated along the boundary between Brazil and Venezuela. Overlying the Parguaza Batholith there developed important supergene concentrations of bauxite. In Brazil there are syenite intrusions of normal and alkaline composition to which an age of about 1.5 Ga has been assigned. Examples are Cachorro, Serra do Acari and Mapari. Along the frontier between Brazil and Guiana there occurs the Mutum alkaline intrusion dated at *c.* 1.0 Ga. In the interior of the Guiana Shield there is evidence for faulting, reworking and the closing of isotopic systems in mica besides igneous intrusions at the end of the Mesoproterozoic and the beginning of the Neoproterozoic. These events are probably related to more intense activity at the margin of the shield between 1.3 and 1.1 Ga. Most of these faults strike NE-SW, and are associated with mylonite and pseudotachylite. This event has been referred to as Orinoquense, K'Mudku, Nickerie or Jari-Falsino. It is possible that these faults involved significant reactivation along zones of weakness, and as such could play an important role in the concentration of gold.

The igneous activity at the end of the Neoproterozoic includes the intrusion of alkaline complexes and volcanic mafic and felsic suites. The youngest Precambrian magmatic events observed on the Guiana Shield are those that refer to the Quebrada Grande diamondiferous kimberlite in Venezuela, dated at 710 Ma (Channer *et al.*, 1998), and to basic dykes in French Guiana that strike NE-SW and NW-SE, dated at 800 Ma (Carte Geologique de la Guyane Française, 1:500 000, in press). The alkaline complexes of Cerro Impacto, Venezuela and Seis Lagos, Brazil, still undated, may also be of Neoproterozoic age.

Swarms of tholeiitic diabase dykes intruded between the end of the Triassic and the beginning of the Jurassic, with strike N-S and NW-SE are very prevalent in the northeastern region of the shield and are related to the opening of the Atlantic Ocean.

The most important mineralizations on the Guiana Shield are described below, with emphasis on gold deposits along with those of iron, manganese, tin, chrome, niobium, tantalite, zircon and diamond. Most of these deposits were enriched significantly by tropical weathering processes.

The Imataca Province

Banded iron formation units (BIFs) with important iron mineralization are associated with the granulite belt of the Imataca Complex. This complex strikes NE-SW and forms a mountainous chain at least 510 km long from the Aro River to the Orinoco Delta at the northern limit of the Guiana Shield (Fig. 5). The protolith of the complex with an age between 3.7-3.4 Ga, consisted of clastic and chemical sedimentary rocks, sub-aerial volcanics of siliceous and calc-alkaline composition, and smaller amounts of plutonic rocks. These rocks were intensely folded and submitted to metamorphism that varies in grade from that of the granulite facies (two pyroxenes) to the amphibolite facies. They consist of orthogneiss, paragneiss, granulite, charnockite and metamorphosed BIF, along with smaller quantities of manganiferous sedimentary rocks, marble, dolomite and anorthosite. Regional metamorphism, deformation and granitic intrusion occurred during the Aroense Event at about 2.8-2.7 Ga. Between 2.1-2.0 Ga, during the Transamazonian Event, they underwent metamorphism in the amphibolite and granulite facies with granitic intrusions (Sidder and Mendoza, 1995).

The banded iron formation units represent less than 1% of the rocks of the complex, and their average thickness varies from a few centimetres up to about 200 m. Several enriched BIF deposits such as Cerro Bolivar and San Isidro are amongst the largest in the world. The iron reserves before mining exceeded 1855 Mt at 63% Fe, and 11 700 Mt at about 44% Fe. The proto-ore of the BIF units consists of an oxide facies assemblage in which magnetite and hematite are the main mineral species. The iron-rich beds are interstratified with siliceous beds containing quartz and metamorphic iron-rich minerals. These deposits are similar to the iron formation units of the Superior Province, notwithstanding some units of the Algoma-type or Carajás-type may also occur. The precious metal content of these rocks is seemingly low.

The majority of the iron ore deposits strike E-W, following the principal structural trend of the complex. The Cuadilatero Ferrifero of San Isidro (San Isidro Iron Quadrilateral) contains the largest reserves known in the complex. The district is underlain by amphibole-pyroxene gneiss, granitic gneiss and amphibolite. The ore was formed as the result of chemical precipitation of volcanic exhalative origin. The average grade of the iron ore is 61% to 68%. At the Cerro Bolivar Deposit the weathered ferruginous laterite is a friable ore formed typically from very fine-grained iron formation. Hematite, magnetite and quartz are the principal minerals. Silicate minerals, mainly sodic amphibole and pyroxene are the most common mineral phases of the iron formation units. The Cerro Bolivar Deposit is hosted in a thick stratigraphic section of iron formation units (220 m) that is repeated by tight folding and by imbricated reverse faulting. Weathering was an important factor in the enrichment of the Cerro Bolivar ore body. This produced an iron oxide cap composed of ferruginous laterite consisting of grains of primary hematite and a hard porous matrix of secondary goethite. The El Pao Deposit occurs intercalated in hypersthene granulite and quartz-feldspathic gneiss. Three types of ore are present: siliceous ore (hematitic gneiss), high-grade hard massive ore, and canga. The first two ore-types consist of lamellar hematite (specularite) in which the crystals are orientated and strongly deformed.

Beds of secondarily enriched manganese are interstratified with migmatitic gneiss, amphibolite and granulite of the Imataca Complex. These rocks form part of a sequence consisting of gondite, quartz-biotite schist, amphibole schist and dolomitic marble, the thickness of which is about 500 m.

The individual manganese beds are generally less than 10 m, with along strike extensions of more than 20 km. There are arguments in favour of the non-volcanogenic sedimentary model as well as the volcanogenic model to explain the genesis of the manganese deposits (Bellizzia *et al.*,1981; Sidder and Mendoza, 1995; Gray *et al.*, 1993).

The Pastora-Barama-Mazaruni and Paramaka Provinces

Geological setting

The Guiana Shield has been proven to host large gold deposits. Taking into account that the production from alluvial deposits overlying the shield in the last century was about 150 t, it is estimated that the remaining gold resources of the shield exceed 700 t of gold. Most of the occurrences show grades of 1.5 g/t Au (Bertoni *et al.*, 1998).

The Pastora-Barama-Mazaruni and Paramaka provinces contain the most important greenstone belt gold producers in the Guiana Shield (Fig. 5). Altogether, the largest mines in these provinces contain about 20 Moz of gold. These greenstone terranes extend for about 1500 km along the Atlantic coast. In Venezuela, the Pastora-Barama-Mazaruni Province comprises the Pastora and Botanamo groups, and in its extension to northern Guyana the Barama-Mazaruni Supergroup (Barama, Cuyuni and Mazaruni groups) represent it. The Paramaka Province occurs in Suriname as the Marowijne Supergroup and the Matapi, Paramaka and Armina groups. In French Guiana it has as its equivalent the Maroni Supergroup and the Paramaka, Bonidoro and Orapu groups. In Brazil (Amapá), the Paramaka Province is represented by the Serra Lombarda Group (Ferran, 1988) and by the Tartarugalzinho gold district in its southeastern extremity. Granitic plutons, domal batholiths, gneiss and migmatite separate the units of the greenstone belts in branching synclinoria. The units of the greenstone belts in the two provinces cited above were deposited mainly in marine environments. In general, the greenstone belts of the Guiana Shield consist of: (a) a marine sequence of mafic volcanic rocks of tholeiitic composition; (b) basalt (tholeiitic to calc-alkaline in composition), andesite, dacite and rhyolite and; (c) a sequence consisting of turbiditic greywacke, volcaniclastic rocks, chemical sediments and pelite. There also occur beds of metaconglomerate, derived mainly from volcanic rocks with associated sediments. Metamorphosed manganiferous and ferruginous sediments, chert and carbonate units are also present. In the several belts there also occur sub-volcanic intrusive rocks of felsic composition. The metasediments include many varieties of tuff, volcaniclastic conglomerate, greywacke and shale units derived from the associated volcanic rocks. In Guyana, the stratigraphic thickness of the greenstone belt is estimated at between 8 to 10 km. The provenance of these metasediments is not known.

Basalt with pillow structures and showing evidence for chemical alteration and mineralogy consistent with submarine spilitization dominates the lower part of the greenstone belt sequence. In the middle part of the sequence there occurs a larger amount of porphyritic andesite, dacite, rhyolite, submarine and possibly sub-aerial lava flows, siliceous sediments and tuffaceous interflow beds. Ultramafic rocks constitute 1% to 2% of the igneous rocks of the greenstone belts of the Guiana Shield, generally forming layered mafic-ultramafic complexes (Sidder and Mendoza, 1995).

Mineralization

The rock chemistry of these sequences has not been studied systematically. The original chemical composition of the igneous rocks has been changed by weathering and hydrothermal alteration (spilitization and potassic metasomatism) as well as by regional metamorphism in the greenschist and amphibolite facies. Trends of tholeiitic and calc-alkaline differentiation are common in the volcanic rocks. The main rock-types are tholeiitic basalt, sub-alkaline and with low K content, in addition to basaltic andesite. The volcanic rocks were mainly extruded in submarine conditions, and have chemical characteristics similar to those published for modern oceanic basalt, island arc rocks and rocks associated with continental arcs. Isotope studies have revealed that the volcanic rocks were derived by mantle melting and do not contain any contribution from Archean continental crust (Sidder and Mendoza, 1995).

The gold mineralization of the Guiana Shield can be related to several epizonal environments including calc-alkaline intrusives (Omai, St. Elie, Yaou, Dorlin, Sophie, Eagle Mountain), deformed terrigenous sediments (Gross Rosebel, Camp Caiman, Regina, Changement, Esperance), metasomatic volcanites and/or intrusive rocks (Las Cristinas, Dorlin) and semi-massive sulphide mineralization (Paul Isnard, Increible, St. Elie). Most of the gold occurrences are hosted in rocks affected by ductil-ruptile deformation in the vicinity of large shear structures (Bertoni *et al.*, 1998) in which developed shear-zone hosted with low sulphide gold-quartz veining.

Native gold, pyrite and smaller amounts of tetrahedrite, chalcopyrite, bornite, molybdenite, scheelite and sphalerite are the most typical metallic minerals present in quartz veins. Carbonate (usually ankerite) in quartz veins and evidence for carbonatization >30 m into the wall rocks is common in some gold districts such as El Callao in Venezuela. In addition to the carbonate alteration, the wall rock shows intense silicification, sericitization and propylitization (replacement of wall rock minerals by epidote and chlorite) many dozens of metres from the veins (Sidder and Mendoza, 1995).

In the Serra Lombarda Group predominates mineralization of the hydrothermal vein-type hosted in gneiss with amphibolite relicts, biotite schist, BIF, and metachert, regarded as the remnants of greenstone belt sequences. In the Tartarugalzinho District, the main gold mineralization is associated with quartzite beds, banded iron formation units and schist (Carvalho *et al.*, 1995; Ferran, 1988).

Large ductile shear zones have not been described in the literature of the Guiana Shield to any extent. One of the largest shear features in the Guiana Shield is related to the so-called Sillon Nord-Guyanais (Milesi *et al.*, 1995) situated in the N of French Guiana, and associated with the Transamazonian Tectonic Event. Several important gold deposits are related to this shear zone, extending from W to

E: Regina, Tortue, Camp Caiman, Changement, Boulanger, St. Elie, St. Pierre, Paul Isnard and Guyanais in the central region of French Guiana, and representing a shear zone striking ESE-WNW (Bardoux *et al.*, 1998). This structure appears to extend to the central region of Guiana, crossing Suriname. Amongst the gold deposits occurring in the vicinity of this megascopic features the following may be cited: Yaou, Dorlin, Sophie, Repentir, Antino, Benzdorp, Omai, Salamangone and Labourrie Siboa (Yoshidome). When analyzed as a whole, it can be noted that the majority of the gold occurrences, until now discovered in the provinces in question, are found in the vicinity of large structures such as that cited above. All the magmatic and sedimentary rocks of the greenstone belts that host gold mineralization in the provinces of Pastora-Barama-Mazaruni and Paramaka underwent at least one phase of ductile deformation. Shear zones and foliation of the first phase of deformation were affected by the structures of a second phase, generally less penetrative. The second phase may be genetically related to the K'Mudku or Nickerie Event (*c.* 1.2 Ga). Thus, only the structures related to the first phase are regarded as being true Transamazonian structures, and the majority of gold occurrences described up to now seem to relate this to a phase of remobilization along the structures of the second phase (Bardoux *et al.*, 1998).

Most of the intrusive rocks together with the Paramaka and Mazaruni volcanics as well as the sedimentary rocks of Armina were deformed concomitantly by a phase of intense ductile deformation. This event has been dated at about 1.99 Ga at Omai and St. Elie, setting the approximate absolute age of the Transamazonian Tectonic Event throughout the Guiana Shield (Lafrance *et al.*, 1999). U/Pb ages obtained on rocks from intrusive bodies from different parts of the Pastora-Barama-Mazaruni and Paramaka provinces suggest that there occurred at least three distinct intrusive events at 2.154 Ga (Las Cristinas), 2.125 Ga (St. Elie) and 2.09 Ga (Omai) (Fig. 5). When examined individually, it can be noted that the majority of these intrusive bodies underwent at least one phase of intense deformation that in many cases is synchronous with the mineralization. Data from Omai and St. Elie show that the mineralization occurred more or less contemporaneously in these deposits at about 1.99 Ga. It also showed that the gold was trapped several dozens of millions of years after the intrusive event (Lafrance *et al.*, 1999).

In summary, the characteristics of the gold occurrences in the Pastora-Barama-Mazaruni and Paramaka provinces are as follows: (a) the host rocks are variable, but volcanic rocks predominate; (b) structural control is the norm, although the types and styles vary; (c) the proximity to intrusions seems to be important, but not always so; (d) most of the deposits may be related to a stronger phase of deformation; (e) the majority of the gold occurrences observed are hosted within or very close to quartz veining, which is syn-tectonic to late-tectonic; (f) most of the occurrences are also hosted in discrete zones or regional shear zones; (g) the gold is generally associated with the sulphide minerals (pyrite, chalcopyrite and pyrrhotite), especially pyrite (Bardoux *et al.*, 1998).

The largest gold deposits in the Pastora-Barama-Mazaruni Province are those at Omai (4.2 Moz) in Guyana, and a set of deposits including Las Cristinas (8.6 Moz), El Callao, Lo Increible and Botanamo in Venezuela. Besides gold, the Las Cristinas deposits contain exploitable reserves of copper (Minérios & Minerales, ed. 225, 1998). In the Paramaka Province the most important deposit is that at Gross Rosebel (2.4 Moz) in Surinama; Paul Isnard (2.2 Moz), Camp Caiman (1.1 Moz), Yaou (0.8 Moz), Dorlin (0.35 Moz) and St. Elie in French Guiana; and Salamongone/Laborrie Siboa (0.35 Moz) in Brazil. The Omai Deposit in Guyana is the largest gold deposit known to date on the Guiana Shield, and one of the largest in South America.

The Vila Nova Province

The Vila Nova province is situated on the southeastern margin of the Guiana Shield to the W of Macapá, Brazil and extends to the international frontier with French Guiana. This province is underlain by volcano-sedimentary sequences of the greenstone belt type (Vila Nova Metamorphic Suite or Vila Nova Group) enclosed in medium to high-grade metamorphic complexes (Ananai and Guianense metamorphic suites). On the other hand, the metamorphites of medium to high-grade may have been derived from the same units that appear as lower grade rocks of the adjacent greenstone belts (João *et al.*, 1978; João and Marinho, 1982; Lima *et al.*, 1974; Gibbs and Barron, 1993). Radiometric data show intense reworking during the Transamazonian Tectonic Event (Tassinari and Macambira, 1999), involving older Archean rocks of high metamorphic grade. The Vila Nova Group occurs as discontinuous elongated and narrow belts that strike NW-SW, forming metamorphic belts of low to medium grade and consisting of broad folds with vergence to the NE. This group is considered to be an integral part of the various volcano-sedimentary sequences of the Guiana Shield dated between 2.25-2.10 Ga. The stratigraphy of the Vila Nova Group is best defined in the area of Serra do Navio. At the base there is a thick sequence of ortho-amphibolite (Jornal Formation) in contact with the Ananaí Suite (Scarpelli, 1966). The ortho-amphibolite is overlain by a parametamorphic sequence consisting of aluminous schist containing lenses of manganiferous proto-ore and beds of quartzite (Serra do Navio Formation). The schist sequence occurs in the quartzose, biotitic and graphitic facies. In the quartzose facies there occur lenses of calc-silicate marble, and in the graphitic facies zones of manganiferous marble with rhodochrosite can be observed. The absence of felsic volcanic rocks in the Vila Nova Province is a characteristic that differentiates this province from the Paramaka and Pastora-Barama-Mazaruni provinces.

In the Serra do Ipitinga, the southern part of the Vila Nova Province, the Vila Nova Group consists essentially of mafic-ultramafic basal metavolcanic rocks, and smaller amounts of metaplutonic rocks, rocks containing cordierite-anthophyllite and quartz-chlorite. Chemical sediments consisting of BIF (oxide and silicate facies), in addition to continental clastic metasediments (quartzite, metapelite and metagreywacke) overlie these rocks. The basal metavolcanic unit represents oceanic tholeiite with a subordinate komatiitic chemical lineage. In the Serra do Ipitinga occur three types of deposits: (a) sulphide mineralization of the syn-depositional hydrothermal volcanogenic type hosted, preferentially, in the quartz-chlorite rocks, and more rarely, in low grade metamorphites (pyrrhotite-pyrite-chalcopyrite and subordinately sphalerite with Au and Ag associated); (b) lode-type deposits in sheared quartz veins containing chalcopyrite, pyrite and covellite; (c) rocks altered by supergene processes such as gossan and laterite (Faraco, 1990, 1997; Faraco and McReath, 1998).

In the region of the rivers Santa Maria and Cupixi there predominate quartzite beds and sericite schist, intercalated with BIF units, and subordinately, metaconglomerate. This unit overlies the Bacuri mafic-ultramafic complex, in which there occurs chromite mineralization (Faraco and Carvalho, 1994).

Furthermore, in the Vila Nova Province occur intrusive rocks of the Mapuera Suite (biotite-alkaligranite and riebeckite-granite) and the Falsino Suite (granodiorite), as well as several bodies of the Mapari alkaline intrusive having Paleoproterozoic to Mesoproterozoic ages (Rb/Sr). On the southwestern margin of the Vila Nova Province the Mapari Intrusive Suite consists of two alkaline-carbonatite complexes (Maecuru and Serra do Maracanaí), with titanium and phosphate mineralization.

From the metallogenic point of view, the Vila Nova Province shows a widespread distribution of gold occurrences. Including the Amapari gold deposit, there are about 100 gold known occurrences. Also of note is the Serra do Navio manganese deposit, and the chromite at Bacuri, also known as Igarapé do Breu. Gold mining has been carried out at *garimpos*, generally in colluvial and placer deposits. In general, the primary gold mineralization is of the hydrothermal vein-type, associated with schist, quartzite and amphibolite. Gold is also found in metaconglomerate. The Mapuera Granite contains disseminated cassiterite; and gold, columbite and tantalite in quartz veins (Carvalho *et al.*, 1995; Faraco and Carvalho, 1994).

The Bacuri (Igarapé do Breu) Chromite Deposit

The chromite deposits are associated with the Bacuri mafic-ultramafic complex (Matos *et al.*, 1992) situated to the W of Macapá, which is intrusive into granite-gneiss medium to high-grade metamorphic terranes, associated with the Vila Nova Group. According to Spier and Ferreira Filho (1999), the Bacuri mafic-ultramafic complex that includes amphibolite, serpentinite, tremolite, and chromitite was deformed and metamorphosed in the amphibolite facies. These rocks include amphibolite, serpentinite, tremolite, and chromitite. They have a stratiform differentiated character defined in function of the magmatic layering of the cumulate and the chemical characteristics of the textures observed. The principal layer of massive chromitite, having an average thickness of 12 m, is situated in the interface between the mafic and ultramafic rocks. Thinner layers of massive and disseminated chromite (up to 3 m), are observed intercalated in the ultramafic rocks. The chromite crystals are euhedral, and their size is uniform (average 0.2 mm). The intense lateritic weathering exceeds depths of 120 m in the topographically elevated localities, permitting the classification of the ore into three categories: lateritic ore, very hard and cemented by iron oxide and hydroxide; friable ore cemented by clay minerals; compact ore cemented by tremolite, chlorite and rarely by orthopyroxene and olivine.

This complex hosts 11 deposits of stratiform chromite with before-mining reserves exceeding 9 Mt of chromitite, of which 2 Mt were mined between 1989 and 1997.

The Serra do Navio Manganese Deposit

The manganese deposits of Serra do Navio have a strike length of 10 km that trends N30°W. They are associated with the volcano-sedimentary sequence of the Vila Nova Group. This sequence has been highly folded, sheared and metamorphosed in the amphibolite facies. At the base occurs a thick sequence of mafic metavolcanic rocks (amphibolite), overlain by a predominantly schistose unit (biotite schist, graphite schist, quartz schist) intercalated with manganiferous layers and quartzite beds in the upper part (Rodrigues *et al.*, 1986).

The manganiferous zones that constitute the proto-ore of the queluzite-type, are intercalated with graphitic schist, consisting preferentially of rhodochrosite amounting to between 50% to 90% of the rock. In the more impure zones the manganese silicate minerals such as tefroite, spessartite and rodonite may predominate to form true gondite. The Mn grade of the proto-ore varies between 19% and 36%. Some sulphide minerals such as pyrrhotite, molybdenite, chalcopyrite and galena are found associated in small amounts in the proto-ore. On account of the equatorial climate, the laterite weathering profile may reach a depth of 100 m, causing the transformation of the carbonate and silicate minerals in the manganese-rich rock to oxide such as cryptomelane, pyrolusite and manganite, resulting an oxide ore richer in manganese up to 56% Mn.

The open-pit working began in 1959 in the oxide ore and continued for decades until 1997 at an output of 52 000 tpa, or a total production over the period amounting to 50 to 60 Mt MnO_2. During the final years of operation, the carbonate-rich proto-ore was mined at about 150 000 tpa, equivalent to a total production of about 3 Mt of manganese ore at an average grade of 35% to 38% Mn.

The Amapari Gold Deposit

The Amapari gold deposit was found recently (AngloGold) some 12 km to the E of Serra do Navio (Borges, 1999). The deposit is hosted in the volcano-sedimentary sequence of the Vila Nova Group, and it is regionally related to a transcurrent shear zone that underwent intense hydrothermal alteration. From the base to the top, there occur the following units: granite-gneiss basement; a volcanic unit (metabasic rocks and amphibolite); a chemical sedimentary unit composed of carbonate rocks, calc-silicate and oxide, silicate and oxide-silicate facies; banded iron formation units, silicate and oxide-silicate; a clastic-chemical sedimentary unit consisting of quartz-amphibole schist and amphibole schist; a unit consisting of pelitic clastic sediments including muscovite-quartzite and micaschist.

Throughout the area occur pegmatite intrusions in the form of elongate bodies, the thickness of which varies from some metres to more than 100 m. The pegmatite is syn-

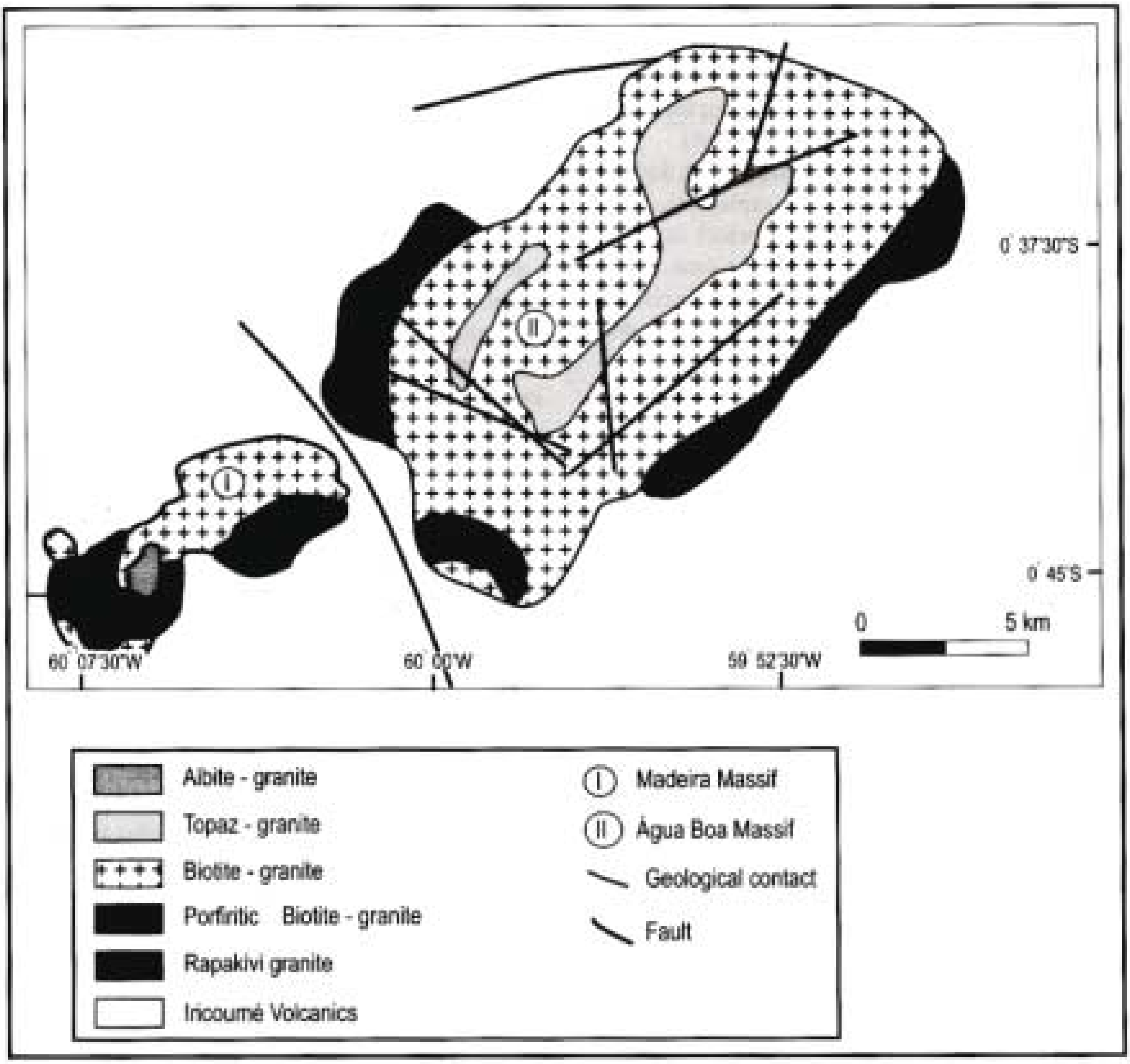


FIGURE 6 - Geological map of Pitinga Mine (modfied after Lenharo, 1998). For situation see Figure 5

tectonic and generally associated with shear zones with strike similar to the wall or country rocks.

The Amapari gold deposit has been classified as a contact metasomatic deposit or a skarn-type deposit (Meinert, 1988 *apud* Borges, 1999) in which the hydrothermalism was controlled structurally by a sinistral, high-angled transcurrent shear zone striking NNW-SSE. Amapari represents a gold deposit that resulted from a combination of factors: lithology (reactive rocks) and favourable structures (shear zones), the latter being the more important. The interaction between the rock-types and structure favoured the gold concentration, both by contact metasomatism as well as by hydrothermal solutions that percolated along the shear zone resulting in intense hydrothermal alteration (principally silicification, sulphidization and carbonatization) in reactive rocks such as BIF units, amphibolite, carbonate beds and calc-silicate. The highest gold grades occur preferentially in the BIF units (oxide, oxide-silicate and silicate facies), with intense hydrothermal alteration manifest as silicification and sulphidization (pyrrhotite and pyrite). The gold-bearing rocks have a brecciated aspect and intense transposition caused by shearing. There also occur carbonate rocks (calcareous marble), amphibolite and calc-silicate. These rocks are quite reactive in the presence of mineralizing hydrothermal and metasomatic fluids that percolated along the shear zone, concentrating the gold in favourable structures in the host rocks. The gold seems to be related to the structure of the sulphide minerals (pyrrhotite and pyrite). Chalcopyrite, sphalerite, arsenopyrite, galena and marcasite also occur in small amounts. However, with the exception of chalcopyrite, these minerals do not seem to be directly related to the gold mineralization.

The mineralization occurs in the zone of weathering as well as in the sulphide-bearing bedrock, where it follows the shear planes at depth. The primary mineralization occurs along a N-S strike-length some 7 km long, associated with zones of hydrothermal alteration in skarn. The weathering profile attains depths exceeding 100 m.

The evidence for considering a part of this deposit as a skarn-type gold deposit is based on the mineralogy, geological setting and the sulphide mineralogy, which may be compared to that of gold-skarns worldwide. The typical skarn minerals are garnet, pyroxene (diopside-hedenbergite), vesuvianite, apatite, titanite, actinolite, epidote and hornblende. There also occurs pyroxene rich in manganese. The sulphide species are pyrite and pyrrhotite. The origin of the mineralizing fluids may be explained in terms of the hypothesis that the deposit came about as the result of metasomatic fluids laden with gold and other elements that were channeled along fractures opened by the pegmatite intrusions. Seeing that the pegmatite is syn-tectonic, it is probable that the mineralizing fluids were also channeled along the shear zone (Borges, 1999).

However, the hypothesis that stands up best in the light of field observations and that best explains the gold concentration of the deposit is that which involves hydrothermal solutions associated with the shear zone.

The mineable reserves of colluvial and oxidized material, for a cut-off grade of 2.13 g/t Au, are about 30 000 kg/Au (*c.* 1.0 Moz Au). There is a considerable potential for primary gold mineralization as sulphide ore that is still being evaluated. Regional geochemical anomalies show anomalies for copper, lead and zinc (Borges, 1999).

The Pitinga Tin District

In the southern part of the Guiana Shield there is important tin, rare metals (Zr, Ta, Y and REE) and cryolite (Na_3AlF_6) mineralization, associated with the granitic massifs of Água Boa and Madeira at Pitinga (Figs. 5 and 6). These granitic rocks are considered to be anorogenic, intraplate and to have originated at high crustal levels. The age of the intrusion into the Iricoumé volcanic rocks is given as 1.8 Ga (Lenharo, 1998). Two types of tin mineralization have been defined:

a) Tin mineralization associated with greisen of the Água Boa Massif. According to Daoud (1988) and Lenharo (1998), the Água Boa Massif is a multiple intrusion consisting of three distinct phases: Rapakivi Granite, Biotite Granite and Topaz Granite. The tin mineralization is associated with post-magmatic processes of hydrothermal alteration in the form of vertical veining of mica-topaz-quartz greisen containing cassiterite, opaque minerals and tourmaline in fractures striking N50°W. The associated minerals are allanite, opaque minerals, zircon, fluorite, siderite, beryl, and sulphide minerals such as sphalerite, pyrite and chalcopyrite. Intense albitization of the granitic wallrock preceded the greisenization.

b) Tin mineralization associated with albite granite of the Madeira Massif. According to Daoud (1988) and Lenharo (1998), the Madeira Massif shows three distinct facies: Rapakivi Granite, Biotite Granite and Albite Granite. The tin mineralization associated with albite-granite is the disseminated type, being composed of cassiterite, zircon, columbite-tantalite, pyrochlore, xenotime and cryolite. The fresh rock contains: 0.176% Sn; 0.223% Nb_2O_5; 0.028% Ta_2O_5; 0.030% U_3O_8; 0.80% ZrO_2. This disseminated mineralization is enriched in the weathered mantle that is about 30 m thick and which is depleted with respect to cryolite. At a depth of about 150 m, the cryolite forms two massive bodies at the nucleus of the albite granite. In 1997, the production was *c.* 11 693 t Sn from 21 700 t of cassiterite concentrate with 53.88% contained Sn. The perspectives for the next fifteen years are to mine 13 M tpa of ore corresponding to the equivalent of 13 000 t/Sn, in adition to 800 t of columbite concentrate with 35% Nb_2O_5 and 3,5% TaO_2.

Diamond in the Guiana Shield

Three distinct diamond-generating epochs may be defined on the Guiana Shield: Dachine, Roraima and Quebrada Grande.

Dachine: This is a new type of occurrence that has recently been described in the Dachine area of French Guiana. The host rock is a volcaniclastic komatiite, an unusual volcanic rock found in the Paleoproterozoic Inini greenstone belt of the Paramaka Province. In the Inini greenstone belt there predominates calc-alkaline andesite and rhyolite and immature sediments. These rocks are intruded by granitoid plutons including tonalite and trondhjemite, suggesting an island arc setting. The ultramafic rocks that host the Dachine diamonds form part of a volcanic sequence in which most of the rocks have been transformed into albite-carbonate-chlorite-talc schist. Primary volcanic texture have been well-preserved locally. The diamond population consists mainly of micro-diamonds. The largest stone recovered was c. 4.6 mm in diameter. The discovery of diamond in ultramafic schist, such as the Dachine Schist, provides an explaination for diamond occurrences from unknown sources in other parts of the Guiana Shield, in addition to elsewhere on the South American Platform (Capdevila *et al.*, 1999).

Roraima: With the notable exception of the Quebrada Grande area, most of the known diamond placers in the Guiana Shield are in areas underlain by the Roraima Group, or are in areas downstream from exposures of this group. Diamond placers associated with the rocks of the Roraima Group form the second group of occurrences on the Guiana Shield. Although no diamond occurrence within the Roraima Group has ever been found, there is a pronounced spatial association between the alluvial diamond deposits and the outcrops of the Roraima Group. The most important occurrence is San Salvador de Paúl. The total historical production exceeds 2 M carats. The monthly production is estimated at 2000 carats. The high percentage of gem-quality diamonds and the absence of other minerals typical of a kimberlitic association that are more resistant to transport suggests a distal source or more than one sedimentary cycle or both (Gray and Orris, 1993). Another important locality for diamond placer deposits is Icabarú. Icabarú occurs at the southern margin of the Roraima Group, near to the frontier of Venezuela and Brazil (Brooks *et al.*, 1995). Likewise, in this area the diamonds are mined from conglomerate derived from the Roraima Group. The main alluvial diamond mining areas occur around or in the interior of the Pakaraima Mountains in Venezuela, Guyana and Brazil. These alluvial diamonds are generally regarded has having been derived from Pre-Roraima kimberlite and preserved in the conglomerate of the Roraima Group. In Venezuela, most of the areas that have been worked are situated in the drainage of the Caroni, Paragua and Icabarú in the eastern part of the State of Bolivar.

Quebrada Grande: The only diamondiferous kimberlite in South America occurs at Quebrada Grande, a tributary of the Guaniamo River, in the State of Bolivar, Venezuela. Placer diamond deposits were discovered in the Guaniamo region in 1968, and since then some 20 to 25 M carats have been recovered, including stones exceeding

40 and 60 carats. In 1982, highly diamondiferous kimberlitic dykes and sills were discovered. These intrusives have been dated at 710 Ma (Kaminsky *et al.*, 1998; Channer *et al.*, 1998). The emplacement of this kimberlite represents the youngest Precambrian event thus far observed on the Guiana Shield. The kimberlite cuts biotite-lamprophyre dykes dated at *c.* 850 Ma (Nixon *et al.*, 1992). The diamond production from this kimberlite by formal mining methods is expected to be 450 000 carats per year. It should be noted that the current production from the entire Guiana Shield is estimated at *c.*250 000 carats per year (L. A. Bizzi, personal communication)

The Carbonatites of Cerro Impacto and Seis Lagos

The Cerro Impacto is a structural feature in the State of Bolivar, Venezuela, that was first observed in radar images. Preliminary studies showed anomalous values for niobium, thorium, barium and other metals besides REE. Although fresh rock has still not been found, the mineralogy of the products of chemical weathering and leaching suggests that the original composition of the proto-ore was a carbonatite (gorceixite, goyasite, florencite, bastnaesite, monazite etc.). The carbonatite is associated with a ring structure, oval on shape, and approximately 10 km in diameter. The weathered cap that is at least 200 m thick does not contain any fragments of the original carbonatite rock. The laterite is enriched in Fe, Mn, Al, Ba, Th, Nb, REE (Ce, La, Nd), Ti, Zn, Pb and other elements.

This body has been emplaced near the intersection of large NE-SW and NW-SE fractures. These fractures may be related to those into which was intruded the Quebrada Grande Kimberlite. The age of the carbonatite is not known. A Proterozoic age is suggested by a possible relationship with the Quebrada Grande Kimberlite at 710 Ma (Bellizzia *et al.*, 1981; Sidder, 1995, Sidder and Mendoza, 1995; Channer *et al.*, 1998).

The Seis Lagos Carbonatite is situated in the headwaters of the Negro River, Brazil, near the frontier with Venezuela, and it consists of three alkaline-carbonatite pipes, mineralized with respect to niobium. In likemanner to the Cerro Impacto the age of this intrusion is not known. The very intense laterite weathering resulted in the development of a weathered cap exceeding 200 m thick in which the pyrochlore was destroyed, and the primary niobium minerals have been neomorphosed to rutile and niobiferous brookite. The reserves at Seis Lagos have been estimated at 2.898 billion t of ore at 2.81% Nb_2O_5 (Justo and Souza, 1986).

The Central Brazil Shield

With the exception of the Maroni-Itacaiunas Belt that is for all practical purposes restricted to the Guiana Shield, the Central Brazil Shield clearly shows a geotectonic zoning developed from E to W around an Archean nucleus by the accretion of successive magmatic arcs between 1.95 and 1.5 Ga, resulting in the formation of a vast continental juvenile crust. From 1.5 Ga to 1.0 Ga, the evolution of this southwestern part of the shield occurred in an ensialic environment.

The nature and distribution of the mineral deposits found on the Central Brazil Shield reflect this geotectonic evolution permitting the definition of several mineral provinces as follows:

- The Carajás Province with gold deposits in Rio Maria granite-greenstone terrane and with deposits of Fe, Mn, Au, Cu, Zn, Cr and Ni associated with the volcano-sedimentary sequences of the Grão Pará, Igarapé Bahia, Pojuca, Salobo groups, and the sediments of the Águas Claras Group;
- The Tapajós and Alta Floresta provinces with gold deposits mainly associated with intrusive granite of the Maloquinha, Serrinha do Matupá and Teles Pires types as well as others, in addition to the Cu-Pb-Zn mineralization related to the sedimentary (volcanic) cover of the Beneficente Group;
- The Alto Jauru Cu-Au District with deposits associated with volcano-sedimentary sequences;
- The Alto Guaporé Province with gold deposits formed during the Sunsas Event and Ni deposits associated with post-orogenic mafic-ultramafic complexes such as that of Rincón del Tigre in Bolivia;
- The Rondônia Province with tin deposits associated with youngest anorogenic granites.

The Carajás Mineral Province

The Carajás Mineral Province (Fig. 7a, b) (Santos, 1983) consists of two sets of Archean rocks of distinct age: the older Rio Maria region, and the younger Serra dos Carajás region.

The Rio Maria region is situated in the SE of the State of Pará and occupies the oldest part of the Archean Amazonian nucleus. These rocks have been assigned to the Rio Maria Granite Greenstone Complex, the evolution of which occurred in an interval between 3.0 and 2.8 Ga (Souza *et al.*, 1996, 2000). The volcano-sedimentary sequences (2.98 to 2.90 Ga) occur in series of irregular belts (Andorinhas, Gradaús, Inajá, Rio Novo and Sapucaia) known as the Andorinhas Supergroup, and are intruded by granodiorite of the Rio Maria type at 2.87 Ga (Pimentel and Machado, 1984; Macambira and Lancelot, 1992). At 2.87 Ga the region underwent an important tectono-metamorphic event, correlative with the Xingu Granite-Gneiss Complex.

The consolidation of sialic crust is taken to have occurred following the development and deformation of the Rio Maria greenstone belt sequence and its associated plutonism. On this sialic crust developed the volcano-sedimentary sequence of the Grão Pará Group, dated at 2.76 Ga (Machado *et al.*, 1991; Gibbs *et al.*, 1986, Wirth *et al.*, 1986). This sequence is correlative with the Igarapé Bahia, Pojuca and Salobo groups, which are overlain by the sediments of the Águas Claras Group, also of Archean age (>2.645 Ga). Several generations of mafic and ultramafic intrusions dated at 2.67 Ga (Machado *et al.*, 1991) and 2.645 Ga (Dias *et al.*, 1996) as well as granitic bodies assigned to the Sossego/Cristalino (2.74Ga), and Plaquê/Old Salobo (2.57 Ga) suites occur in the province (Dall'Agnol *et al.*, 1987). The deformation and the regional metamorphism associated with regional shear zones occurred between 2.58 and 2.5 Ga (Machado *et al.*, 1991; Pinheiro and Holdsworth, 1997), constraining the final stabilization of the area. At the

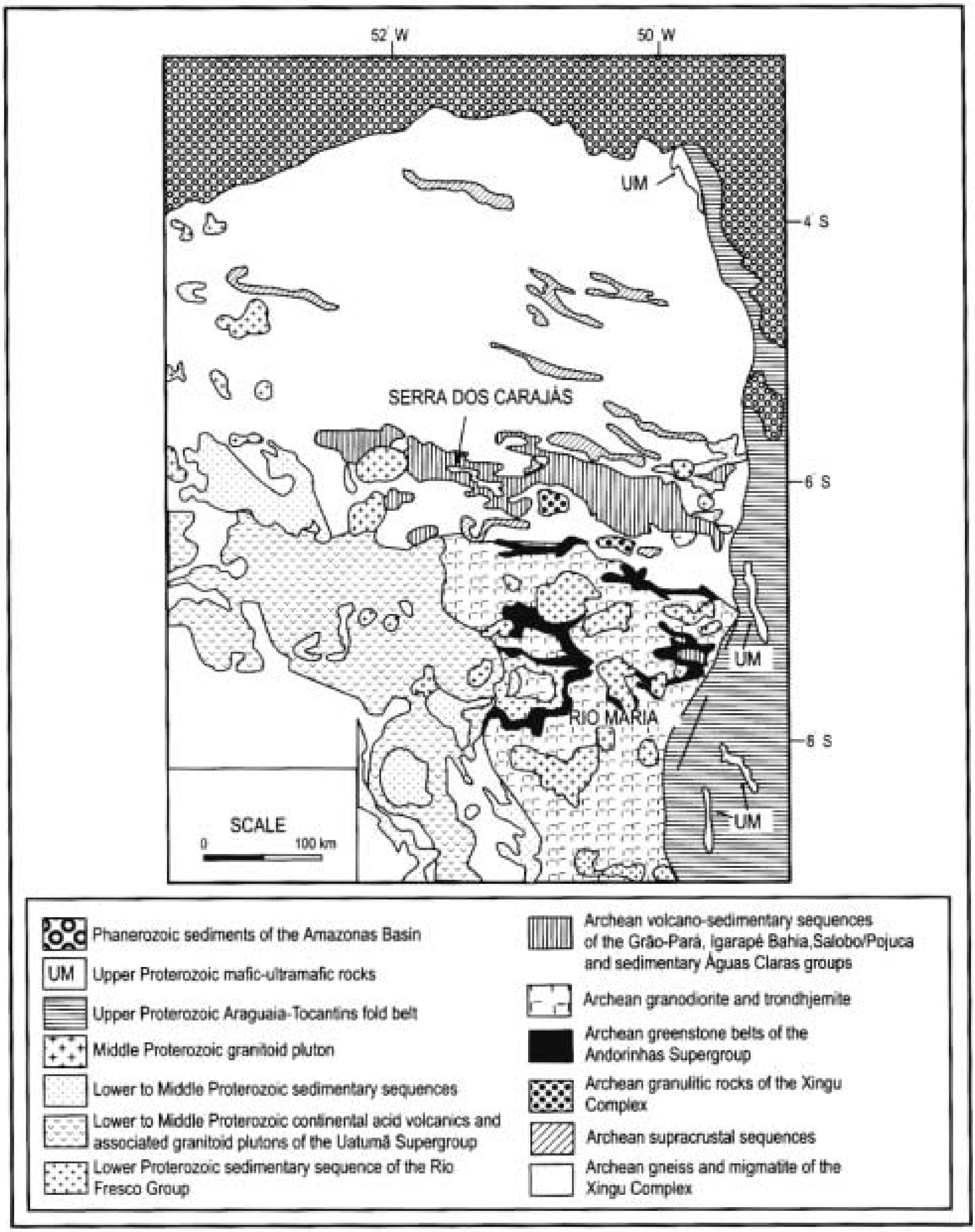
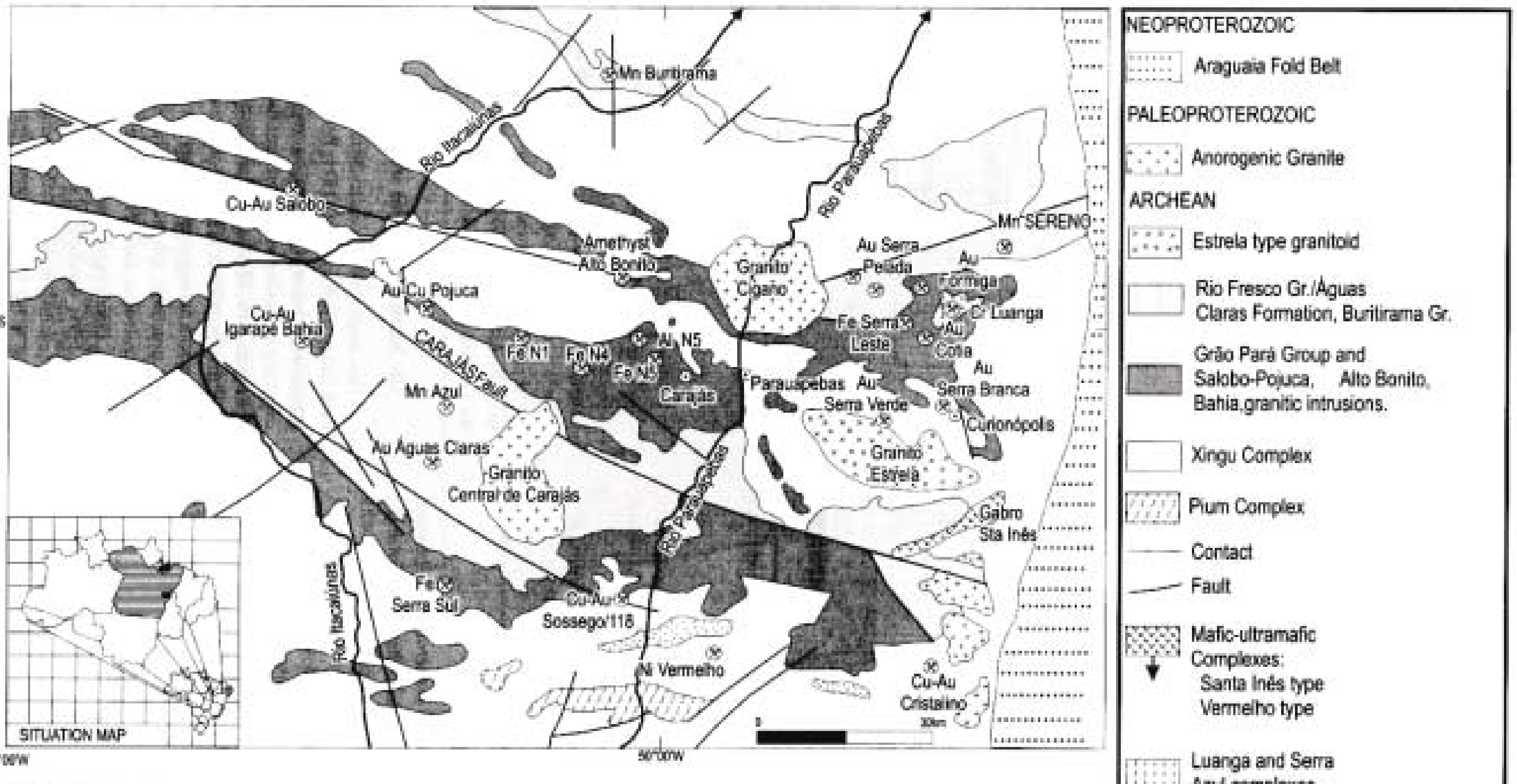


FIGURE 7a - Geological map of the eastern sector of the Amazonian Craton (modified after DOCEGEO, 1988; Araújo and Nilson, 1988).



RADIOMETRIC AGES

Anorogenic Granites		**Águas Claras Formation**	2.681+0.005 Ga
Central de Carajás	1.888+0.002 Ga	**Grão Pará Group**	
Pojuca	1.874+0.002 Ga	Carajás Fm.	2.740+0.008 Ga
Cigano	1.883+0.002 Ga	Parauapebas Fm.	2.759+0.002 Ga
Young Salobo	1.880+0.080 Ga	Luanga Complex	2.763+0.006 Ga
Orogenic Granites		**Salobo-Pojuca Group**	2.732-2.742 Ga
Estrela	2.527+0.034 Ga	**Xingu Complex**	2.859+0.002 Ga
Itacaiúnas	2.560+0.030 Ga	**Pium Complex**	>3.0 Ga
Old Salobo	2.573+0.002 Ga		

FIGURE 7b - Geological sketch map of the Carajás Region (modfied after DOCEGEO, 1988).

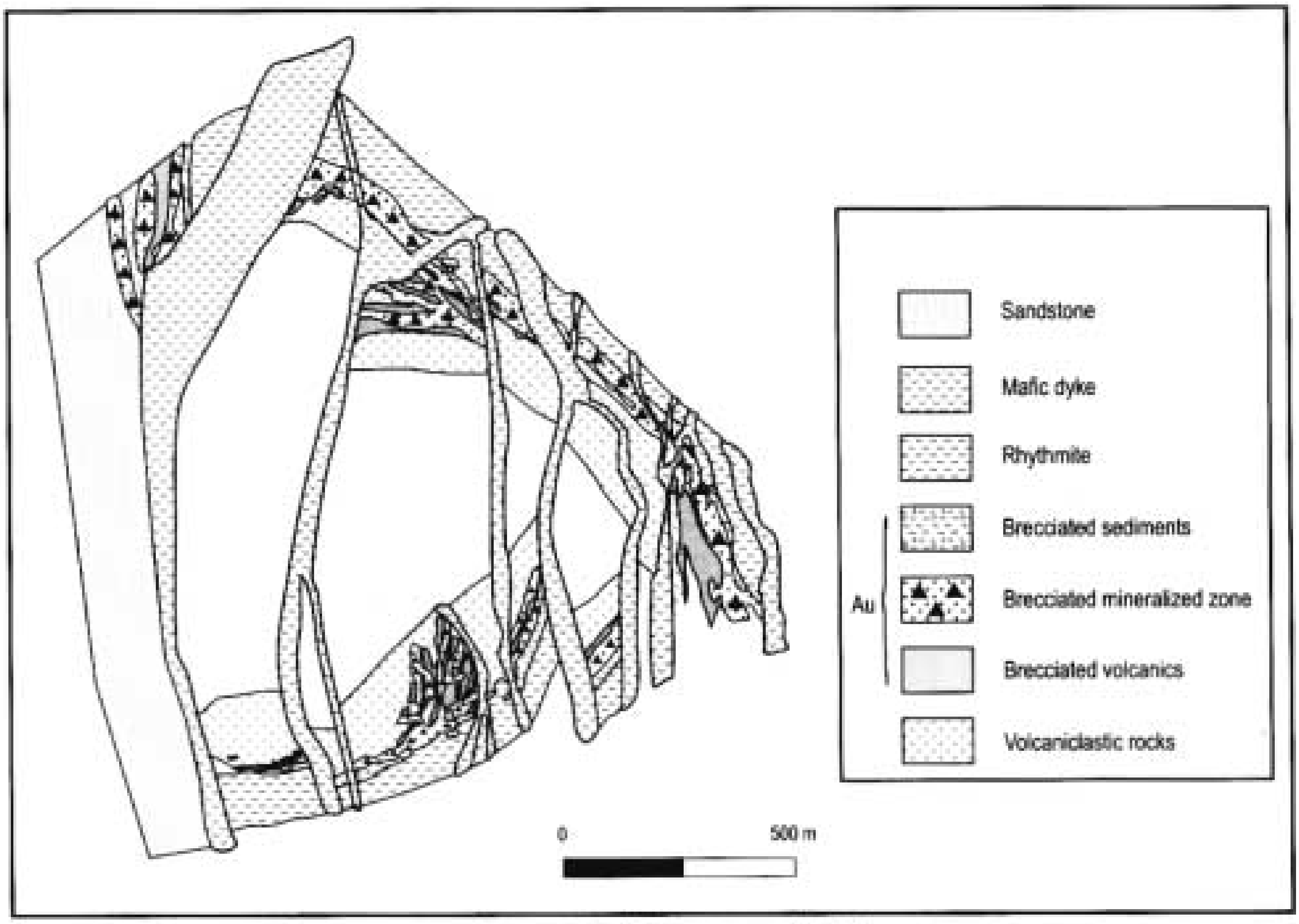


FIGURE 8a - Geological sketch map of Cu-Au Igarapé Bahia Deposit (modified after Soares et al., 1999).

beginning of the middle Proterozoic (1.88 Ga), occurred in the region several anorogenic intrusions such as the Central Carajás Granite/Young Salobo Suite.

These units are associated with Cr, Fe, Cu-Au, Mn, Au and W mineralization that define the Carajás polymetallic province.

Lode-type Gold Deposits

Several small deposits and numerous gold showings are found in pyrite-rich quartz veins that developed as the result of intense hydrothermal alteration associated with regional shear zones that cut the rocks of the volcano-sedimentary sequence. The best known deposits are the Diadema in the Sapucaia greenstone belt (Oliveira *et al.*, 1995; Oliveira and Leonardos, 1990; Nascimento and Biagini, 1988) and Babaçu/Lagoa Sêca in the Andorinhas greenstone belt (Huhn, 1991; 1992; Souza *et al.*, 1990; Silva and Cordeiro, 1988).

Au-Cu-Bi-Mo Deposits of the Porphyritic Lode-type

The Cumaru gold deposit (Leonardos *et al.*, 1991; Santos, 1995; Santos *et al.*, 1998) is associated with a stockwork of quartz veins and veinlets, rich in pyrite, and related to a late-tectonic granodiorite intrusion dated at 2.87 Ga (Lafon and Scheller, 1994; Lafon and Macambira, 1990). The association of the metals Au-Cu-Bi-Mo, and the geochemical and isotope data, in addition to the fluid inclusion studies, led these authors to propose a mixed origin for the gold mineralization, classifying the deposit as a porphyritic lode-type exhibiting both magmatic and structural controls: The Cumaru Granodiorite is defined as I-type calc-alkaline intrusion of volcanic arc affiliation, similar to porphyry systems, that was affected by metamorphic aquo-carbonic fluids, which have circulated in the shear zone on the southern flank of the Gradaús greenstone belt.

The Serra dos Carajás Iron Ore Deposit

The Carajás Formation is an intermediate unit of the Grão Pará Group (Tolbert *et al.*, 1971; Beisiegel *et al.*, 1973) intercalated between two mafic volcanic units, and consisting of jaspilite which represents the proto-ore of the iron ore deposits. This jaspilite has as its main characteristic the alternation of bands rich in hematite/magnetite and bands of jasper, with numerous sedimentary structures such as cut-and-fill structures, spherulites, intraformational breccia, slump structures, and dehydration nodules and veins (Meirelles, 1986; Meirelles and Dardenne, 1993; Macambira and Silva, 1995). Geochemical studies on the jaspilite showed a positive anomaly for Eu and intermediate Cu and Zn values in relation to the values for BIF of the Algoma and Lake Superior types. The REE spectra are very similar to those of the mafic volcanic wall rocks. The set of geological and geochemical data indicate a volcano-sedimentary origin in a rift environment for the formation of the jaspilite in the Carajás Formation.

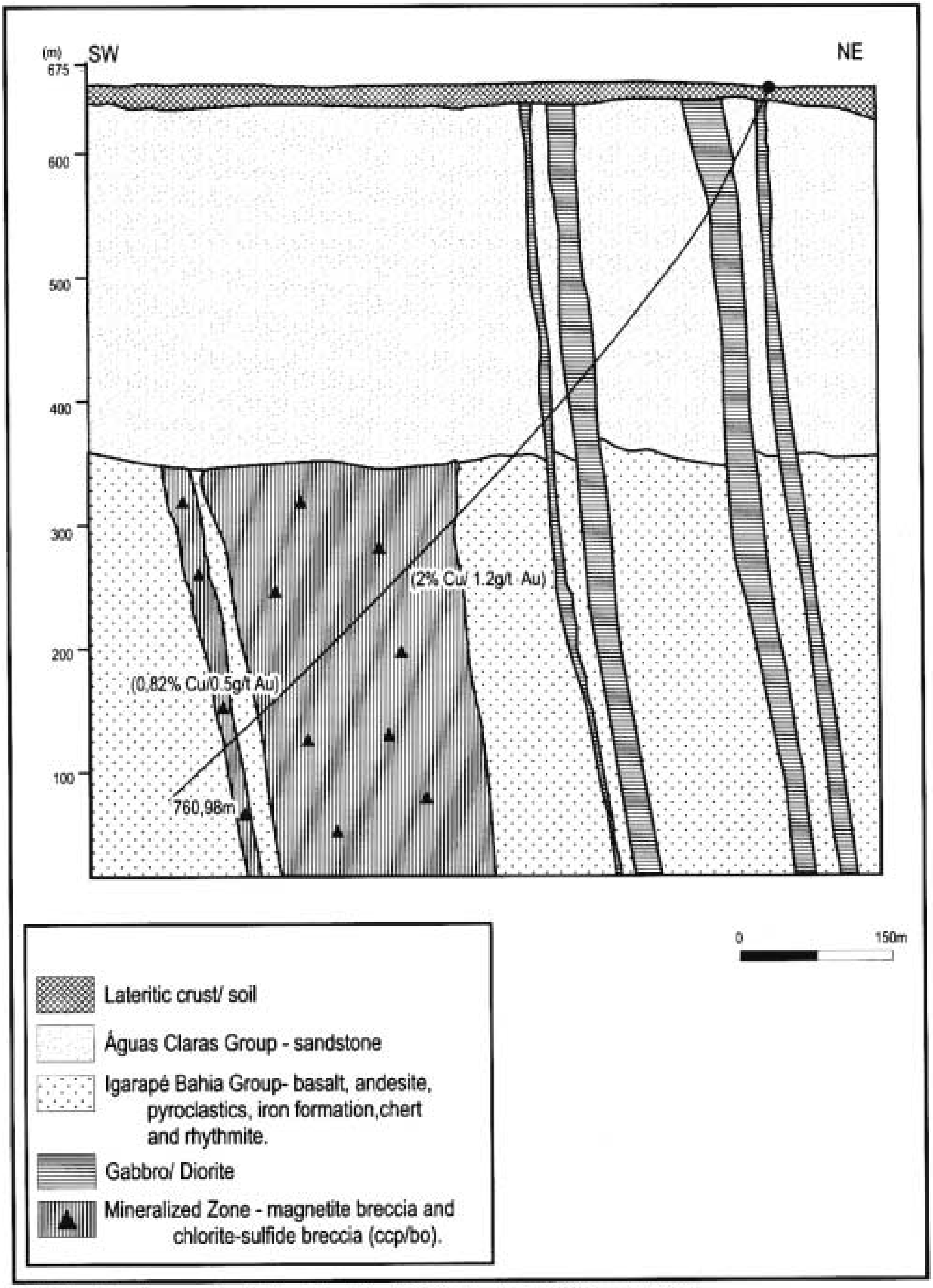


FIGURE 8b - Schematic cross-section of the Alemão Deposit (modified after Barreira et al., 1999; Soares et al., 1999)

These iron formation units cannot be classified either as the Algoma-type or as Superior-type. Moreover, they seem to be a unique type: the Carajás-type. On one hand, the strict relationship between the jaspilite of the Carajás region with volcanic rocks permits a comparison with the Algoma-type. On the other hand, the iron formation units occur only in the oxide facies suggesting similarities with the depositional environment of the Superior-type (Meirelles, 1986; Hoppe and Schobbenhaus, 1990; Meirelles and Dardenne, 1993).

The iron ore presently being worked in open pits at the huge Carajás Deposit came about as a result of laterite weathering that caused the supergene leaching of the silica in the jaspilite and the residual concentration of hematite (Costa, 1997). The reserves of the Serra dos Carajás have been estimated at 18 billion t of ore at 60% to 67% Fe. At the N4E area the mineable reserves are about 1.251 billion t of ore at an average grade of 60.9%Fe (Coelho, 1986). In 1998, the production was about 54 Mt of iron ore.

The controversy over the Carajás iron deposit involves the geotectonic context that preceded the development of the rift; (a) the basaltic volcanism is intraplate and of the Paraná Basin-type (Wirth *et al.*, 1986; Gibbs *et al.*, 1986; Lindenmayer, 1998) and the high-K values are related to the percolation of hydrothermal fluids (Lindenmayer *et al.*, 1998); (b) the basaltic volcanism associated with the jaspilite is defined as high-K with shoshonitic affinities (Meirelles, 1986; Meirelles and Dardenne, 1991; Dardenne *et al.*, 1988; Teixeira, 1994; Teixeira and Eggler, 1994) that developed as a mature magmatic arc, dominated by subduction mechanisms.

The Luanga Chrome Deposit

The stratiform bodies of massive and disseminated chromitite are associated with the orthopyroxenite of the mafic-ultramafic layered and differentiated complex of Luanga, intruded at 2.76 Ga (Machado *et al.*, 1991) at the base of the Rio Novo greenstone belt, thus being contemporaneous with the volcanism of the Grão Pará Group. This complex of mafic tholeiitic association is composed of a set of cumulate rocks, beginning with dunite and peridotite at the base, grading to orthopyroxenite with zones of chromitite, and transitioning to norite and leuconorite in the upper part (Hirata *et al.*, 1982; Medeiros and Meireles, 1985; DOCEGEO, 1988; Suita, 1998; Suita and Nilson, 1991). According to Suita (1996), the massive chromitite is enriched in PGE, with grades attaining 3.2 g/t PGE + Au and 3.0 g/t Pt + Pd.

Cu-Au Deposits

Grouped under this heading are the Igarapé Bahia/Alemão, Pojuca and Salobo deposits, in addition to new targets presently in phase of economic evaluation such as the deposits of Sossego, Cristalino, S118, etc. All these deposits have as their main characteristic the association iron oxide Cu-Au (-U-REE).

The Igarapé Bahia and Alemão Cu-Au deposits

Associated with the volcano-sedimentary sequence of the Igarapé Bahia Group, these two deposits (Figs. 8a and 8b) represent, for all practical purposes, a single mineralized entity. The Alemão Deposit, which was discovered by geophysics (Barreira *et al.*, 1999) is the along-strike extension in subsurface of the Igarapé Bahia Deposit. Three types of mineralization are known: volcano-sedimentary, hydrothermal and supergene.

- Volcano-sedimentary mineralization: this type of mineralization was known before the others as the result of drilling. It occurs in the form of disseminated chalcopyrite and pyrite associated with chlorite, principally in the sedimentary rocks and rhythmite beds in the form of banded iron formation units composed of magnetite, fluorite and chalcopyrite (Ferreira Filho, 1985; Ferreira Filho and Danni, 1985; Althoff *et al.*, 1994). Up to the present, this type of mineralization is not of economic interest.

- Hydrothermal mineralization: this type of mineralization was investigated by drilling and described at Igarapé Bahia by Ferreira Filho (1985); Althoff *et al.* (1994), Lindenmayer *et al.* (1998), and Tallarico *et al.* (2000b); and at Alemão by Barreira *et al.* (1999), and Soares *et al.* (1999). There are two types of ore: vein ore and brecciated ore.

The vein ore occurs in the form of small veins (0.7 to 2.7 cm thick) which cut several rock-types, and are surrounded by a chlorite envelope consisting of an association of calcite-quartz-chalcopyrite; quartz-magnetite-chalcopyrite; and quartz-chalcopyrite. Molybdenite and digenite (Althoff *et al.*, 1994), as well as uraninite (Angélica *et al.*, 1996) and fluorite have been described in the area.

The brecciated ore is found in association with hydraulic breccia that is situated preferentially at the contact of granophyre intrusions with sedimentary rocks and mafic volcanic rocks,. The ore is composed of fragments derived from the wall rock of the iron formation units. The matrix is rich in magnetite and quartz, and cemented by chlorite and calcite in addition to chalcopyrite-pyrite-chalcocite-covelite. The gold is mainly associated with chalcopyrite. The mineralization is accompanied by intense hydrothermal alteration represented by the assemblage of chlorite-quartz-albite-carbonate (Ferreira Filho, 1985; Ferreira Filho and Danni, 1985).

At Alemão, the reserves of primary sulphide ore are estimated about 170 Mt of ore at 1.5% Cu and 0.81 g/t Au (or 113Mt of ore with 1.98% Cu and 0.94 g/t Au).

According to Lindenmayer *et al.* (1998), the fluids responsible for the general chloritization and for the mineralization were rich in CO2, U, REE, Cu, Ag, Mo, F end Cl, as well as being highly saline. For these reasons Lindenmayer *et al.* (*op. cit.*) proposed that these fluids came from a magmatic source associated with the anorogenic granite intrusions dated at 1.88 Ga. However, other workers such as Huhn and Nascimento (1997), Oliveira *et al.* (1998), Tallarico *et al.* (1998, 2000 b) related the installation of the hydrothermal system to Archean dioritic intrusions, thus enabling a comparison with the Olympic Dam Deposit in Australia. The Igarapé Bahia was classified as a Fe oxide Cu-Au (-U-REE) deposit.

- A good argument in favour to the second hypothesis is the geological observation that the brecciated mineralization does not cut the Águas Claras sediments that unconformably overlie the Igarapé Bahia volcano-sedimentary sequence.

- Supergene mineralization at Igarapé Bahia: the ore is of lateritic origin (Zang and Fyfe, 1993; Costa *et al.*, 1996; Costa, 1997) consists mainly of hematite, maghemite, goethite with gibbsite, and subordinate kaolinite and quartz. This assemblage formed in the thick weathered zone (20 to 50 m) that developed over the veining rich in chalcopyrite. The gold grains found in the lower part of the ferruginous zone are almost pure. The deposit contained before mining reserves of about 12 Mt of ore at 5 g/t Au, and in 1999 the production was about 10 t of gold.

The Pojuca Cu-Zn-Au deposit

This deposit is situated on the northern side of the Serra dos Carajás (Fig. 7b). The Pojuca volcano-sedimentary sequence that hosts the Cu-Zn-Au deposit consists of a thick sequence of ortho-amphibolite intercalated with banded iron formation units, and overlain by beds of metamorphosed sandstone and siltstone. This sequence is intruded by dykes and sills of metagabbro and metadiabase (Farias and Saueressig, 1982; Farias *et al.*, 1984; Medeiros Neto and Villas, 1985; Medeiros Neto, 1986). The volcano-sedimentary sequence was intruded at 1.88 Ga by the Pojuca anorogenic granite. According to the workers cited above, three types of mineralization can be observed at this deposit.

- Mineralization associated with iron formation units: this type occurs as banded and disseminated sulphide, as massive sulphide, and as siliceous breccia. Although this type of mineralization has been considered as stratiform and of volcano-sedimentary origin, the mineralization associated with siliceous breccia is very similar to the hydraulic breccia of hydrothermal origin described at Igarapé Bahia.

- Mineralization associated with hydrothermal veining of quartz-feldspathic composition: this type of mineralization cuts all the rock-types of the volcano-sedimentary sequence, and is accompanied by propylitic alteration that is symmetric with respect to the axis of the veins. The main minerals are quartz, fluorite, calcite, tourmaline, albite and microcline together with chalcopyrite, pyrrhotite, sphalerite and bornite, besides molybdenite, ilmenite, pyrite, marcasite, cobaltite, hematite, mackinavite, cubanite and pentlandite. These quartz-feldpathic veins seem to be related to remobilization brought about by the intrusion of the Pojuca anorogenic granite at 1.88 Ga. In the vicinity of the Pojuca Granite, the same type of mineralization has been described as that related to the intrusion of the Gameleira Granite, dated at 1.88 Ga (M. Pimentel and Z. Lindenmayer, personal communication).

The Salobo Cu-Au deposit

The Salobo Cu-Au deposit was discovered by DOCEGEO in 1977 (Farias and Sauesserig, 1982). The deposit is situated some 30 km to the N of the Serra dos Carajás (Fig. 7b), and is hosted in a homonymous volcano-sedimentary sequence, the attitude of which is vertical. The rocks consist of quartzite, amphibolite, metagreywacke and iron formation units rich in magnetite to which is associated the copper mineralization (Fig. 9a,b). The deposit is limited by extensive shear zones, the strike of which is WNW-ESE, and is intruded by syn-tectonic granite, foliated and mylonitized (OSG), and dated at 2.57 Ga. At the beginning of the middle Proterozoic, at 1.88 Ga, this sequence was intruded by an anorogenic quartz-syenite (YSG) that caused contact metamorphism and intense hydrothermal potassic and propylitic alteration. The ore of the Salobo Deposit, studied mainly by Lindenmayer (1990, 1998), Lindenmayer and Fyfe (1994), Réquia *et al.* (1997) consists mainly of disseminated or massive bornite and chalcocite, always associated with lenses of magnetite. The other minerals present are chalcopyrite, molybdenite, uraninite, gold, ilmenite, graphite, safflorite, Co-pentlandite, covelite, digenite, hematite and native copper. The ore is hosted in the siliceous and aluminous iron formation units that are tectonically foliated and even mylonitized. This deformation reaches the lenses of magnetite-bornite-chalcocite, which show a plastic flux texture, mylonitization and brecciation. With the intrusion of the anorogenic quartz-syenite (YSG), the iron formation units and the mineralization underwent contact metamorphism in the pyroxene hornfels facies (750°C), followed by potassic alteration (650°C-550°C; 2.5 kbar) and by intense propylitization (350°C-270°C) manifest in a generalized chloritization of the iron formation units by the paragenesis almandine, biotite, hastingsite, titanite, rutile, greenalite in association with fluorite and uraninite, and by the development of veining with stilpnomelane, fluorite, molybdenite, allanite, chalcopyrite, cobaltite, gold and quartz.

The Salobo Cu-Au deposit was initially thought to be a VMS-type deposit of volcanogenic origin. It was successively defined by Lindenmayer (1990) to be volcanogenic; and then of mixed origin: exhalative for copper and epigenetic for Au-Mo-U (Lindenmayer and Fyfe, 1994). Later, it was considered to be porphyritic for Au (Lindenmayer, 1998, 1999), and finally as a deposit of the Fe-oxide Cu-Au (-U-REE) type associated with anorogenic granitic intrusions at 1.88 Ga by Lindenmayer *et al.* (1998) or with Archean granitic intrusions (Huhn and Nascimento, 1997; Huhn *et al.*, 1999). According to the contrasting views of Siqueira (1996), the Salobo Deposit is of hydrothermal origin, entirely related to a shear zone. A fundamental point to be clarified is the observation that the magnetite-chalcopyrite-bornite massive bodies have been affected by the Archean deformation as described initially by Lindenmayer (1990).

The reserves of the Salobo Deposit are estimated at 789 Mt of ore at 0.96% Cu, 0.52 g/t Au and 55 g/t Ag.

The Sossego-Cristalino/S118 Cu-Au deposits

The new discoveries of Cu-Au deposits known as Sossego-Cristalino-S118 are situated to the S of the Serra dos Carajás (Fig. 7b). According to Huhn *et al.* (1999), these deposits are associated with diorite and quartz diorite intrusions, dated at 2.74 Ga, in the volcano-sedimentary sequence of the Grão Pará Group. The attitude of these rocks is vertical and metamorphism has occurred in the greenschist to amphibolite facies. The mineralization occurs in brecciated and hydrothermalized zones, forming stockworks both in the volcanic wall rock as well as in the diorite. The mineralization is accompanied by a distinct enrichment in magnetite, apatite and allanite in addition to varied types of hydrothermal alteration including

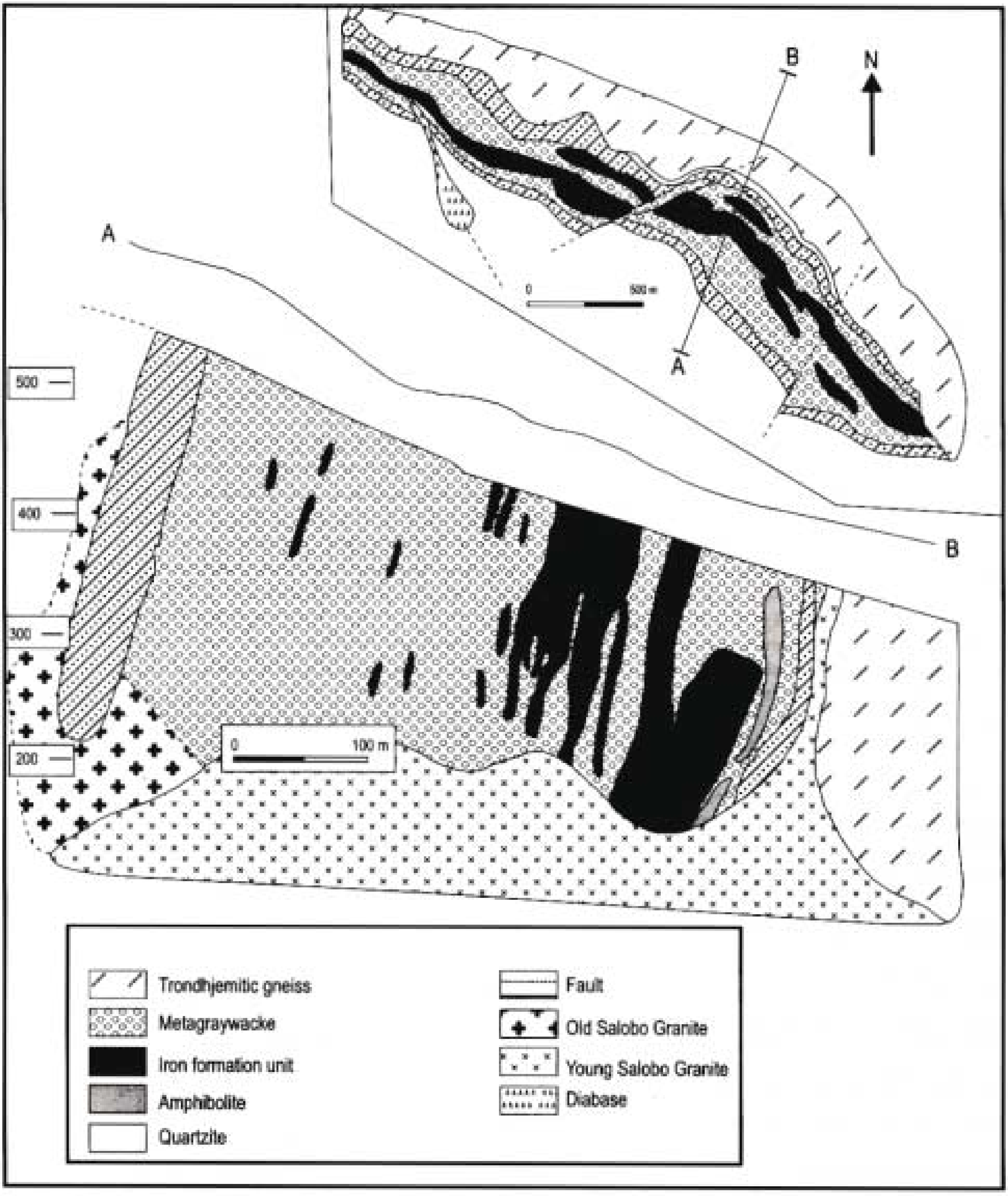


FIGURE 9a,b - Schematic geological map and cross-section of the Cu-Au Salobo Deposit (modfied after Lindenmayer,1998).

microclinization and biotitization (potassic alteration), albitization and scapolitization (sodic alteration), chloritization and carbonatization, silicification and tourmalinization. The ore consists of chalcopyrite, pyrite, magnetite, bravoite, willerite, cobaltite, vesite and gold. Mineralized intersection from the first drillholes showed grades of 1.4% Cu and 0.25 g/t Au, and the potential is estimated to exceed 200 Mt of copper ore. The partial reserves of the Sossego Deposit have been estimated at about 219 Mt of ore with 1.14% Cu and 0,34g/t Au. Huhn *et al.* (1999) consider that these new discoveries show features that are also common to the Cu-Au deposits of Alemão, Bahia, Pojuca and Salobo, and suggest these deposits will fall into the Fe-Cu-Au-U-REE model of Hitzman *et al.* (1992), in line with the initial view proposed by Huhn and Nascimento (1997).

The Azul Mn deposit

The Azul manganese deposit (Coelho and Rodrigues, 1986; Beisiegel *et al.*, 1973; Bernadelli and Beisiegel, 1978; Bernadelli, 1982) is associated with the lower member of the Águas Claras Group (Nogueira *et al.*, 1995) and dated indirectly at 2.645 Ga. In the Serra dos Carajás area the denomination Águas Claras Group has substituted the former Rio Fresco Group (Barbosa *et al.*, 1966). The lower member of the Águas Claras Group consists of argillite with subordinate siltstone and sandstone, deposited on a marine platform. The primary mineralization, or manganese proto-ore (Bernadelli and Beisiegel, 1978; Valarelli *et al.*, 1978) is found intercalated in the pelitic sequence in the form of two manganiferous units consisting essentially of rhodochrosite, quartz, phyllosilicate, feldspar and organic material. The ore, of lateritic origin, developed from the alteration and supergene enrichment in or on the lower manganiferous unit in the form of cryptomelane. In 1996 the mineable reserves were estimated at 54.36 Mt of manganese ore. The annual production was 1.1 Mt of metallurgical-grade ore and 70 381 t of manganese dioxide.

The Azul manganese deposit may be compared to sedimentary deposits developed at the margin of stratified and anoxic basins (Force and Maynard, 1991).

The Buritirama manganese deposit, isolated to the N of the Carajás Mineral Province may represent a metamorphic equivalent of the Azul sedimentary sequence or the Pojuca/Salobo volcano-sedimentary groups.

The Serra Pelada/Serra Leste Au Deposit

The Serra Pelada/Serra Leste gold deposit (Meireles *et al.*, 1982; Meireles and Silva, 1988) is hosted in the sediments of the Águas Claras Group, that form a regional synclinal structure, and is associated with a dextral transtensional system the strike of which is NNE-SSW, developed between regional shear zones the strike of which is ENE-WNW (Freitas-Silva, 1999). In this dilational environment the mineralizing fluids deposited gold in manganiferous and carbonate-rich tectonic breccia. The gold is free and rich in palladium (1% to 8% Pd) and platinum. The ore also contains silver (*c.* 0.5% Ag), iron (between 0.5%-1% Fe) and copper (*c.* 0.5% Cu). According to Tallarico *et al.* (2000 a) the ore bodies are found in the contact between the carbonate-rich siltstone and dolomitic marble and are surrounded by intense silicification. Quartz, dolomite, chlorite, actinolite, biotite, muscovite, magnetite, calcite, tourmaline, hematite, pyrite, chalcopyrite, molydbenite, galena, digenite as well as uranium-bearing minerals and rare earths accompany the Au-Pd-Pt mineralization. A hydrothermally-altered dioritic intrusion is known to occur underneath the deposit to which can be related the epigenetic mineralization.

The deposit has been highly weathered, and is known for the development of large nuggets (weighing up to 6 kg of massive gold) as well as huge gold agglomerates having dendritic and skeletal habits weighing between 26 and 62 kg. The deposit was discovered in 1977, and became famous on account of the number of *garimpeiros* (over 40 000) that worked there. Between 1980 and 1984 the Serra Pelada Deposit produced about 32.6 t of gold. The oxidized mineralization is known to a depth of 300 m.

The Águas Claras Cu-Au Deposit

The Águas Claras Cu-Au deposit is associated with shear zones and normal faults, the attitude of which is N20° to N40°E/70°NW, which cut the sediments of the Águas Claras Group, consisting of siltstone, sandstone, and sills and dykes of diabase. According to Soares *et al.* (1994) and Silva and Villas (1998), the primary mineralization is of hydrothermal origin. The first phase consisted of the development of massive quartz veins with cassiterite and wolframite accompanied with silicification of the mafic rocks and the tourmalization of the sediments. Chloritization and sericitization, intense brecciation of the quartz veins and the precipitation of sulphides mark the second phase of hydrothermal mineralization. The dominant sulphide species are chalcopyrite, pyrite and arsenopyrite; subordinate species are pyrrhotite, sphalerite, stannite, cobaltite, bismuthinite and galena, besides magnetite. The gold, rich in silver (*c.* 25% Ag), appears as irregular grains (0.01 to 0.26 mm) in the contact between pyrite and chalcopyrite crystals, as inclusions in arsenopyrite, and as isolated grains in chalcopyrite. The origin of the Au-Cu-Sn and W mineralization may be related to the Central Carajás granite intrusion dated as 1.88 Ga. The reserves of the deposit have been estimated at 9.5 Mt of ore at 2.43 g/t Au (Silva and Villas, 1998).

The lateritic weathering resulted in the formation of gossan at the surface (Angélica *et al.*, 1996), and the secondary mineralization in the form of gold nuggets is associated with goethite, limonite, hematite and martite, besides quartz, kaolinite and tourmaline. This weathered cap is worked for support of the industrial gold complex at the Igarapé Bahia Mine.

The Pedra Preta W deposit.

The Pedra Preta wolframite deposit (Cordeiro and Silva, 1986; Rios *et al.*, 1998) is associated with the Musa anorogenic granite intrusion (1.88 Ga) (Dall'Agnol *et al.*, 1994). The mineralization is contained in quartz veins with strike N80°W/subvertical, which cut the volcano-sedimentary sequence. Together with the quartz, there occur besides wolframite, topaz, muscovite, tourmaline, pyrite,

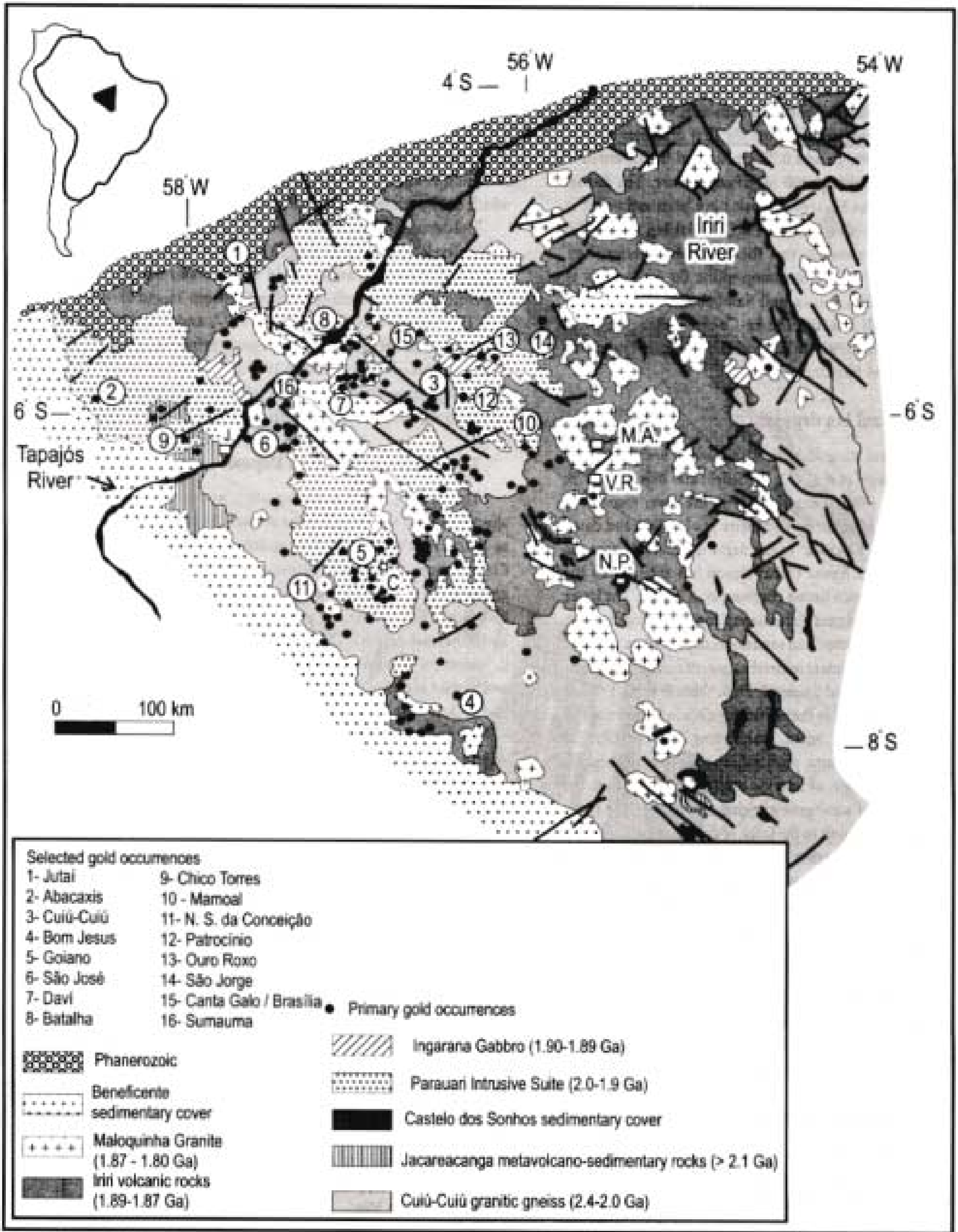


FIGURE 10 : Simplified geological map of the Tapajós Province in the Central Brazil Shield, showing the distribution of primary gold occurrences. Sources: Faraco et al (1996); Robert (1996). J.= Jacareacanga; M.A.= Morais Almeida; N.P.=Novo Progresso; V.R. = Vila Riozinha; C.=Creporizão

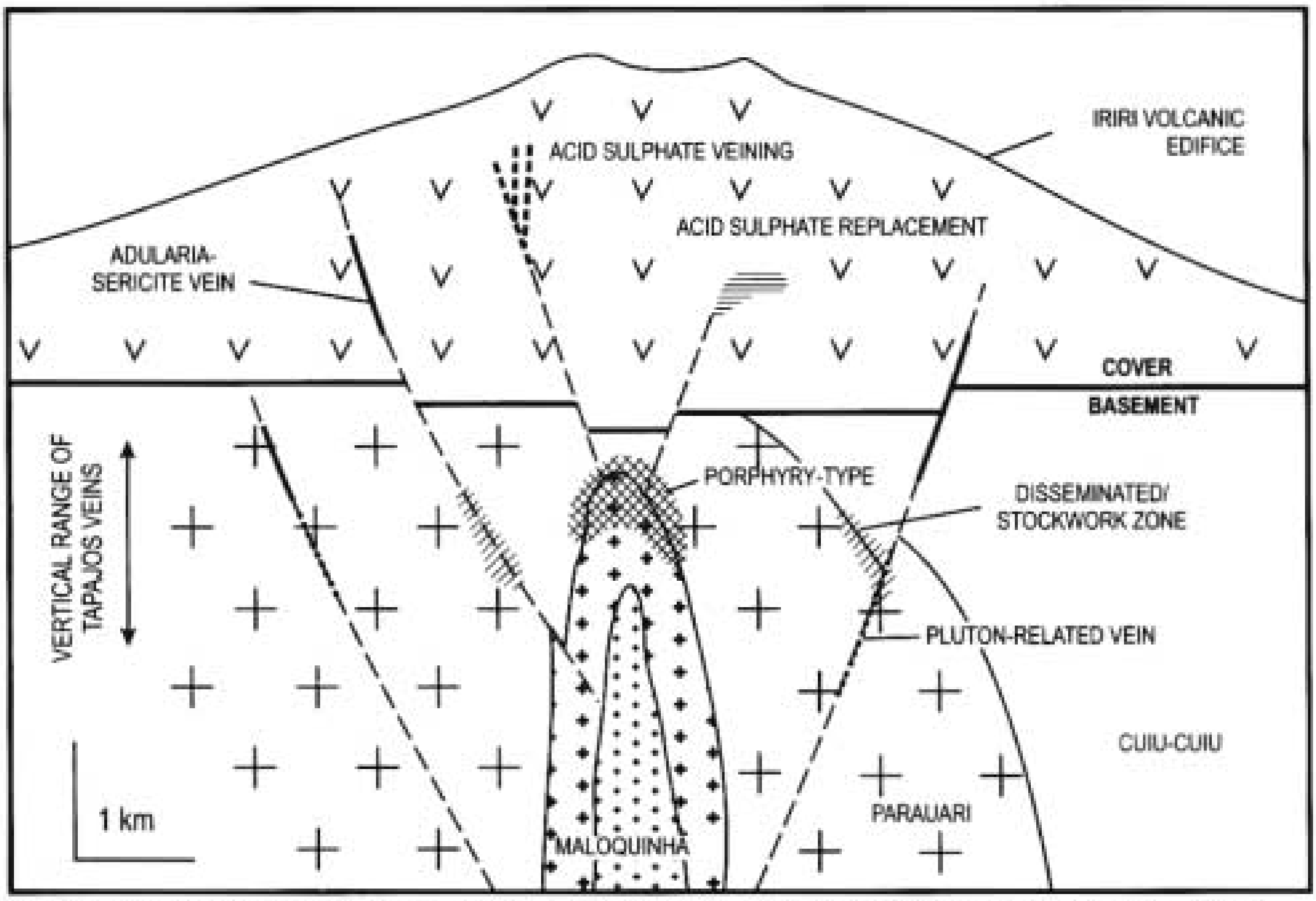


FIGURE 11 - Schematic geological model based in an intrusion - centered hydrothermal gold system applied to the Maloquinha Granite in the Tapajós Province (modified after Sillitoe, apud Robert, 1996)

pyrrhotite and chalcopyrite. In the contact with the thicker veins, the wolframite is also disseminated in the intensely silicified wall rock. The ore reserves have been estimated at about 322 753 t at an average grade of 1.03% WO3.

The Tapajós Province

Geological context

Tapajós is the largest gold province in Brazil. It is situated in the central region of the Central Brazil Shield between the Tapajós River and the headwaters of the Crepori and Jamanxim rivers, the Serra do Cachimbo and the Iriri River (Fig. 10). The basement of this province consists of two main units: the Cuiú-Cuiú and the Jacareacanga Metamorphic suites of possible Paleoproterozoic age (>2.0 Ga). The Cuiú-Cuiú Suite consists of gneiss, migmatite, granitoid and amphibolite whereas the Jacareacanga Suite represents a supracrustal volcano-sedimentary sequence that has been metamorphosed and deformed in the high greenschist facies. The relationship of this unit with the Cuiú-Cuiú Metamorphic Suite is not known. Both these metamorphic suites were intruded by granodiorite and monzodiorite plutons and batholiths of the Parauari Suite considered to be syn to late tectonic (2.0-1.9 Ga). The Parauari Granitoid is calc-alkaline in composition, and often displays a rapakivi texture. All these units form the basement over which was deposited an extensive cover of sub-aerial volcanic rocks of acid to intermediate composition known as the Iriri Group (1.89-1.87 Ga), and an anorogenic co-magmatic intrusive suite, the Maloquinha Suite (1.87-1.80 Ga). The magmatism of the Iriri-Maloquinha rocks occurred during a geotectonic regime that was mainly extensional. The volcanic rocks and their sub-volcanic plutons are not penetratively deformed. The volcanic rocks with pyroclastic units and subordinate sediments have low dips, except in the proximity of faults. Clastic fluvial sedimentary rocks of the Castelo dos Sonhos Formation, underlie the Iriri Formation. These sediments generally form elongate quartzite ridges. An intrusion known as the Ingarana Gabbro Intrusion (1.90-1.89 Ga) cuts the Parauari Intrusive Suite and the Cuiú-Cuiú basement sequence. All these units are overlain discordantly by a sedimentary cover, locally with volcanic rocks of taphrogenic character assigned to the Beneficente Group. Basic sills and dykes of the Crepori unit intruded these rocks at about 1.69 Ga, as well as the mafic-ultramafic rocks of the Cachoeira Seca unit, dated at between 1.2 and 1.1 Ga (Robert, 1996; Faraco *et al.*, 1996).

The late stage Maloquinha Granitoid of alkaline to sub-alkaline affinities intrudes the Parauari Suite and the Iriri volcanics. These younger granitoid plutons are more fractionated than those of the Parauari Suite with Rb/Sr >1, and high Nb/Zr and Gd/Yb ratios. The Maloquinha intrusive rocks were emplaced at a very shallow depth. They have a porphyritric texture, and they are frequently reddish in colour. In addition to these characteristics they are anomalous with respect to F, Zr, REE, Y, Sn, Au and Cu, and play an important role in the mineral economy of the Tapajós Province (Jacobi, 1999).

Gold Mineralization

Primary gold occurrences are found mainly in the western part of the province in rocks of the basement units as well as in the Maloquinha intrusions. There are relatively few occurrences in the volcanic rocks of the Iriri cover. This distribution of the gold deposits (Fig. 11) may reflect the influence of the level of erosion in as much as these may expose certain styles of gold mineralization (Robert, 1996; Coutinho *et al.*, 1998).

Of the diverse occurrences of primary gold, two main styles of mineralization are present: (1) gold in quartz veins and (2) gold as disseminations in zones and stockworks. Quartz veining is the most common style of mineralization and can be observed at many localities including Abacaxis, Bom Jesus, Goiano, São José, Davi, Batalha, Chico Torres, Mamoel, Cuiú-Cuiú and at N. S. da Conceição. In general, the gold occurs in narrow discontinuous zones. Very often the gold is visible forming high-grade concentrations. The hydrothermal alteration is restricted to the veins and is never pervasive. Disseminated gold mineralization can be observed at Jutaí and Abacaxis.

The principal characteristics of the quartz veins may be summarized as follows: (a) they are polymetallic and generally contain pyrite, along with variable amounts of chalcopyrite, galena, sphalerite, pyrrhotite and molybdenite. Some veins contain alkaline feldspar, amethyst, and fluorite; carbonate is common; (b) at the occurrences there may be observed massive quartz to banded quartz structures (or comb quartz structures), and structure filling open spaces are also common; (c) the associated alteration consists of proximal sericite-pyrite (Bom Jesus) or K-feldspar (Batalha) or peripheral mineral assemblages of chlorite-epidote-calcite (Bom Jesus, Davi) or chlorite/sericite-sulphide-carbonate (Ouro Roxo); (d) the veins are associated spatially with ruptile faults, and there is some evidence that, locally, the veins may have been emplaced in active brittle faults (Bom Jesus, Goiano). These characteristics are typical of veins that have formed in relatively shallow crustal levels and indicate an *epithermal affinity for the veining of the Tapajós Province; (f)* the quartz veins occur in a variety of host rocks: granitic basement (Goiano), Maloquinha Granite (Bom Jesus) and non-metamorphosed feldspathic sandstone (Abacaxis); (g) the absence of penetrative deformation and significant metamorphism of the Maloquinha intrusives and of the sandstone at Abacaxis may indicate that the veining formed at shallow crustal levels, a view that corresponds to the inferred environment of the veins; (h) in spite of the diversity of the different types of host rock and their ages, the similarity of the characteristics of the veins suggests that the quartz veins have similar ages. Consequently, the veins are contemporaneous or younger that the Maloquinha intrusives (Robert, 1996).

The absence of primary gold occurrences in the Beneficente Group suggests that the veining is older than this group. Based on available data, the most likely interpretation is that the quartz veins were formed during the magmatic event that resulted in the Maloquinha intrusions at the end of the Paleoproterozoic (1.87-1.80 Ga). A possible exception is the veining at São José where the veins are parallel to the penetrative foliation in basement granitoid, and which form boudins along the foliation planes. The implication of this boudinage is that the veining was developed during the phase of basement deformation and is therefore older (2.4-2.0 Ga), this is to say at the base of the Paleoproterozoic. This fact raises the important possibility that there occurred two events that brought about the gold veining in the Tapajós Province: one that affected the basement rocks consisting of mesothermal veins in the lower Paleoproterozoic, and a second phase of veining in the younger Paleoproterozoic. Additional work is required to examine these two possibilities. Finally, it is important to note that the quartz veining is related to the Iriri-Maloquinha magmatic event. These veins are not abundant in the Iriri volcanic cover. The tendency is for these to occur in rock-units below this cover and within the Maloquinha intrusions.

The presence of disseminated auriferous sulphide mineralization in the Abacaxis Granodiorite and in weak to moderate fracture stockworks at Jutaí and São Jorge is highly significant in spite of the fact that this mineralization is not ore-grade. It shows that the disseminated stockwork mineralization, generally associated with the porphyritic environment (Sillitoe, 1991; Robert, 1996) could constitute a valid exploration target for the Tapajós Province. At Abacaxis, the granodiorite hosting disseminated sulphide mineralization intrudes non-metamorphosed sandstone and siltstone beds that are believed to be coeval equivalents of the Iriri volcanic rocks. This also implies that the granodiorite belongs to the Maloquinha Intrusive Suite, and that the diseminated mineralization may possibly have the same age as the quartz veining.

As mentioned above, the quartz veins have epithermal affinities. Specifically, these show several common characteristics with the epithermal deposits of the adularia-sericite type (Heald *et al.*, 1987; Robert, 1996). These include K-feldspar/sericite/chlorite alteration; weak sulphide mineralization in veins (sphalerite, chalcopyrite, and galena); and hematite and adularia (Dreher *et al.*, 1998) in some veins, and an absence of hypogene alunite, enargite-tenantite and a high degree of argillization. However, in the Tapajós veins *there occur a few instances in which the veining diverges from* the classical model of the adularia-sericite-type: the veins occur predominantly in the basement rather than in the volcanic cover; many veins are polymetallic, and as emphasized by Robert (1996) and Buchanan (1981) do not show a vertical separation of the basic and precious metals. The veins have some similarities with what Sillitoe (1991) and Robert (1996) referred to as plutono-related veining. These are transitional in character between adularia-sericite epithermal and mesothermal veins that occur at a slightly greater depth as compared to the epithermal veins. Figure 11 shows the position of the vein-types within a hypothetical hydrothermal system centered on an intrusion, and provides a plausible geological model for the gold deposits of the Tapajós Province.

More than 90% of the gold produced in the Tapajós Province was mined from thousands of placer deposits. Production figures show that about 159 t of gold were mined between 1959 and 1996. *Garimpos* of the Cuiú-Cuiú, Canta Galo, Abacaxis and Patrocinio types are classical examples of these large Recent placer systems. Gold is also found in Tertiary paleoplacers, some 10 to 20 m below the

Recent surface (Nova Brasilia) or in paleoplacers above the present surface (Sumaúma) (Robert, 1996; Dreher *et al.*, 1999; Martini, 1998; Costa and Carvalho, 1999; Coutinho *et al.*, 1998; Jacobi, 1999)

The Alta Floresta Province

Geological Context

The Alta Floresta Province is also essentially auriferous, in like manner to the Tapajós Province. It lies between the Serra do Cachimbo to the N and the Serra dos Caiabis and the Chapada dos Dardanelos to the S. To the E it is limited by the Peixoto de Azevedo/Matupá region, and by the Aripuanã River to the W (Fig. 12).

With the exception of some studies in specific areas the level of geological knowledge of the Alta Floresta Province is limited for the best part to geological mapping at 1:1 000 000 scale. Presently, geological mapping at 1: 250 000 scale is being carried out by the Geological Survey of Brazil. According to Tassinari (1996) the Alta Floreta region corresponds to the limit of two geological provinces or tectonic belts: the Ventuari-Tapajós Belt (1.9-1.8 Ga) to the N, and the Rio Negro-Juruena Belt (1.8-1.55 Ga) to the S, both defined by this author mainly on the basis of geochronology, supplemented by geological data. The limit between these two provinces passes approximately along a line between the towns of Matupá – Alta Floresta – Paranaita – Apiacás (Fig. 12). These two geochronological provinces or tectonic belts were interpreted by Tassinari (1996) and Teixeira *et al.* (1989) as having developed in magmatic arcs originating by collision directed against a continental block situated to the NE (Central Amazonian Province) (Fig. 4).

The basement of the Alta Floresta Province consists essentially of granitoid plutons of granitic to monzogranitic composition and gneiss of granitic to tonalitic composition. There also occur schist and mafic and ultramafic rocks, banded iron formation units, migmatite and other rock-types. In the basement of the Alta Floresta Province occur large batholiths (Juruena-type granitoid) that are isotropic to foliated banded and monzogranitic to granodioritic in composition (Paes de Barros, 1994; Paes de Barros *et al.*, 1999). The batholiths of the Juruena-type have not been adequately mapped yet. Recent U/Pb dating of a number of samples of basement granitoid, collected between the towns of Apiacás and Peixoto de Azevedo, show ages between 1.9 and 1.8 Ga for these intrusive rocks (JICA/MMAJ, 2000). On the other hand, a dozen samples of granite collected from the areas of Apiacás and Paranaita, dated by K/Ar, showed that these rocks have been reworked isotopically between about 1.35 and 1.10 Ga (JICA/MMAJ, 1999).

Overlying the above-mentioned basement occur acid to intermediate volcanic rocks with calc-alkaline affinities, including pyroclastic rocks (Teles Pires Group). There are few reliable dates available for this volcanism correlated by some authors to the Iriri Group. U/Pb dates collected near the town of Apiacás to the NW, and to the N of Matupá, showed ages of 1.78 Ga (JICA/MMAJ, 2000) and 1.8 Ga (C. Schobbenhaus, personal communication), respectively, for the Teles Pires Volcanics. About 400 km to the W of Apiacás, in the Moreru area, an ignimbrite sample presents an U/Pb age of 1.81 Ga for this volcanism (Pinho *et al.*, 1999). These dates indicate that the Teles Pires volcanism is about 100 Ma younger than that of the Iriri volcanic rocks exposed in the Tapajós Province. The Teles Pires volcanic rocks are cut by granitic intrusions (Teles Pires Granite) that are generally circular in plan, and similar in this respect to the Maloquinha Intrusive Suite, occurring associated with sub-volcanic rocks with alaskite affinities. U/Pb age determination for this granite gave *c.*1.76 Ga (J. Orestes Santos, personal communication). Mesoproterozoic sediments of the Beneficente Group, the thickness of which exceeds 1000 m, cover the volcanic rocks and granite intrusives of the Teles Pires Group. These sediments were probably deposited between 1.8 and 1.5 Ga in a continental rift, the strike of which was NW-SE. Morphologically, they form an extensive plateau known as the Chapada do Cachimbo. The Beneficente Group consists of continental and marine sediments intercalated with pyroclactic rocks. Sedimentary rocks from this group revealed Rb/Sr ages of 1.4 Ga, interpreted as the diagenetic age of the sediments (Tassinari *et al.*, 1978). In the Aripuanã region there occurs associated sedimentary and volcanic rocks of low metamorphic grade that are also correlative with the Beneficente Group. K/Ar and Rb/Sr ages of *c.* 1.5 Ga and 1.45 Ga, respectively, for diabase and alkaline rocks intruded into the sediments represent minimum ages for the sedimentation The southern limit of the Alta Floresta Province is marked by two important physiographic features: the Serra dos Caiabis and the Chapada dos Dardanelos in which there occur continental sedimentary sequences known as the Dardanelos Formation. This formation overlies the volcanites of the Teles Pires, the Beneficente Group and the basement. Alkaline basalt flows intercalated in the Dardanelos Formation gave K/Ar ages of between 1.4-1.2 Ga. To the NW of the Alta Floresta Province, in the region of the Sucunduri Dome, continental sediments of the Prosperança Formation, of possible Neoproterozoic age, overlie the Beneficente Group (Iwanuch, 1999).

Gold Mineralization

Gold is the most important mineralization related to the Alta Floresta Province. In second place occurs base metal mineralization. Gold mineralization, whether this be in secondary alluvial deposits or primary deposits, is widespread throughout the province for more than 500km, and especially along the southern margin of the Cachimbo Graben that strikes WNW-ESE. Most of the gold occurrences have received little study. Mining began in 1966 following the discovery of gold by *garimpeiros* along the Juruena River. Official production figures and estimates of the production from small workings in alluvial deposits between 1982 and 1995 (DNPM) are 112 and 148 t of gold, respectively (Peixoto de Azevedo, Colider, Matupá, Terra Nova do Norte, Guarantã do Norte, Alta Floresta, Apiacás, Paranaita and Aripuanã areas).

Paes de Barros *et al.* (1999) divided the occurrences of gold mineralization into four districts: Peixoto de Azevedo, Teles Pires, Cabeça and Aripuanã. According to these authors, in the Peixoto de Azevedo area that extends from the vicinity of the town of Peixoto de Azevedo to the vicinty of the town of Alta Floresta, important gold mineralization

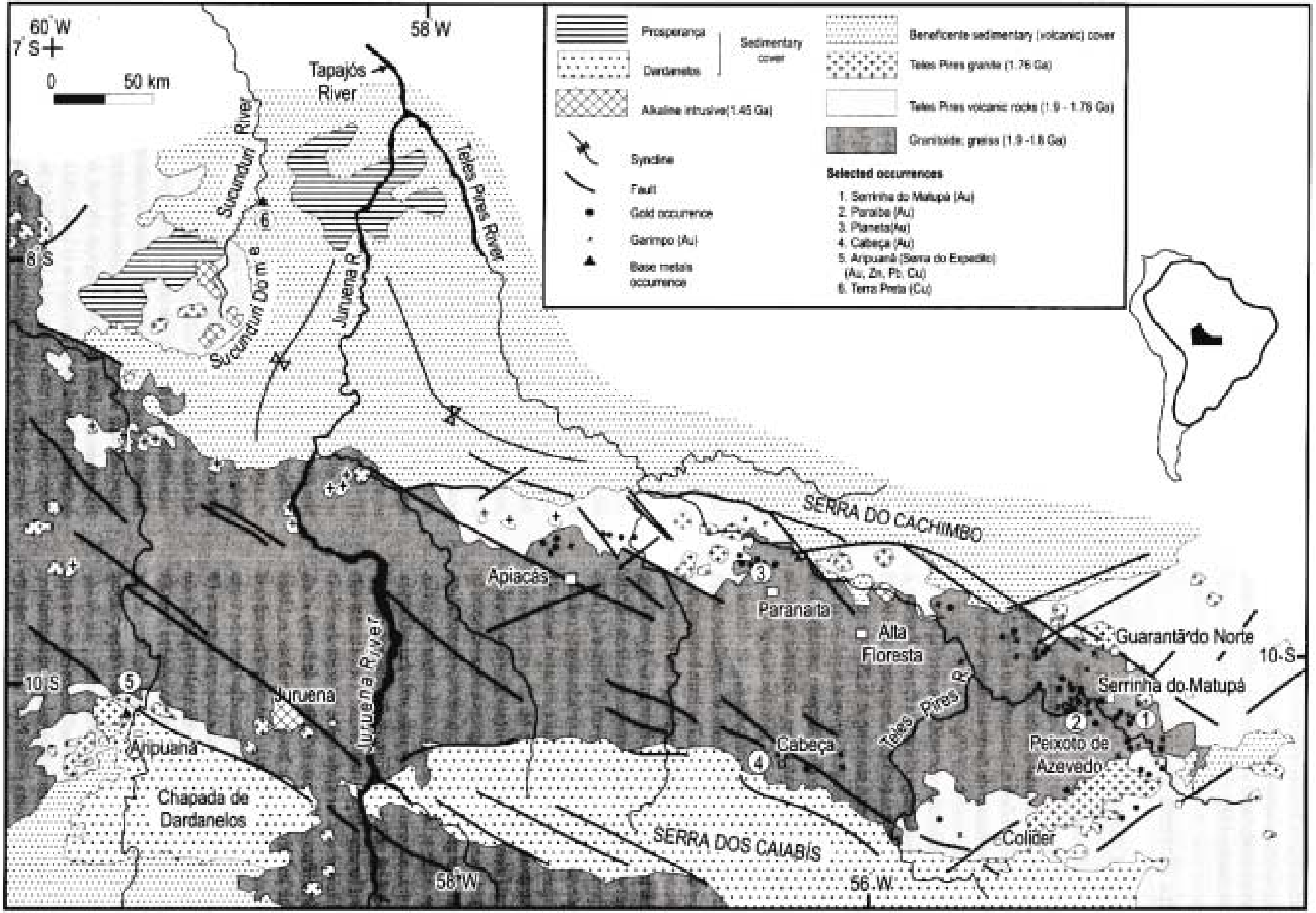


FIGURE 12 : Simplified geological map of the Floresta Province in the Central Brazil Shield. Sources of information: JICA-MMAJ (1999), Carvalho and Figueiredo (1982), Paes de Barros et al. (1999), and others.

is hosted in shear bands and in extensional structure that strike NNW-SSE and WNW-WSE. The ductile shear zones have as their main characteristic, continuous lineaments representing quartz-mylonite with the development of extensive pervasive hydrothermal alteration including silicification, chloritization, sericitization, epidotization and propilitization. Dozens of occurrences are hosted in shear fractures, including those at the Paraiba, Cubu, Pezão, Edu, Edson, Goiano, Mineiro workings and others. At the Serra do Guarantã occurrence, vein mineralization can be observed hosted in talc-chlorite schist that form mega-enclaves of ultramafic nature. Deposits related to granitic apophyses and stocks are frequently related to the Teles Pires magmatism. These may be observed as veins, veining and in stockworks, as for example, at the Pé Quente, Trairão, Aluizio, Naiuram *garimpos* and others. The Teles Pires area (Paes de Barros *et al.*, 1999) lies in a belt, over 200 long, that strikes E-W to NW-SE extending from Paranaita, passing through Apiacás to the Juruena River and reaching the area of the Moreru River. In general, the mineralization of this region is related to pre-Teles Pires magmatism represented by granitic batholiths, the rocks of which consist of equigranular biotite-monzogranite, light grey in colour. In the vicinity of the mineralized zones occurs a facies showing a higher degree of hydrothermal alteration with large bluish quartz crystals associated with an epidote, chlorite and pyrite paragenesis. The gold mineralization occurs associated with sulphide in quartz veins and as disseminations hosted in closely-spaced multiple shear bands. Gold also occurs associated with sub-volcanic acid rocks in pockets and in disseminated pyrite with grades exceeding 10 g/t Au. In the Cabeça region the gold mineralization occurs associated with a probable volcano-sedimentary sequence that underwent several phases of cataclasis, and which is locally intruded by the Teles Pires Granite. This volcano-sedimentary sequence is conditioned by a ductile shear zone, the strike of which is N70°-80°W. Thin gold-bearing veins with high-grade ore strike N20°-30°E and N5°-15°W.

The primary gold mineralization of the Alta Floresta Province may be divided into three types: (1) shear zone-hosted quartz vein type; (2) porphyry or disseminated type; (3) stockwork type. Examples of each of these three types are Paraiba, Matupá and Nova Planeta, respectively (Fig. 12).

The gold mineralization of the shear zone-hosted quartz vein type is related to a ductile shear zone that has the regional strike NW-SE, which cuts the entire province. This shear zone may be several kilometres wide, and includes some dozens of important gold veins and hundreds of zones with smaller gold veins.

This type of mineralization is best seen at the Paraiba underground mine, which has been considered to be the most important working in which veining is hosted in a shear zone. Reserves are given as *c.* 4.3 t of gold. The zones with quartz veins and veinlets have the preferential strike N20°-60°E, NNE, N30°-60°W and E-W. The Paraiba lode consists of a network of quartz veins with gold and copper displayed in parallel bands with variable sulphide content (Siqueira, 1997; Paes de Barros, 1994; JICA/MMAJ, 1999).

The gold mineralization of the porphyry or disseminated-type is associated with calc-alkaline granitic Type I plutons, of which the Serrinha do Matupá Granite is the best example (Botelho *et al.*, 1997; Moura, 1998; Botelho and Moura, 1998). The age of this granite (1.87 Ga, Pb/Pb), together with its geochemical characteristics are favourable indicators for gold prospecting in the region. The gold in this granite occurs in small high-grade vein-type deposits as well as in disseminations in wide hydrothermally-altered zones with sericitization, feldspathization, and pyritization. The association of gold with oxidized Type I granite and the hydrothermal alteration style are analogous to the associations present in world-class porphyry-type deposits (JICA/MMAJ, 1999). The Serrinha do Matupá Massif is a homogeneous, non-deformed monzogranite, with equigranular to porphyritic texture. The geochemical characteristics are similar to those formed in volcanic arcs as well post-collisional bodies, formed in association with oceanic lithosphere.

The stockwork-type gold mineralization is observed in the Teles Pires Intrusive Suite. It is controlled by regional lineaments or shear zones, and best observed in the area of Nova Planeta around the periphery of a monzogranite of the Teles Pires type. Here the strike is E-W, a direction that also coincides with that of a prominent shear zone. This monzogranite is intruded in basement granitoid as well as in volcanic rocks of the Teles Pires Group, as shown by the presence of dykes and apophyses of the intrusion (JICA/MMAJ, 1999; Veiga, 1988).

Base Metals Mineralization

In the Aripuanã region (Fig. 12), low metamorphic acid volcanic and pyroclastic rocks, problably related to the Teles Pires volcanism, host hydrothermalized polymetallic Zn, Pb, Cu, Au and Ag deposits. At the Serra do Expedito *garimpo*, situated to the NW of Aripuanã, the country rocks are metasiltstone that has been correlated to the Beneficente Group, and the mineralization resembles a gossan. There also occurs in this area polymetallic mineralization of the stratabound type related to the intersection of shear zones, striking E-W, in carbonate-rich pelite of the Beneficente Group. The mineralization occurs as massive, disseminated and veined sulphide with pyrite, pyrrhotite, sphalerite galena, and subordinately, gold and silver. Gold mineralization is also associated with the periphery of sub-outcropping intrusives (Neder *et al.*, 1998; Paes de Barros *et al.*, 1999).

In the Moreru area on the Aripuanã River, situated about 250 km NNW of Aripuanã area, Pinho *et al.* (1999) refer to hydrotermalized acid volcanics (1.81 Ga) and pyroclastic rocks with Au mineralization associated with pyrite, chalcopyrite, galena and ilmenite. Sulphides occur as disseminated form or in massive bands in subvolcanic bodies and in quartz-carbonate-chlorite veins cutting rhyolite. These rocks are related to the Teles Pires magmatism.

At Terra Preta, on the Sucunduri Dome (Fig. 12), the Beneficente Group bears copper mineralization (chalcopyrite and bornite) in beds of sandstone, argillite, calcareous sandstone and limestone. The mineralized zones are 7 to 12 m thick and contain copper grades of about 0.35% Cu (Carvalho and Figueiredo, 1982).

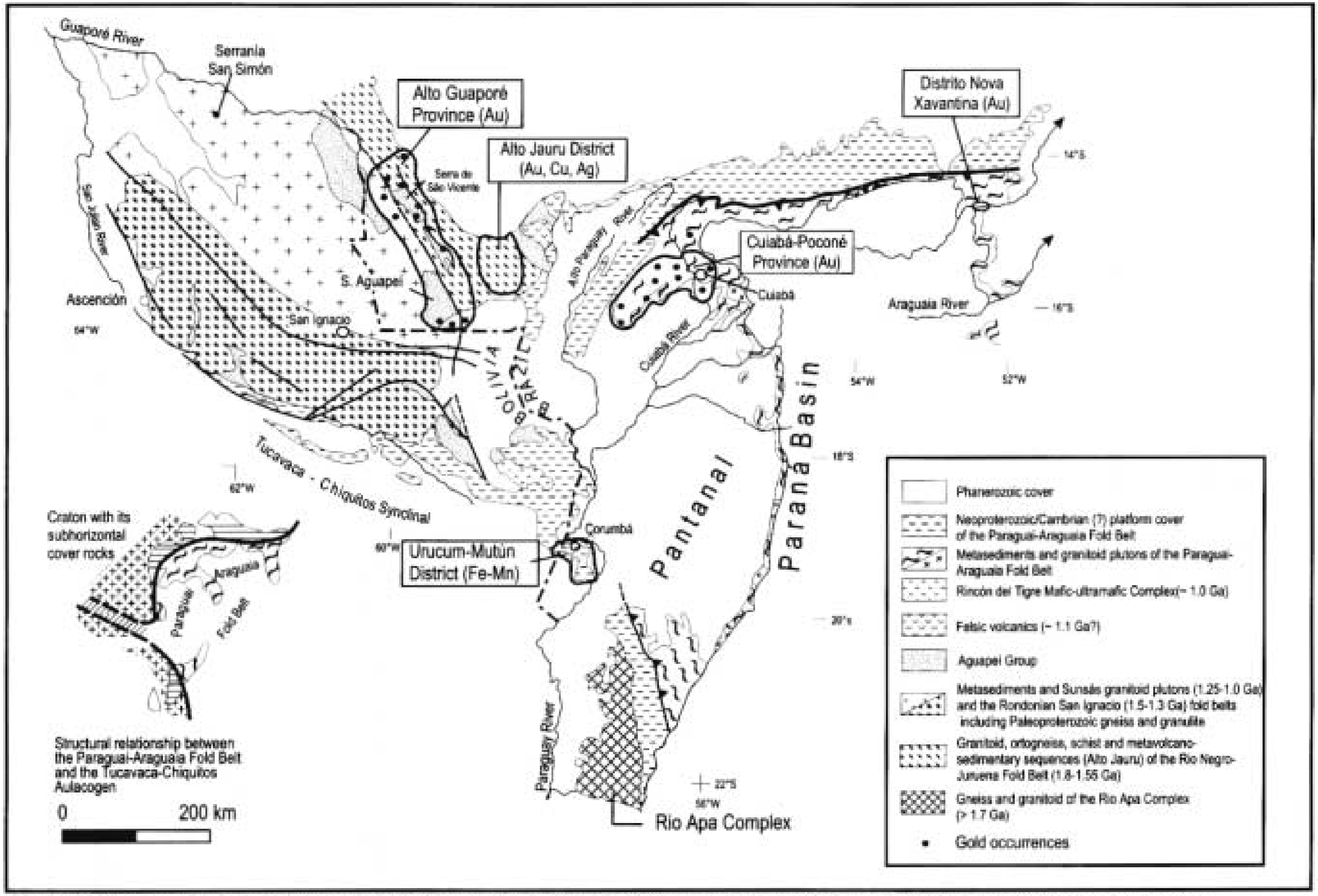


FIGURE 13 - Schematic geological map of the southwestern area of the Amazonian Craton and part of the Paraguai-Araguaia Fold Belt, showing the location of the Alto Guaporé and Cuiabá-Poconé provinces and the Alto Jauru, Nova Xavantina and Urucum-Mutún districts. Modified after Trompette (1994); Litherland et al. (1986); Tassinari and Macambira (1999); and other sources cited in text.

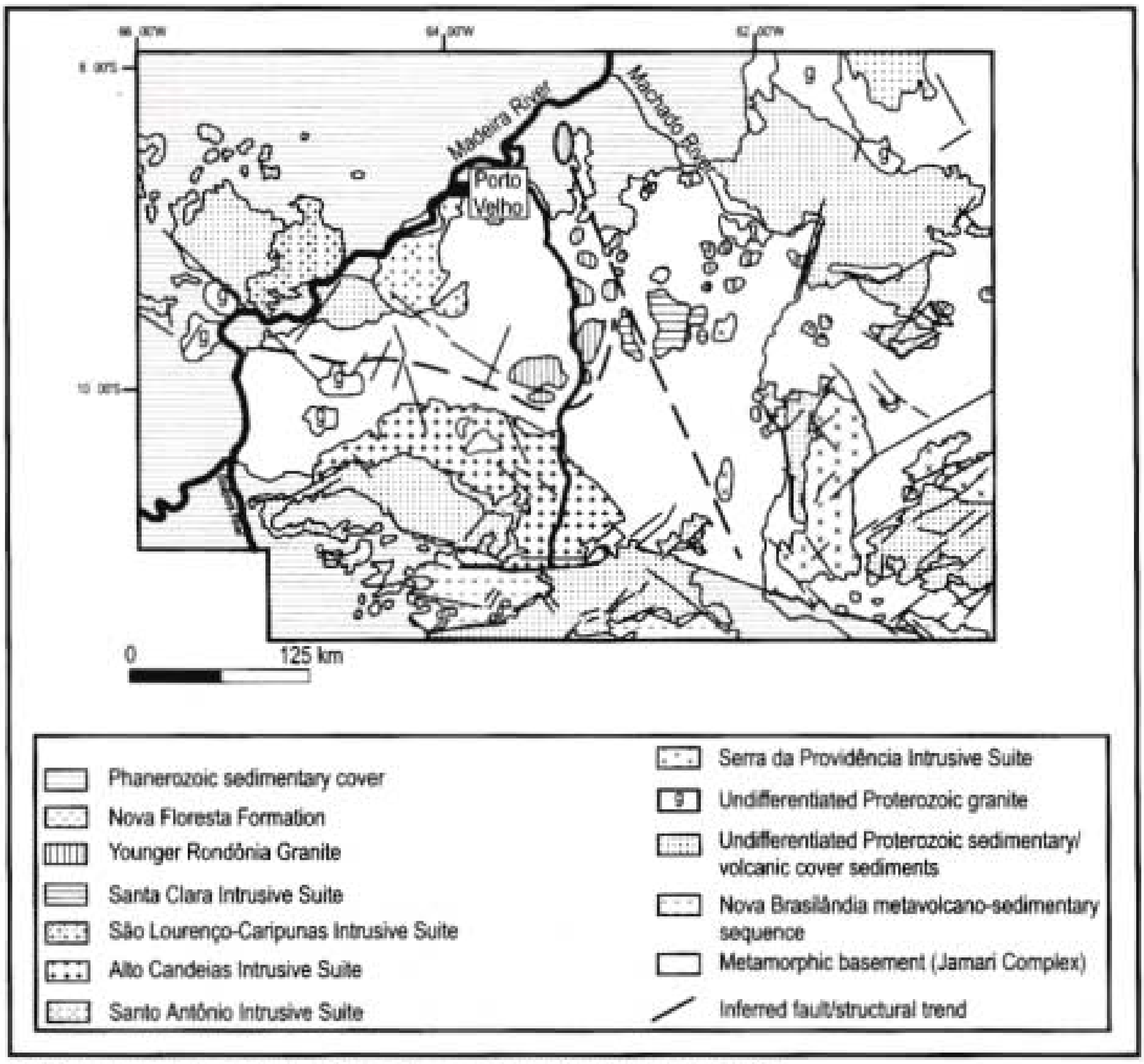


FIGURE 14 - Geological map of the Rondônia Tin Province (modified after Bettencourt et al., 1997).

The Alto Jauru District

The Alto Jauru gold district is situated in the State of Mato Grosso, Brazil, at the southeastern extremity of the Alto Guaporé Gold Province (Fig. 13) and is included in the Rio Negro-Juruena Belt (1.8-1.5 Ga). It consists of three volcano-sedimentary belts striking N25°W, separated by bands of granite-gneiss known from E to W, as Cabaçal, Araputanga and Quatro Meninas, constituting the Alto Jauru Greenstone Belt (Monteiro *et al.*, 1988). There is a Au-Cu-Ag deposit at Cabaçal. The volcano-sedimentary sequence may be divided into three units: (1) a basal unit consisting of mafic-ultramafic volcanic rocks of the Mata Preta Formation; (2) an intermediate unit consisting of acid volcanic rocks of the Manoel Leme Formation; (3) an upper sedimentary unit assigned to the Rancho Grande Formation. Gneissified plutonic rocks of tonalitic composition intrude this sequence. The age of these rocks was determined to be between 2.0 and 1.7 Ga (Geraldes *et al.*, 1996; Pinho, 1996). The Cabaçal gold deposit is associated with a volcano-sedimentary belt that developed in an island arc setting (Pinho *et al.*, 1997). It hosts a sequence of tuff units and intercalated volcaniclastic rocks with chert zones of the Manuel Leme Formation (Monteiro *et al.*, 1988). Three main types of mineralization have been recognized: associated with a shear zone; volcanogenic massive sulphide (VMS) type; and disseminated mineralization in tonalite (Pinho *et al.*, 1997). The third type is disseminated, banded, veined, brecciated and massive, and is composed of sulphide species such as chalcopyrite, pyrite, pyrrhotite, marcasite, sphalerite, cubanite, galena and molybdenite. In addition, there are minerals with selenium, tellurium, Au-Ag and Au-Bi compounds (Pinho, 1996; Pinho *et al.*, 1997). The total reserves of the Cabaçal Deposit are about 1.9 Moz Au, 0.6 Moz Ag and 43 000 t Cu (Souza, 1988), partially mined during the last decade. The mine is now closed.

The Alto Guaporé Gold Province

This province is situated on the boundary between Brazil and Bolivia in the upper reaches of the Guaporé River, in the southern part of the Amazonas Craton (Fig. 13). There occurred between 1.2 and 1.0 Ga the development of the Sunsas passive margin, representing a zone of oceanic expansion between Amazonia and Laurentia, as well as the development of the Aguapeí Aulacogen by intracontinental rifting (Sães and Fragoso Cesar, 1994; Sães, 1999). The closure of this basin by intercratonic collision of the Amazonian Craton and the Grenville Province brought about the formation of the Grenville-Sunsas collision belt, the inversion of the Aguapeí Rift, and the amalgamation of the Rodinia Supercontinent at 1.0 Ga.

The tectonic deformation related to the Sunsas Event is reflected by the development of an extensive shear belt of dextral character striking N20°W, which affected especially, the central zone of the Aguapeí Rift, to which are associated the main gold deposits and occurrences.

The gold mineralization is associated with quartz veining: Laurinha, Pau-a-Pique, São Francisco Xavier and São Vicente, constituting the Alto Guaporé Gold Province (Sães *et al.*, 1991; Silva and Rizzotto, 1994; Geraldes *et al.*, 1996; Sães, 1999). This veining generally occurs at the contact between the sediments of the Aguapeí Group and the granite-gneiss of the basement, and locally in the rocks of the Aguapeí sequence. The gold ore consists of quartz, pyrite and gold, accompanied by magnetite, chalcopyrite, galena and arsenopyrite. The hydrothermal alteration associated with the mineralization involved silicification, chloritization and sericitization which has been dated at between 960 and 910 Ma by K/Ar. The deposits are small with grades between 0.6 and 20 g/t Au, and reserves amounting to some tonnes of gold. Gold mining in the Serra de São Vicente has been going on for almost a century. The gold occurs in placer deposits formed from the erosion of sediments and quartz veins of the Aguapeí Group. The total reserves in alluvial-coluvial material are given as 0.14 Moz at an average grade of 0.14 g/m^3 Au. The gold reserves of the Alto Guaporé Province exceed 90 t Au.

In this province there also occurs mineralization associated with volcano-sedimentary units of the San Ignacio Group. However, the most important occurrence is situated to the NW, outside the province at Serrania San Simon, Bolivia, near the Guaporé River (Fig. 13). At this locality the gold mineralization is associated with the San Ignacio tectonic event (1.4-1.3 Ga) and occurs in hydrothermal gold-quartz space-filling lodes in greywacke, sandstone and conglomerate beds of the low-grade metamorphic San Ignacio schist belt. It is estimated that some 4 t of gold have been mined from vein and secondary deposits (Litherland *et al.*, 1986).

The Rondônia Tin Province

According to Bettencourt *et al.* (1997) cassiterite was discovered in 1952 in the then Territory of Rondônia, and has since been intensively mined by *garimpeiros* and mining companies. The total production up to 1995 has been estimated at about 220 000 t Sn. Present production is about 7500 tpa from the still active Bom Futuro and Santa Bárbara mining districts.

In the Rondônia Tin Province (Fig. 14) the Sn deposits and associated metals (W, Nb, Ta, Cu, Zn, Pb) are spatially related to the final phases of the rapakivi anorogenic granite intrusives of São Lourenço-Caripunas (SLC: 1.3 Ga) and the Younger Rondônia Granites (YRG: 950 Ma). These two suites are correlated with the development of the Sunsas-Aguapeí Orogen that occurred between 1.30 Ga and 900 Ma. The SLC is related to an extensional phase that preceeded the orogenic phase, whereas the YRG are interpreted as a distal manifestation of the orogenesis.

The YRG occur mainly in the Massangana, Ariquemes, São Carlos, Caritianas, Pedra Branca, Santa Bárbara and Jacundá massifs, situated in the central part of the province. The intrusive suite displays three distinct intrusive granite phases:

(a) an early phase, only observed at the Massangana Massif, consisting of coarse-grained biotite syenogranite with subordinate hornblende and accessory minerals that include zircon, apatite, ilmenite, magnetite and fluorite.

(b) an intermediate phase consisting essentially of fine to medium-grained syenogranite and alkali-feldspar granite with biotite, and locally with hornblende. The most common accessory minerals are zircon, monazite, ilmenite and fluorite.

(c) A late phase consisting essentially of topaz, Li-mica-albite granite and topaz-quartz-feldspar porphyry.

The tin mineralization and associated metals are spatially associated with the two last phases, occurring mainly in the form of Li-mica-albite granite with disseminated cassiterite and less abundant quantities of columbite-tantalite; pegmatite with topaz, beryl, cassiterite and subordinate columbite-tantalite; greisen bodies with cassiterite; quartz veins with cassiterite and wolframite; and quartz veins with Cu-Pb-Zn-Fe sulphides.

The São Francisco Craton

The São Francisco Craton (Fig. 15) (Alkmin *et al.*, 1993; Almeida, 1977) started its evolution in the Archean and ended at the close of the Mesoproterozoic. It is delimited by Neoproterozoic mobile belts related to the Brasiliano Cycle: Brasilia, Rio Preto, Riacho do Pontal, Sergipano, and Araçuaí (Fuck *et al.*, 1993).

The São Francisco Craton may be divided into three main compartments (Barbosa, 1997) the strike of which is broadly N-S: Eastern, Central and Western that are exposed over extensive areas of the states of Bahia, Minas Gerais and Goiás. The limits of these compartments correspond to large tectonic features, also with N-S strike: the tectonic limits of the marginal coastal basins; the Contendas-Jacobina Lineament, the Espinhaço Lineament; and the limit of the São Francisco Craton with the Brasília Belt.

These three compartments contain important deposits, of which the most important are associated with:

- Granite-greenstone terranes: Au and Mn of the Rio das Velhas greenstone belt (Quadrilátero Ferrífero); gold of the Rio Itapicuru greenstone belt; magnesite of the Brumado greenstone belt; barite of the Mundo Novo greenstone belt;
- Mafic-ultramafic complexes: Fe-Ti-V of the Jacaré and Campo Alegre de Lourdes sills; chrome at Rio Jacurici and Campo Formoso, and copper at Rio Curaçá/Caraiba;
- Exhalative sedimentary sequences: Fe of the Minas Supergroup (Quadrilátero Ferrífero); Pb-Zn at Boquira; Fe-Mn at Urandi-Licinio de Almeida;
- Sedimentary sequences: Au-U-Py at Jacobina and Moeda; diamond at Diamantina and on the Chapada Diamantina.

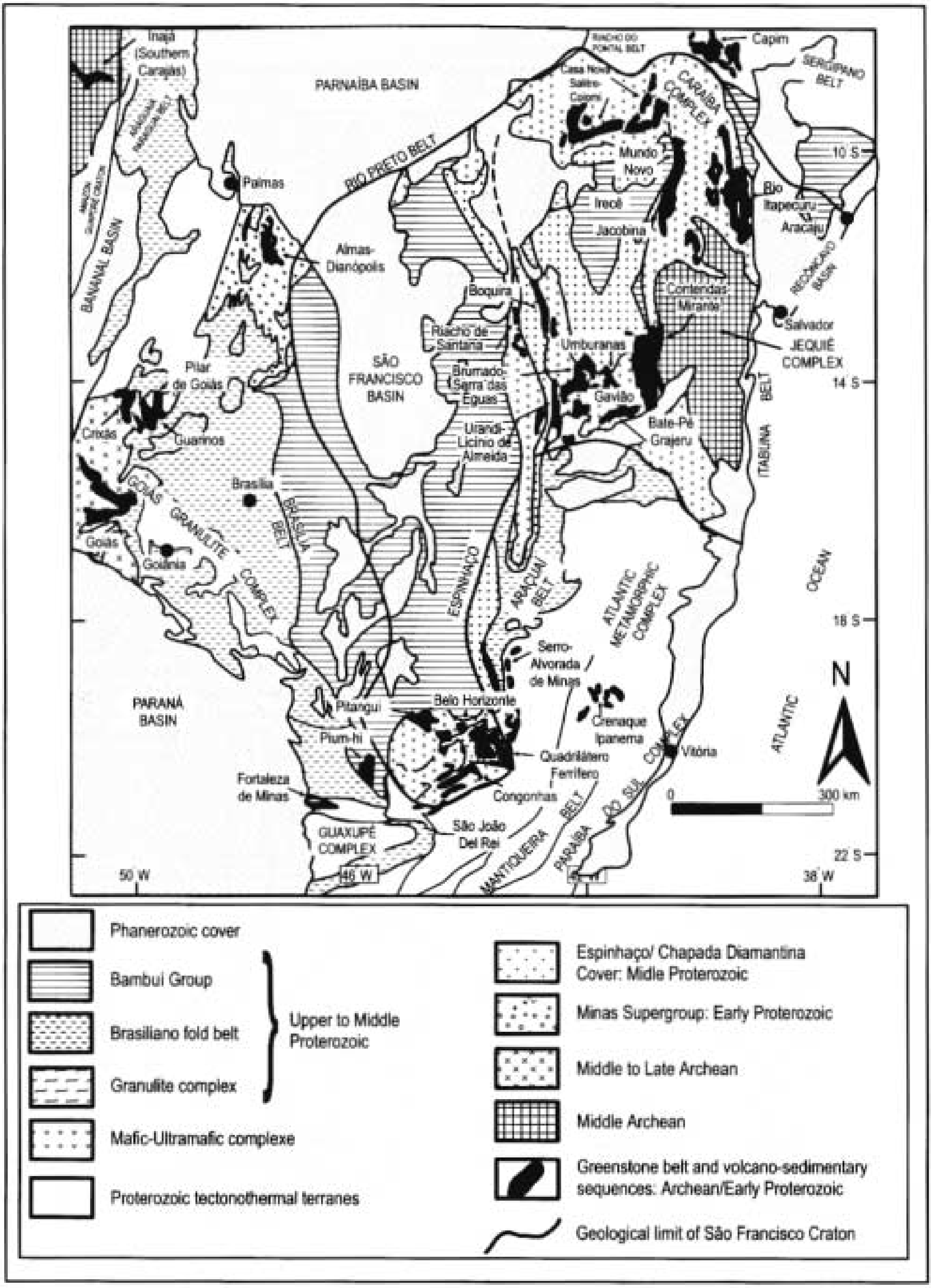
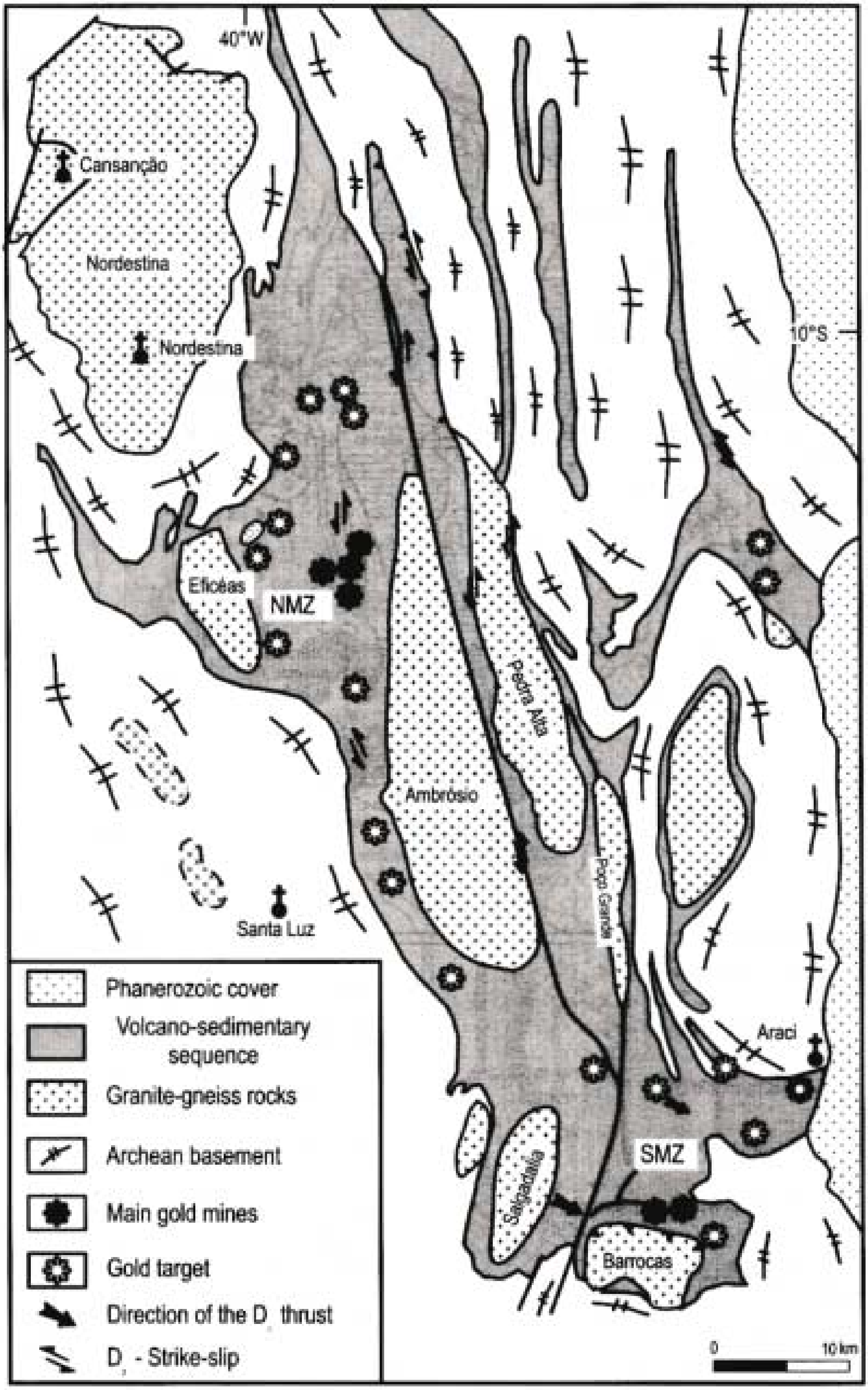


FIGURE 15 - Geological map of the São Francisco Craton and surrounding areas.



FIGURES 16 - Simplified geological map of the Rio Itapicuru Greenstone Belt. The North Mineralized Zone (NMZ) and the South Mineralized Zone (SMZ) are indicated (modified after Silva and Villas, 1998)

The Eastern Compartment

The Eastern Compartment includes the Jequié and Serrinha blocks with extensions into the Macururé Domain of the Sergipano Belt, the Itabuna and Curaçá-Salvador belts, as well as with the Jacobina rift sequence.

The Itapura Barite Deposit

The Itapura barite deposit (Castro *et al.*, 1997) is situated near the town of Miguel Calmon and is not well known. It occurs as hydrothermal veins that cut the metasedimentary unit of the Mundo Novo greenstone belt. However, one can observe banded iron formation units and zones with stratiform barite, intercalated with beds of chert, which favour the hypothesis of an exhalative origin of a distal facies of a volcano-sedimentary sequence. In the Mundo Novo greenstone belt, defined by Mascarenhas and Silva (1994), also occur some gold showings and those of base metals (Cu-Pb-Zn) at Fazenda Coqueiro where disseminations and zones of massive sulphide (pyrite, pyrrhotite, chalcopyrite, sphalerite, galena and associated gold) can be observed in mafic and felsic volcanic rocks, intercalated with pelitic and chemical metasediments.

Rio Itapicuru Gold Province

The Rio Itapicuru greenstone belt (Fig. 16) (Davison *et al.*, 1988), metamorphosed in the greenschist facies, is hosted in the Serrinha Block, which is represented by the Santa Luz Complex. This complex of Archean age consists of gneiss, granodioritic orthogneiss, amphibolite and intrusions of differentiated mafic-ultramafic complexes containing chromite associated with the Pedras Pretas Deposit, situated near the town of Santa Luz. The volcano-sedimentary sequence consists of a basal mafic volcanic unit of tholeiitic composition of ocean-floor origin; an intermediate to felsic calc-alkaline volcanic unit, having the characteristics of a continental arc; and an upper sedimentary unit consisting of turbidite beds, chert and BIF units. It is intruded by syn to late tectonic Type I granitoid plutons. The evolution of the supracrustal and granitoid rocks occurred between 2.2 Ga (basalt) and 2.0 Ga (syn-tectonic granitoid) as determined by Sm/Nd, Pb/Pb, U/Pb geochronology in zircon, and by Rb/Sr (Silva and Cunha, 1999). Important gold concentrations are found in the Fazenda Maria Preta District to the N of the Rio Itapicuru greenstone belt and in the Weber Belt, situated to the S of the Rio Itapicuru greenstone belt (Kishida *et al.*, 1991).

The Fazenda Maria Preta deposits (Coelho and Freitas-Silva, 1998; Alves da Silva *et al.*, 1998) are found in three second order sinistral shear zones, the attitude of which is N-S/50°70°W. These zones lie parallel to the main shear zone farther to the E that affected the Maria Preta unit, which consists of andesitic lava flows intercalated in pyroclastic rocks and metasediments. The main mineralized zones are associated with quartz veins parallel to the mylonitic foliation and to breccia and stockworks of the wall rock. Pronounced veining, silicification and carbonatization attest to the intense circulation of hydrothermal fluids along the shear zone, resulting in the growth of quartz, carbonate, albite, and the precipitation of sulphides (pyrite, arsenopyrite, pyrrhotite), besides sericite and chlorite. The gold is free in the quartz and in the sulphide minerals. The deposits of the Fazenda Maria Preta district are not important economically. The reserves are estimated at 12.5 t of gold. The deposit was mined by open-pit methods to a depth of 100 m.

The deposits of the Weber Belt (Santos *et al.*, 1988;Teixeira *et al.*, 1990; Reinhardt and Davison, 1990; Alves da Silva *et al.*, 1998) are found to the S in an E-W bend that corresponds to a zone of overthrusting, resulting in the overturning of the stratigraphic units, which from S to N are: the Incó unit, consisting of carbonate-chlorite schist representing basalt; the Fazenda Brasileiro unit, bearing the most important gold mineralization; the Canto unit, containing pelitic-carbonaceous sediments intercalated with pyroclastic volcanic rocks; and the Abóbara unit, consisting of a thick sequence of basalts with thin sedimentary intercalations.

The Fazenda Brasileiro unit consists mainly of two zones of quartz-chlorite-magnetite-schist (magnetic schist) which host the gold mineralization. The upper magnetic schist zone is overlain by a sequence of graphitic schist. The magnetic schist zones have a thickness of 20m and 3m, respectively. Both magnetic schist zones are separate by an intermediate sequence of sericite-chlorite-carbonate-schist and plagioclase-actinolite-schist, the latter representing altered gabbroic bodies.

At the Fazenda Brasileiro Mine, the reserves of which are estimated at about 150 t of gold at 7 to 8 g/t Au, the mineralization is associated with zones of more or less graphitic quartz-chlorite-magnetite schist. The mineralized bodies may be up to 500 m long and 40 m wide, and they contain several generations of quartz veins with associate sulphide, locally forming a sulphide alteration halo with pyrite, pyrrhotite, arsenopyrite, carbonate and albite. Arsenopyrite is the most important sulphide species on account of the fact that it is always associated with gold. However, at the other mines in the Weber Belt, the gold may be associated preferentially with pyrrhotite. The gold is free, and occurs in the intergranular contacts, microfractures, and as fillings in sulphide. The circulation of hydrothermal fluids that gave origin to the mineralization of the Weber Belt are related to D_2 and D_3 events, and were for the best part channeled along shear zones resulting from the D_1 deformational phase.

The Rio Jacurici and Campo Formoso Chromite Districts

The chromite deposits of Medrado and Ipueira, situated in the Valley of the Jacurici River are associated with a mafic-ultramafic complex, some 7 km long and 300 m thick, intruded into the Salvador-Curaçá Belt. These deposits were initially interpreted as a stratified sill (Barbosa de Deus and Viana, 1982) in the contact between quartz-feldspathic granulite at the base, and a metasedimentary sequence at the top consisting of serpentinite-marble, diopsidite and metachert, forming a large synform the axis of which strikes N-S. The sill consists mainly of dunite, harzburgite, pyroxenite and gabbro (Fig. 17) in which there is intercalated a zone of cumulate chromite having an average thickness of 5 to 8 m, locally attaining 15 m. The chromite grains are subhedral and fine-grained, having an average diameter of 0.4 mm. The chromite is rich in chrome (48.8% Cr2O3) with a low Cr/Al ratio of about 3.2. According to Marques (1999), the cryptic variation of the minerals along the sill suggests injections of a primitive magma during the formation of the main chromite zone. The chromite reserves of the Ipueira-Medrado Sill are estimated at about 8 Mt (Mello *et al.*, 1986).

The Campo Formoso Complex is about 40 km long and 900 m wide. It is intruded into the granulitic rocks of the Mairi Block, and cut by the Campo Formoso Granite, dated at 2.0 Ga. The complex is covered discordantly by rocks of the Jacobina Group. From base to top it consists of actinolite gneiss, tremolite-actinolite serpentinite and serpentinite-chlorite-carbonate-talc schist; the two upper units contain seven chromitite beds with massive, disseminated and stringer ore (Barbosa de Deus *et al.*, 1982). The chromite grains are euhedral, and have an average diameter of 1 mm. They display a network texture, and locally an olivine-fill texture. The chromite is rich in Cr_2O_3 (up to 60%), and has high Cr/Al (*c.* 6.5) and Cr/Fe (*c.* 3.0) ratios.

The emerald deposits known in the Campo Formoso area are associated with a granite intruded at 2.0 Ga into the ultramafic rocks of the complex (Giuliani *et al.*, 1993).

The Rio Curaçá Copper District

The Caraíba copper deposit (D'el Rey Silva and Oliveira, 1999; Lindenmayer, 1981) is situated in the valley of the Curaçá River. It is associated with a mafic-ultramafic intrusive complex in the high-grade metamorphic Curaçá-Salvador Belt generated by the collision of the Serrinha and Mairi continental blocks at about 2.0 Ga. The sequence consists of gneiss with intercalations of amphibolite, paragneiss, BIF units, calc-silicate rocks, olivine marble and quartzite at the base; gabbro, gabbronorite, leucogabbro, peridotite, olivine-pyroxenite, hypersthenite rich in Cu, melanorite and norite in the intermediate unit; and migmatitic gneiss with syn-tectonic granite intrusions (tonalite and granodiorite) in the upper unit. The structure of the complex is vertical following polydeformation and polymetamorphism, in addition to shearing along ductile structures striking NNW-SSE and NNE-SSW. The mineralization consists essentially of chalcopyrite and bornite, which are disseminated or form massive irregular bodies in the coarse-grained hypersthenite, concordant with the mafic and metamorphosed sequence of the wall rock. The mineralized sill is attributed to successive intrusions of mantle-derived material before or during the D_1 deformation. The entire tectonic evolution is - considered to have occurred between 2.2 and 1.9 Ga. Mining since 1978 in open-pit and underground workings, the Caraíba Deposit has produced up to 1998 about 600 Mt of ore at 1.6% Cu.

To the N, in the Macururé Block, the Serrote da Laje Cu-Ni-Co deposit, situated in the proximity of Arapiraca (Alagoas), shows Cu-Ni-Co sulphide mineralization (Figueiredo, 1992; Horbach and Marimon, 1988).

The Jacobina Au-U-Py Deposit

The Jacobina gold deposit (Fig. 18) was mined continuously between 1973 and 1996. The ore-zone is hosted in a thick sequence of clastic metasediments assigned to the Jacobina Group of vertical attitude overlying the basement consisting of older granite-gneiss. The Jacobina Group is divided into three formations: The Serra do Córrego at the base, consisting of conglomerate and quartzite beds that host the gold-bearing zones; followed by an intermediate unit known as the Rio do Ouro Formation, consisting mainly of quartzite; and the Cruz das Almas Formation at the top, formed mainly of pelite beds. The Au-U-Py mineralization is associated with the conglomerate beds of the Serra do Córrego Formation. The more important zones are known as the Basal Reef, Main Reef, João Belo and Canavieiras, in which the gold grade varies between 2 and 100 g/t Au (Cox, 1967; Molinari, 1982; Molinari and Scarpelli, 1988; Scarpelli, 1991). The gold is associated with pyrite and may constitute up to 30% of the matrix of the conglomerate and with brannerite/uraninite. The most common accessory minerals are tourmaline, zircon, thorite, magnetite and chromite. The gold mineralization is considered to be a paleoplacer of the Witswaterrand-type and has been strongly affected by metamorphism and by percolating waters along shear zones, following the classification of a paleoplacer-type, modified as proposed by Ledru and Bouchot (1993).

The Central Compartment

In the Central Compartment are included the Gavião and Paramirim Blocks with extension to the N into the Sobradinho Domain; the Umburana, Ibitira-Ubiraçaba, Brumado, Guajeru and Contendas-Mirante granite-greenstone terranes; the Boquira and Licinio de Almeida-Urandi sedimentary-exhalative sequences, as well as the Mesoproterozoic cover of the Chapada Diamantina and of the Northern Serra do Espinhaço.

The Gold and Base Metals Occurrences Associated with Archean Greenstone Belt Sequences

In the Gavião Block occur several supracrustal volcano-sedimentary sequences of the greenstone belt-type: Umburana (UGB), Ibitira-Ubiraçaba, Brumado (BGB), Guajeru. The Contendas-Mirante (CMGB), which is situated in the ambit of the Jacobina-Contendas-Mirante Lineament is here incorporated in the Central Block in the interests of clarity. Numerous gold and base metals showings (Cu, Pb, Zn) are known in the UGB and CMGB, which are being prospected (Silva and Cunha, 1999).

The Jacaré Sill Fe-Ti-V-Pt Deposit

The stratified Jacaré Sill, dated at 2.9-2.8 Ga (Brito *et al.*, 1999) or 2.4 Ga (Marinho, 1991) is intruded into the Contendas-Mirante greenstone belt along the Jacobina Lineament (Galvão *et al.*, 1986; Brito, 1984), and represents

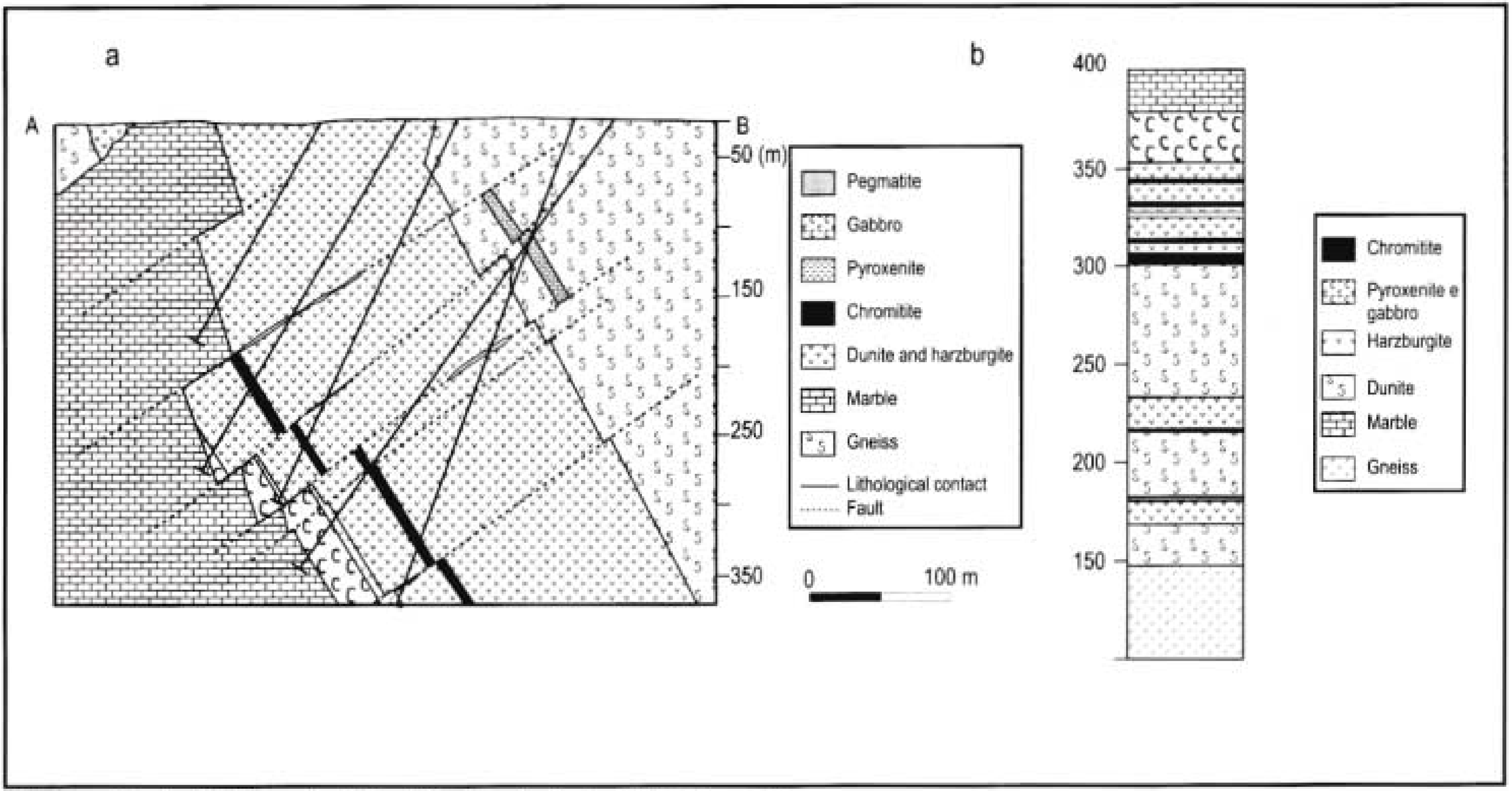


FIGURE 17 - a - Schematic geological cross-section of Ipueira Sill; b - Stratigraphic column (modified after Marques,1999).

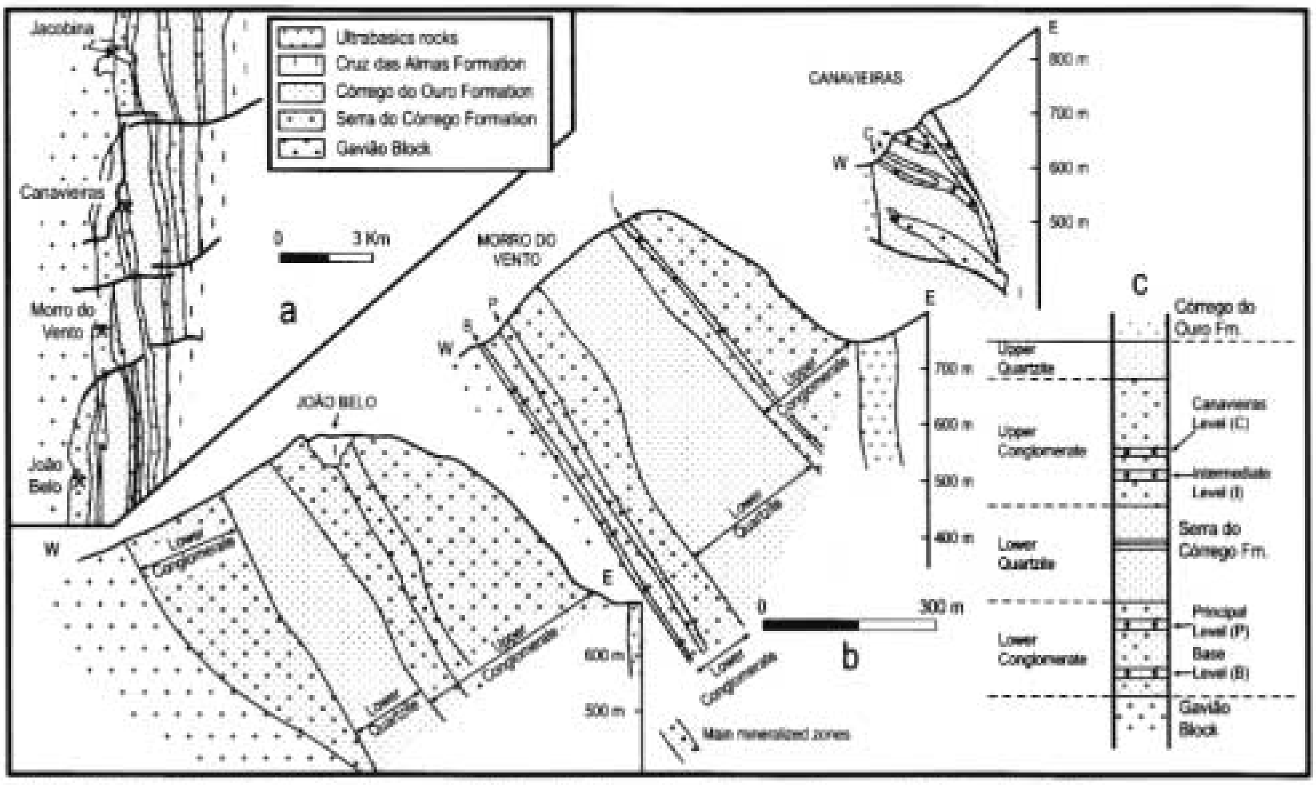


FIGURE 18 - Gold mineralization of Jacobina. a - simplified geological map; b - main cross-sections; and c - lithostratigraphic column.

a potential target for Fe-Ti-V mineralization. The Fe-Ti-V deposit at Fazenda Gulçari is associated with the Jacaré Sill. The deposit is divided into two zones: a Lower Zone and an Upper Zone. The Lower Zone (400 m) consists of gabbro, gabbronorite and anorthosite of generally coarse-grained and massive aspect without magnetite. The Upper Stratified Zone (600 m) consists of medium to coarse-grained gabbro with rhythmic layering in the Lower Member, manifest by alternations of magnetite and pyroxene at the base, grading gradually to melanogabbro with pyroxenitic bands in the Central Member; and by layered gabbro intercalated with zones of pyroxenite and magnetitite in the Upper Member.

The Fazenda Gulçari Fe-Ti-V deposit is hosted in the Lower Member of the layered part of the sill. It has an oval pipe-like shape (400 m x 150 m), displaying broadly concentric zonation, with an external aureole of hornblende grading to pyroxenite, magnetite-pyroxenite and magnetitite at the centre. Pegmatoid structures are often observed in the several aureoles. There occur two types of ore: massive and disseminated. The mineralization consists mainly of titanomagnetite, ilmenite and ulvospinel. The ilmenite forms discrete grains or ribbon-like exsolution structures composed of magnetite. Disseminated sulphide (>1%) is present. The gangue mineral consists of diopsidic augite. The reserves at Fazenda Gulçari are given as 6.1 Mt of ore at an average grade of 1.27% V_2O_5. Important PGE anomalies have been defined in association with magnetitite and are investigated.

More or less of the same age is the Campo Alegre de Lourdes Sill Fe-Ti-V deposit, situated in the Sobradinho Domain, N of the Central Compartment in the State of Bahia (Sampaio *et al.*, 1994).

The Serra das Éguas Magnesite Deposit

The Serra das Éguas magnesite deposit (Bodenlos, 1960) is associated with the Archean volcano-sedimentary sequence of the Brumado greenstone belt that overlies the basement rocks of the Gavião Block. The sequence consists of three units: the Lower Ultramafic Unit (200 m) consisting of ultramafic flows intercalated with siliceous-carbonate and carbonate rocks; an Intermediate Unit (500 m) consisting essentially of chemical sediments such as magnesitite and dolomite with intercalations of tuff and ultramafic flows; and an Upper Unit (*c.* 700 m) consisting of quartzite, ferruginous quartzite and itabirite with intercalated tuff and volcanic flows. The Serra das Éguas (Fig. 19) is situated near the town of Brumado, Bahia, and hosts the largest magnesite deposits in Brazil with reserves of about 150 Mt, and production of about 1.7 M tpa. The talc reserves are about 1 Mt, and the production is 30 000 tpa (Oliveira *et al.*, 1997). The main characteristic of the magnesite deposits is the sedimentary nature of the magnesitite, and the continuity and thickness of the beds, as well as the intimate association of these with dolomite. These factors favour an origin related to the chemical precipitation of magnesite, permitting a comparison of the Serra das Éguas Deposit to deposits of the Veistsch-type. However, the presence of submarine volcanism associated with sedimentation suggests the possibility that this volcanism may have contributed significantly to the supply of magnesium involved in the precipitation of magnesite.

The Boquira Pb-Zn Deposit

The Boquira Deposit (Fig. 20) is situated in the valley of the Paramirim River. The deposit was discovered in 1952,

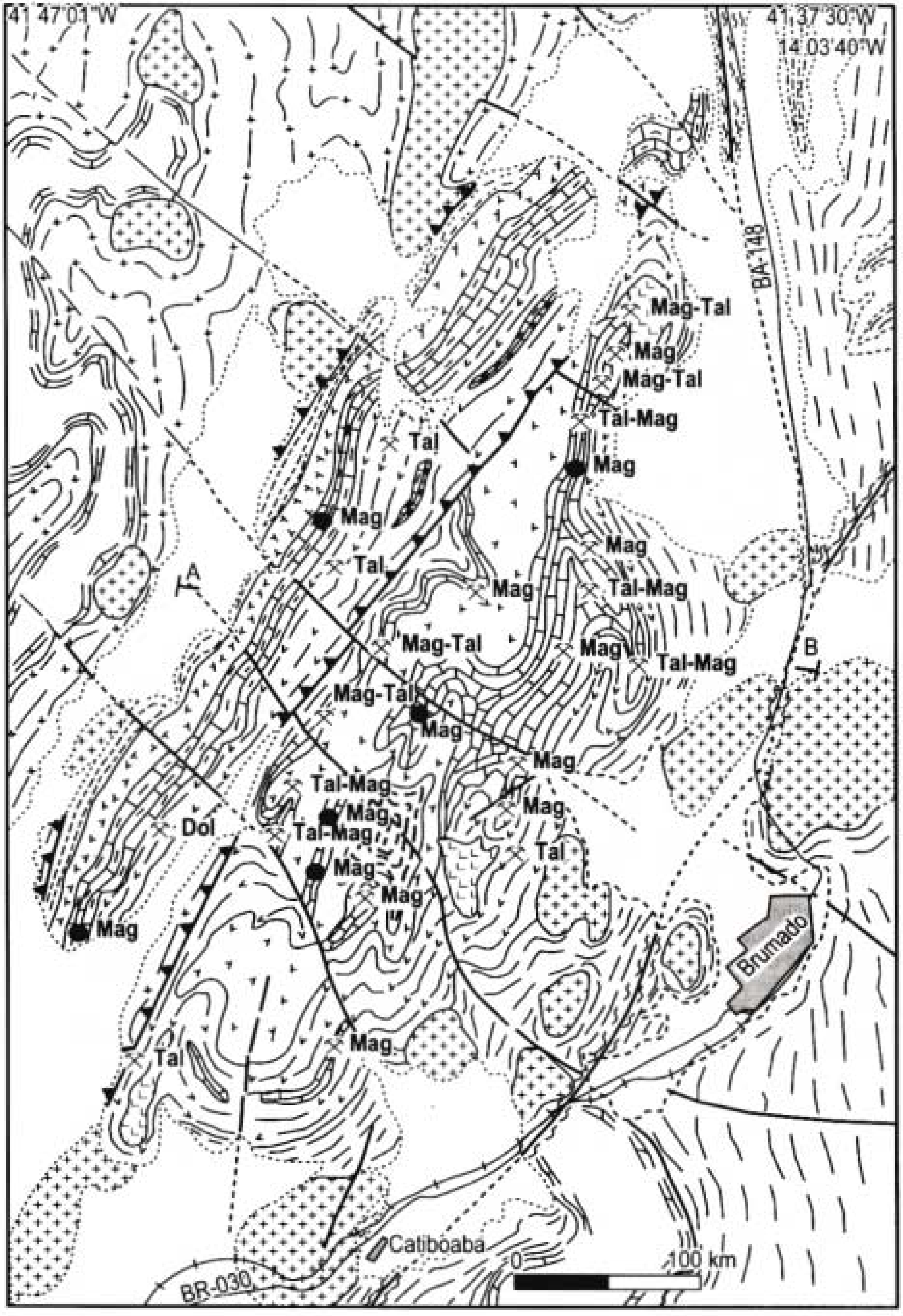


FIGURE 19 - Geological sketch map of the Serra das Éguas magnesite deposit: mag = magnesite; ta = talc (modified after Oliveira et al.,1997).

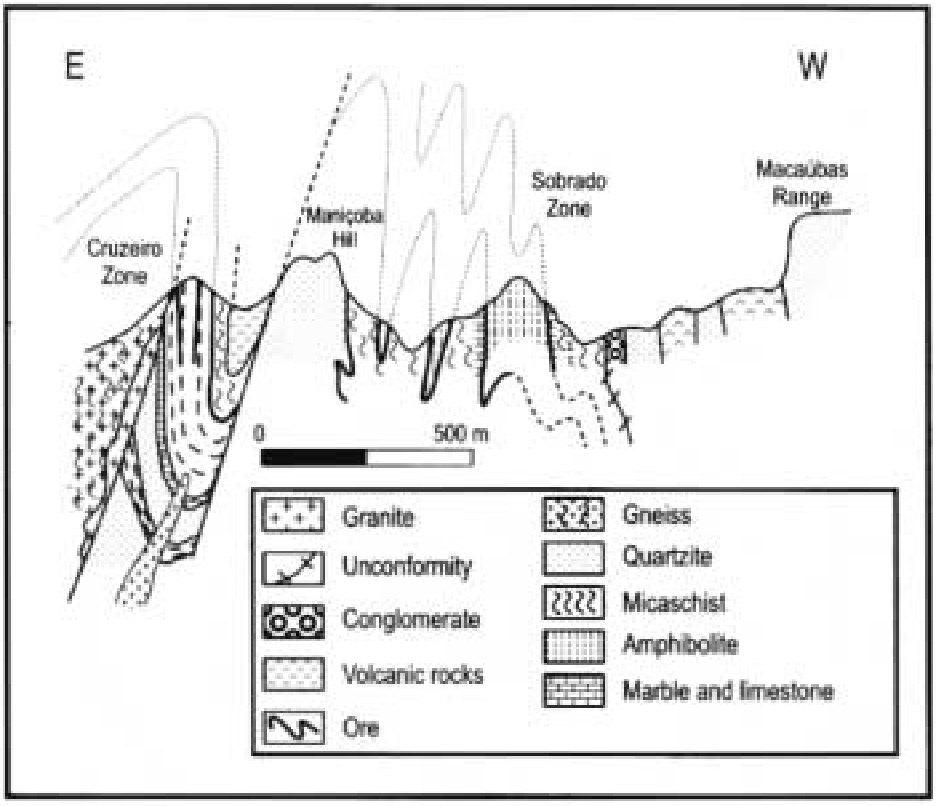


FIGURE 20 - Geological cross-section through the Boquira Mine (modified after Fleischer, 1976; Espourteille and Fleischer, 1988).

and between 1959 and 1992, mine production was about 650 000 t of Pb+Zn. The deposit is hosted in the Boquira Formation (Fleischer and Espourteille, 1999; Espourteille and Fleischer, 1988) consisting of quartzite, amphibolite, marble, BIF units and chlorite-garnet-biotite schist that pass transitionally to gneiss and migmatite of the Paramirim Block. The Boquira Formation, which is intruded by granite and pegmatite, disappears to the NW under the Mesoproterozoic cover of the northern Espinhaço range. The mineralized zones have been intensely folded and sheared, and display a characteristic banding that appears as quartz-magnetite, silicate-magnetite and carbonate-silicate amphibolite facies (Rocha, 1985). The mineralization has been dated at 2.7 Ga by Pb/Pb, and consists essentially of galena, rich in silver (up to 260 g/t Ag), and sphalerite associated with pyrite and subordinate chalcopyrite. Near the surface the mineralization is affected by weathering to a depth of 20 m with the formation of oxides such as cerussite, smithsonite, limonite and anglesite, along with pyromorphite, hemimorphite, hydrozincite, chrysocolla, bornite, covellite, malachite and smaller amounts of azurite.

The stratiform character of the mineralization, intimately associated with primary banding and the presence of banded iron formation units, favours a sedimentary-exhalative origin for the mineralization (Misi *et al.*, 1996, 1999; Carvalho *et al.*, 1997).

The Chapada Diamantina Diamond Deposits

The region around the towns of Lençóis, Andaraí, Mucugê, Xique-Xique, in the Chapada Diamantina (diamond tableland), in the State of Bahia, has been famous for its diamond production since the last century. Mesoproterozoic diamondiferous conglomerate occurs (a) in the Tombador and Morro do Chapéu formations of the Chapada Diamantina Group (*c.*1.2-1.0 Ga) and (b) in weathered alluvium and colluvium as the product of the erosion of the conglomerate beds by river action. The former has been worked mainly by *garimpeiros*.

The diamond fields have a large distribution throughout the Chapada Diamantina reaching about 300km from one extreme to another. Five different diamond fields are distinguished: Lençóis-Andaraí-Mucugê, Santo Inácio, Piatã-Serra do Bastião, Chapada Velha and Morro do Chapéu. With the exception of Morro do Chapéu, all the other fields are related to the Formação Tombador, the most important diamond-bearing unit of the Chapada Diamantina.

A characteristic feature of the Chapada Diamantina is the abundance of *carbonado* associated with gem diamond. This region has been mainly a producer of *carbonado*, which on average has larger dimensions than of the gem diamond. The color of the carbonado varies from dark grey to black. One *carbonado* weighing 3,167 ct, and believed to be the largest in the world, was found in 1895 near the town of Lençóis. Gem diamond is small, usually weighing less than one carat. The largest concentrations of diamonds occur in alluvial placers along the Paraguaçu, Santo Antonio and São José rivers with estimated reserves exceeding 1.5 M carats. The diamond reserves of the whole Chapada Diamantina are estimated at 3.8 M carats. The source of the diamonds remains unknown. Paleocurrents of the Tombador Formation, measured in the Lençóis-Andaraí-Mucugê field, indicate an extra-basinal origin for the diamonds coming from E-NE (Sá *et al.*, 1982; Sampaio *et al.*, 1994; Misi and Silva, 1996; Schobbenhaus, 1996).

The Lagoa Real Uranium Deposit

The Lagoa Real uranium deposit (Oliveira *et al.*, 1985; Forman and Waring, 1981) is situated near the town of Caetité (BA), to the E of the Serra do Espinhaço. The underlying rocks are Archean granite-gneiss and migmatite (Lagoa Real Complex) is intruded by several porphyritic granite bodies known as the São Timóteo Granite, dated at 1.74 Ga (Turpin *et al.*, 1988; Cordani *et al.*, 1992; Pimentel *et al.*, 1994). The gneiss of the Lagoa Real Complex and the São Timóteo Granite are cut by regional shear zones, along which occur a number of lenticular albitite bodies that host the uranium mineralization (Geisel Sobrinho *et al.*, 1980; Brito *et al.*, 1984; Lobato *et al.*, 1983; Raposo *et al.*, 1984). These bodies are distributed along two main lineaments that show that the shearing, sodium metasomatism and mineralization are contemporaneous (Lobato and Fyfe, 1990; Lobato *et al.*, 1998).

The metasomatic albitite forms lenticular bodies, the length of which varies from metres to kilometres, with thickness varying from a few centimetres up to a hundred metres. The bodies plunge along the lineation of cataclastic origin and form elongate ore shoots that may be 850 m long at depth, such as occur at the Rabicha Deposit. The mineralization consists mainly of uraninite in the form of microcrystalline and microgranular crystals (F » 0.023 mm). The age of the mineralization that has been dated at about 1.5 Ga (Turpin *et al.*, 1988; Cordani *et al.*, 1992) is related to a hypothetical Espinhaço event. The same mineralization was dated at 960 Ma by U/Pb in titanite (Pimentel *et al.*, 1994), undergoing recrystallization and remobilization at about 500 Ma, occurring during the Brasiliano tectono-thermal cycle.

The Lagoa Real District is the most important uranium district in Brazil with reserves given as 93 190 t U3O8.

The Western Compartment

In the Western Compartment are situated the Almas-Dianópolis, Guanambi-Correntina and Quadrilátero Ferrífero blocks, the granite greenstone terranes and volcano-sedimentary sequences of the Riacho de Santana, São Domingos and Rio das Velhas, the sedimentary-exhalative sequences of the Minas Supergroup and the Espinhaço sedimentary sequences.

The Occurrences of Gold and Base Metals of the Volcano-Sedimentary Sequence of Riacho de Santana

In the Guanambi-Correntina Block, the Riacho de Santana volcano-sedimentary sequence, considered to be a greenstone belt (Silva and Cunha, 1999; Lobato *et al.*, 1999), are found the most promising anomalies for gold and base metals in association with gossan (1.3% Cu and 2 to 5 g/t Au) overlying metatuff, chert and carbonate beds.

The Gold Deposits of the Almas-Dianópolis Volcano-Sedimentary Sequences

These sequences that occur on the São Francisco Craton to the N of the Brasilia Fold Belt are considered to be Paleoproterozoic greenstone belts (2.2 Ga). They contain numerous gold occurrences associated with volcanic rocks and intercalated banded iron formation units. The largest gold deposit (Córrego Paiol Mine) is situated near the town of Almas in the State of Tocantins where the gold is associated with a shear zone and intense hydrothermal alteration that affected the mafic rocks of the sequence (Cruz, 1998; 1993; Lobato *et al.*, 1999).

The Quadrilátero Ferrífero Au, Fe, Mn Province

The Quadrilátero Ferrífero Mineral Province (Fig. 21) is situated in the southern part of the São Francisco Craton, is well known for its deposits of gold, manganese and Cu-Ni-Co-Pt in the rocks of the Rio das Velhas Supergroup *(Archean), as well as for its deposits of iron and gold in the* Minas Supergroup (Paleoproterozoic).

The Mineral Deposits of the Rio das Velhas Supergroup

The Rio das Velhas Event (2.78-2.70 Ga) was coeval with the development of the Rio das Velhas greenstone belt at 2.772 Ga, and with the intrusion of tonalite, granodiorite and granite between 2.78 and 2.77 Ga (Noce, 1995; Noce *et al.*, 1998). The final phase of Archean cratonization is marked by granite intrusions at 2.612 Ga, and metamorphism at 2.61-2.59 Ga (Romano, 1989; Romano *et al.*, 1991; Machado and Carneiro, 1992). The Rio das Velhas Supergroup (Dorr, 1969) is divided into the Nova Lima and Maquiné groups. At its base, the Nova Lima Group consists of mafic and ultramafic volcanic rocks, including komatiite and tonalite associated with banded iron formation units of the Algoma-type, phyllite with chlorite and graphite, greywacke, felsic volcanic rocks and pyroclasts, all metamorphosed in the greenschist facies. At the top of the sequence the Maquiné Group consists mainly of metasediments consisting of conglomerate, quartzite, phyllite and greywacke, being subdivided into the Palmital and Casa Forte formations.

The Nova Lima Group gold deposits

The most famous of all the gold deposits in Brazil are associated with the rocks of the Nova Lima Group: Morro Velho (> 470 t Au), Raposos (> 40 t Au), São Bento (> 80t Au), Faria, Bicalho, Bela Fama, Brumal (> 30 t Au), Cuiabá (> 180t Au), Lamengo (> 10 t Au) (Ladeira, 1980, 1988, 1991; Ribeiro-Rodrigues, 1998; Ribeiro-Rodrigues *et al.*, 1996; Sales, 1998; Lobato *et al.*, 1998; Vieira and Oliveira, 1988; Abreu *et al.*, 1988).

The mineralized bodies have an elongate shape (Fig. 22), and they are controlled by the stretching lineation the strike of which coincides with the fold axes, and which is associated with ductile shear zones locally kilometres in

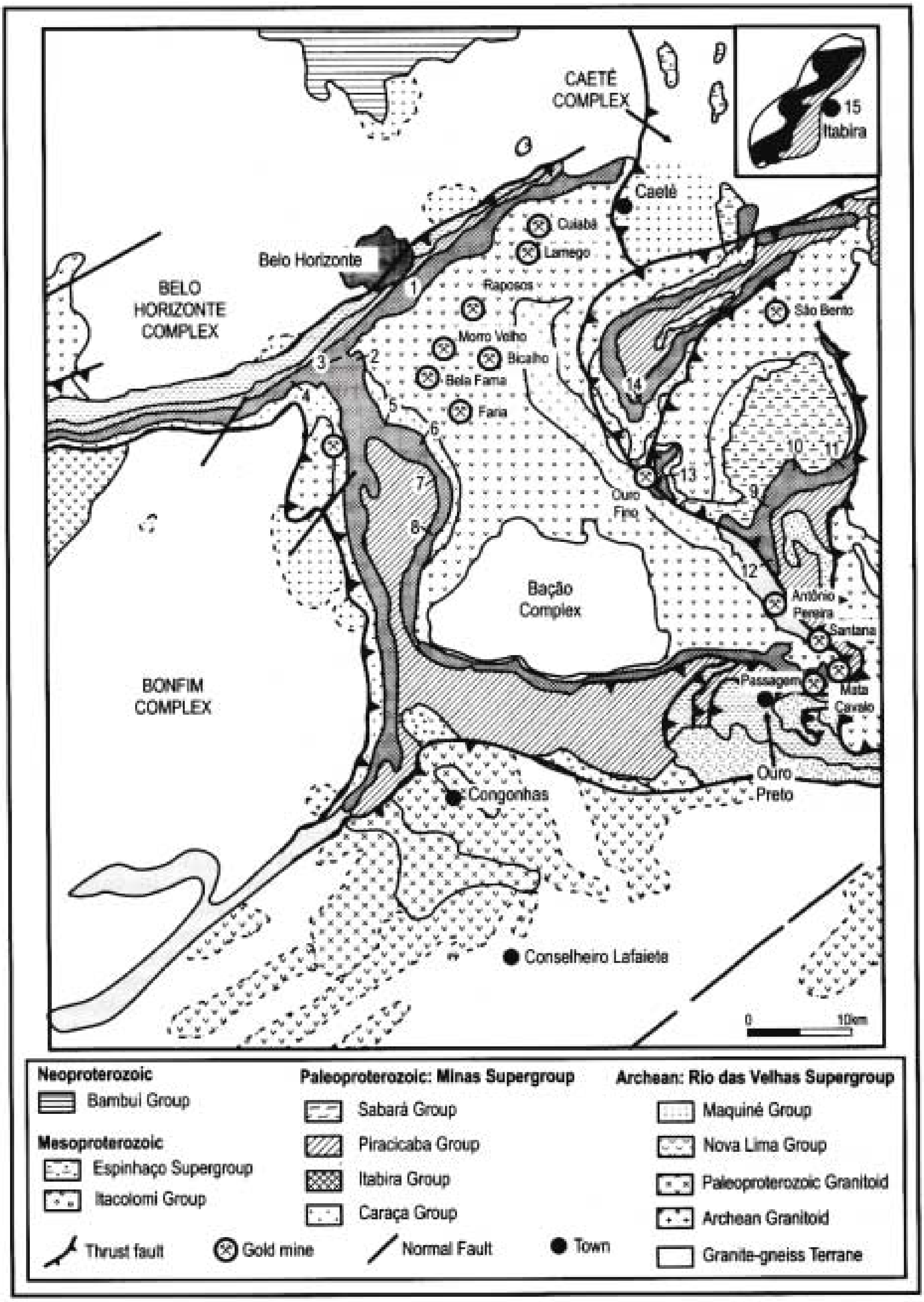


Fig. 21 - Geological map of the Quadrilátero Ferrífero (modified after Noce,1995; Pedrosa-Soares et al., 1994) showing main gold ore deposits and iron ore deposits. Gold deposits: ..Iron deposits: 1 - Águas Claras; 2 - Mutuca; 3 - Capão Xavier; 4 - Jangada/Samambaia; 5 - Tamanduá; 6 - Serra/Capitão do Mato; 7 - Abóboras; 8 - Pico; 9 - Miguel Congo; 10 - Alegria; 11 - Fazendão; 12 - Timbopeba;13 - Capanema; 14 - Gandarela, 15 - Itabira

length. The mineralization is concentrated in the hinges of the folds the axes of which are displayed parallel to the bedding. These shear zones are associated with intense hydrothermal alteration, attention to which was drawn by Vieira who also described these features in detail (Vieira, 1987, 1988, 1991; Lobato *et al.*, 1998). In the mafic and ultramafic volcanic rocks, the hydrothermal alteration is manifest in an external chloritization zone, an intermediate zone of carbonatization (ankerite), and an internal zone with sericitization, albitization, silicification, tourmalization and sulphidization. The sulphide mineralization occurs preferentially in the Algoma-type banded iron formation units that have as their principal feature magnetite-rich banding and/or siderite as may be seen at the Cuiabá, Lamengo, Raposos, São Bento deposits. Elsewhere, such as at Morro Velho, Bicalho, Bela Fama, the sulphide mineralization occurs in a generally massive carbonate zone (Lapa Sêca), locally banded and consists of siderite and ankerite together with quartz, albite, sericite and sulphide. However, some deposits and many mineralized zones in diverse deposits are associated with the development of shear zones in hydrothermally-altered schist, as may be seen at the Juca Vieira Deposit in which the sulphide mineralization occurs with sericitization and carbonatization, intimately related with quartz veining (Lobato *et al.*, 1998), but the main structural features are related to the Transamazonian Event.

The main point of discussion over the gold deposits of the Nova Lima Group is over the existence of syn-sedimentary stratiform mineralization associated with banded iron formation units prior to shearing; and the age of the shearing that gave origin to the hydrothermal mineralization: Archean (*c.* 2.6 Ga) or Transamazonian (2.0 Ga). Some consensus exists that the first and mineralizing tectonic event is of Archean age (DeWitt *et al.*, 1994; Thorpe *et al.*, 1984).

The Conselheiro Lafaiete Mn deposit.

In the region of Conselheiro Lafaiete-Ritápolis-Nazareno there are numerous occurrences associated with manganese found in the volcano-sedimentary sequence of the Barbacena Group, correlative with the Nova Lima Group of the Rio das Velhas Supergroup. The manganiferous ores are of two types (Pires, 1977, 1983): gondite, rich in silicate, and manganese with rodonite, spessartite, tefroite etc.; queluzite, rich in manganese carbonate such as rhodochrosite.

Taking for example the deposit of Morro da Mina, the mining operation initially started working the products of lateritic weathering, enriched in Mn (average grade 46% Mn), and consisting of cryptomelane and pyrolusite. After the oxide ore had been worked out, mining started in the carbonate-rich proto-ore (queluzite) having manganese grades between 30% and 37%. In 1997, the reserves at Morro da Mina were about 3 Mt, and the production was 150 000 tpa. The origin of the primary mineralization is attributed to the classical volcano-sedimentary model.

The Morro do Ferro greenstone belt Ni-Cu-Co-PGE + Au deposit

The Rio das Velhas Supergroup is also related to the greenstone belt sequences that occur to the W of the Quadrilátero Ferrífero. These sequences have been dated at about 3.0 Ga (Noce, 1995) and they are associated with the chromite deposit of the Pium-hi greenstone belt, the nickel laterite deposit at Morro de Niquel, and the O'Toole Ni-Cu-Co-PGE + Au deposit in the Morro do Ferro greenstone belt of the Fortaleza de Minas region.

The O'Toole Ni-Cu-Co-PGE + Au deposit of the Morro do Ferro greenstone belt (Brenner *et al.*, 1990; Cruz *et al.*, 1986; Teixeira *et al.*, 1987) is hosted in a unit of a komatiitic suite consisting of olivine peridotite, peridotite, pyroxenite and basalt, all metamorphosed in the greenschist facies. The basalt has a massive or layered aspect with olivine cumulate at the base and spinifex textures at the top, as well as pillow structures and breccia intercalated with tuff and banded iron formation units. Clinopyroxenite, amphibolite and BIF overlie the serpentinite body. The main ore-types are brecciated, disseminated, banded and stringer ore. The mineralization consists of pyrrhotite, pentlandite, chalcopyrite, cobaltite and PGM. The reserves are given as 6.6 Mt of ore at 2.2% Ni, 0.4% Cu, 0.05% Co, and 1.2 ppm PGM + Au. The O'Toole Deposit is similar to Ni sulphide deposits associated with Archean komatiite sequences. However, the mineralization occupies an unusual position in the ultramafic sequence, being situated in the upper part of the sequence, whereas it is more usual to find this type of deposit at the base.

The Deposits of the Minas Supergroup

The evolution of the Minas Supergroup (Dorr, 1969) (Fig. 21) probably began in Paleoproterozoic times with the deposition of the Caraça Group and the Cauê and Gandarela formations of the Itabira Group in basins that resulted from the rifting of the Archean Platform (Renger *et al.*, 1994). At the base of the Caraça Group, the Moeda Formation contains conglomerate beds with Au-U-Py Witwatersrand-type mineralization (Renger *et al.*, 1988; Minter *et al.*, 1990). The Gandarela Formation was dated at 2.42 Ga. The sedimentation of the Piracicaba Group defines a period of oceanic expansion and the subduction of the ocean crust manifest in the intrusion of the Maranhão Batholith at 2.124 Ga. The basin closed with crustal collision between 2.065 and 2.035 Ga during the Transamazonian Event (Marshak and Alkmin, 1989; Marshak *et al.*, 1992), and was succeeded by the deposition of the Itacolomi molasse. The Transamazonian Event brought about the individualization of the extensive Cinturão Mineiro (Teixeira, 1985) that lies along the southern margin of the São Francisco Craton, with prolongation to the NE, where its definition is complicated by tectonic events occurring in the Mesoproterozic and Neoproterozoic.

The iron ore deposits of the Minas Supergroup

The huge iron ore deposits of the Quadrilátero Ferrífero (Melo *et al.*, 1986; Gomes, 1986; Barcelos and Büchi, 1986) are associated with the Cauê Formation of the Itabira Group, and resulted from the lateritic weathering of the Cauê Itabirite, increasing the iron grade to 36% to 49%, and locally to 65% by the total or partial leaching of silica and the precipitation of iron oxides and hydroxides together with the residual hematite and the transformation of the compacted itabirite to a rich, friable iron ore.

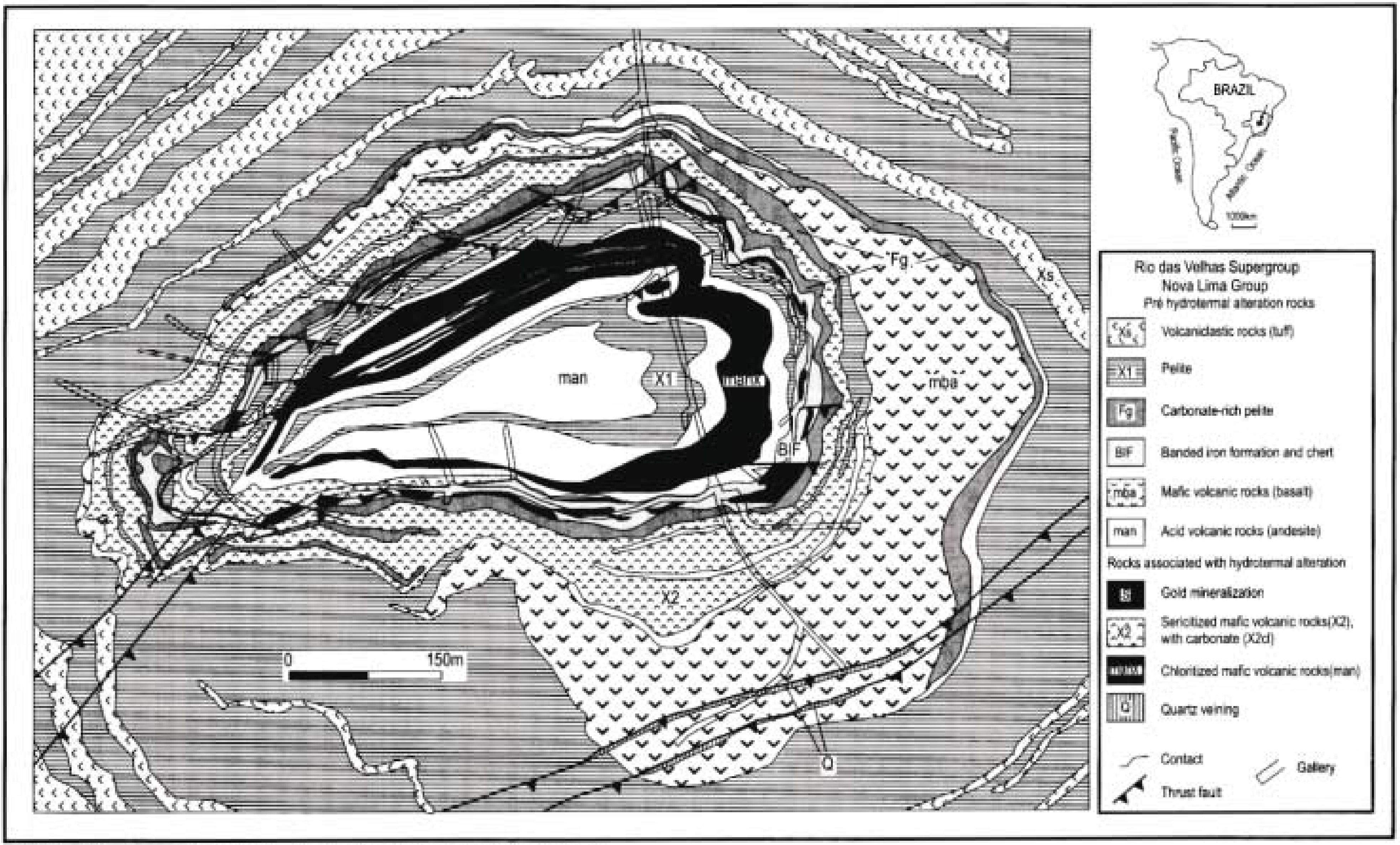


FIGURE 22 - Geological map of the Cuiabá Mine, level no. 3 (modified after Vial, 1988; Vieira, 1991).

The itabirite of the proto-ore is generally preserved in a synform, greatly affected by overthrusting and shear zones associated with the Transamazonian tectonic event at about 2.0 Ga (Noce, 1995).The itabirite displays a banding typical of banded iron formation units of the Superior Province-type, with an alternation of siliceous lamina in the form of quartz, and lamina rich in iron oxide, principally hematite and subordinately secondary magnetite in large martitized crystals. In many deposits, such as the Águas Claras Deposit, the itabirite shows banding consisting of dolomite and quartz/hematite at the base. Lateritic weathering is especially deep over the dolomitic itabirite in function of the higher solubility of the carbonate. The reserves of the Quadrilátero Ferrífero have been estimated at about 29 billion t of iron ore at grades between 50% and 65% Fe. In 1998, the production from all the mines of the Quadrilátero Ferrífero was about 200 Mt of iron ore. The genetic model advanced here proposes a sedimentary exhalative SEDEX-type for this iron deposit.

The gold deposits of the Mariana District

Near the towns of Ouro Preto and Mariana occur several gold deposits: Santana, Antônio Pereira, Passagem de Mariana etc., situated along the Mariana Anticline that constitute a gold district (Duarte and Pires, 1996; Chauvet and Menezes, 1992, Chauvet *et al.*, 1994; Ledru and Bouchot, 1993). These deposits show distinct characteristics, which permit their distinction from deposits associated with the Nova Lima Group. In the Mariana District the gold mineralization occurs in the tectonic contact between units assigned to the Nova Lima Group and the Itabira Group. The gold is hosted in quartz veins with ankerite, tourmaline and arsenopyrite (Vial, 1988), situated between the Paleoproterozoic itabirite and the Archean schist. The gold is found in association with bismuthinite and in the interstices and microfractures of arsenopyrite, quartz, carbonate and tourmaline. The mineralization was already considered to be syngenetic by Fleischer and Routhier (1973), and is related to an extensional phase following the development of a low-angled shear zone (20°-25°) of Transamazonian age (*c.* 2.0 Ga) (Ribeiro, 1998) or during the Brasiliano Tectonic Event as proposed by Chauvet *et al.* (1994).

The gold deposits associated with itabirite

At the Cauê and Conceição mines of the Itabira District, situated to the NE of the Quadrilátero Ferrífero, the gold deposits are hosted in the itabirite. The gold mineralization (*Jacutinga*) is very rich, with grades varying from 10 g/t Au to 1000 g/t Au. The mineralization is contained in at special type of itabirite distributed along the stretching lineations during the Transamazonian Event (2.0 Ga). However, according to Galbiatti *et al.* (1999), Galbiatti (1999), and Pereira *et al.* (1999), the formation of ore bodies is associated with dextral transcurrent shear zones having a transtensive component generating a fracture system that cuts the main foliation of the itabirite and hosts the mineralization. The transcurrent fault is related to the Brasiliano Event.

The Emerald Deposits of Itabira

In the region of Itabira, important emerald deposits are hosted in the hydrothermal alteration zone that occurs around the intrusions of Borrachudo-type granite (1.75 Ga) into the ultramafic rocks of the Nova Lima Group (Giuliani *et al.*, 1993) and Brasiliano pegmatite (570 Ma).

The Diamond Deposits of Diamantina

The Diamantina region is well known for its diamond production since the middle of the XVIII century. The diamonds occur in Mesoproterozoic conglomerate beds of the Sopa-Brumadinho Formation of the Espinhaço Supergroup (1.77-1.71 Ga). These deposits have been mined mainly by *garimpeiros* from weathered conglomerate as well as from alluvial deposits derived from conglomerate beds by fluvial erosion. The placers along the Jequitinhonha River are the source of the largest part of the diamond production of Brazil.

The source of the diamonds has been subject of controversy: intra-basin source versus extra-basin source (Chaves and Uhlein, 1991). The main diamond fields in the Diamantina District are: Campo Sampaio-São João da Chapada, Sopa-Guinda and Extração.

Abreu *et al.* (1997) recognized a quartzitic metabreccia of the top of the Sopa-Brumadinho Formation as a primary source of the diamonds. This breccia is interpretaded by the cited authors as vent-breccia deposited in craters of the *maar*-type.

The measured reserves of the Diamantina District are of about 15 M carats. The grade varies between 0.01 and 0.20 ct/m³ (DNPM-Brazilian Mineral Yearbook, 1978).

The Upper Proterozoic Fold Belts and Related Cover Deposits

At the end of the Mesoproterozoic the São Francisco and Amazonian cratons were surrounded by Neoproterozoic elongated sedimentary basins, the closure of which by orogenic collage at the end of the Brasiliano Cycle resulted in the development of extensive mobile belts known as the Brasília, Araçuaí, Ribeira, Dom Feliciano and Paraguay-Araguaia belts. In Northeastern Brazil, the Borborema Province represents a complex network of old basement nuclei and Neoproterozoic belts.

The Paraguay-Araguaia Belt

Barbosa *et al.* (1966) and Almeida (1967) first described this fold belt that lies around the margin the Amazonian Craton for over 2500 km. Although it constitutes a prominent tectonic feature of apparent continuity, the Paraguay and Araguaia belts probably represent two independent units with distinct sedimentary and tectonic histories.

The Araguaia Belt (Hasui and Costa, 1990; Hasui *et al.*, 1994; Abreu *et al.*, 1994) that can be traced along a N-S strike length for over 1000 km is about 150 km wide, and may be divided into two main domains: a) the internal zone occupied by the Estrondo Group, consisting of gneiss, mica schist and quartzite, with basement exposure at the centre of structural domes (e.g., the Colmeia, Xambioá and Lontra domes); b) the external zone is represented by rocks of the Tocantins Group consisting mainly of psammite and phyllite. The limit between the external zone and the Amazonian Craton is marked by the Tocantins-Araguaia

Lineament, over 700 km long, and which is expressed by a series of mafic-ultramafic bodies that represent ophiolite fragments (Gorayeb, 1989; Souza *et al.*, 1995). The Araguaia Belt was affected by two main tectonic events: the first displays regional compression with vergence to the NW; the second is characterized by ductil-ruptile shears striking N-S. In this belt the most important prospects are related to base metals associated with ophiolitic bodies (Teixeira, 1996; Kotschoubey *et al.*, 1996).

The Paraguay Belt developed during the Vendian (650-550 Ma) and displays sedimentary and tectonic zonation described by Almeida (1945, 1964), Alvarenga and Trompette (1993). From W to E three zones may be distinguished: a cratonic zone with subhorizontal beds; a pericratonic zone, the characteristic of which is the presence of long, large-amplitude holomorphic folds; and a deep basin zone containing metamorphites with vergence to the W. Regionally, the stratigraphy of the Paraguai Belt may be divided from base to top into four units: a) the Puga Formation, of glacial origin and lateral equivalents corresponding to the Jangada Group, a talus deposit, with the deposition of proximal glacio-marine turbidite beds; b) the Cuiabá Group represented by distal turbidite beds and basin pelite; c) the Corumbá Group, consisting essentially of carbonate rocks including limestone and dolomite; and d) the Alto Paraguai Group consisting mainly of sandstone and arkose.

The only known large deposits in the Paraguay-Araguaia Belt are those of the Fe-Mn Urucum-Mutún mines, near Corumbá, Mato Grosso do Sul.

Historically, the phyllite of the Cuiabá group have been noted for the presence of numerous small gold deposits associated with hydrothermal veining (Alvarenga *et al.*, 1990), and surficial concentrations of gold of lateritic origin (Cuiabá-Poconé Province and the Nova Xavantina District).

In the Cuiabá-Poconé or Baixada Cuiabana Province gold deposits occur as (a) quartz and quartz-pyrite veins cutting low-grade metasediments of the Cuiabá Group; (b) supergene enrichment in laterite; and (c) as placer deposits. The gold grade varies between 0.3 and 2 g/t Au. In 1984/85, the district produced 2.5 t/Au from *garimpeiro* workings (Souza, 1988). In the Nova Xavantina District, situated some 650 km NE of Cuiabá, gold occurs in like manner to the Cuiabá-Poconé Province. Highly brittle phyllite and felsic volcanic rocks represent the Cuiabá Group in this district. The average gold grade in quartz veins is 15 to 20 g/Au m^3. Monthly production from *garimpeiro* workings is between 500-600 kg (Souza, 1988).

The Urucum-Mutún Fe-Mn Deposit

The Corumbá Graben is situated in the junction of the Paraguay Belt with the Chiquitos-Tucavaca Aulacogen that separates the Amazonian Craton from the Apa Block (Litherland *et al.*, 1986), along the border between Bolivia and Brazil. In this extensional environment (Haralyi and Walde, 1986), the graben was filled with sediments of the Jacadigo Group (Fig. 23), which may be divided into three formations (Dorr, 1945, Walde *et al.*, 1981): Urucum Formation, green in colour at the base, and consisting of conglomerate and arkose; the Córrego das Pedras Formation, an intermediate unit, consisting of conglomerate and reddish arkose, enriched in hematite; and the Band'Alta or Santa Cruz Formation consisting of jaspilite with intercalated manganiferous beds. To the N and S, the Jacadigo Group is overlain by carbonate sediments of the Corumbá Group. The conglomerate and arkose of the Urucum Formation represent piedmont sediments along the base of the fault-scarps that delimit the graben. The beds of jaspilite and manganese are attributed to chemical precipitation of alternating bands of iron oxide and silica. However, the jaspilite sequence is intercalated with very many beds of diamictite and arkose with gradational structures that are intensely transformed and substituted, partially or totally, by iron oxide and silica. These observations support the view that the siliciclastic sedimentation persisted during the phase of chemical sedimentation in the form of turbidite beds and subaqueous gravity flows (Dardenne, 1998; Trompette *et al.*, 1998). However, according to a divergent view, the control over the siliciclastic deposits was tectonic rather than sedimentary. The pure jaspilite that formed only by chemical precipitation with alternation of hematite and silica lamina occurs in the upper part of the rhythmic sequence, showing from base to top: ferruginous diamictite and conglomerate, ferruginous arkose with gradational structures, ferruginous shale, and finally, pure finely-laminated jaspilite, with an ocellar texture due to the presence of numerous small pink siliceous nodules, coloured by a fine-grained hematite powder. During supergene alteration, these nodules were leached, preferentially, resulting in a vacuolar aspect, very specific to these jaspilites. The manganese beds consist mainly of cryptomelane, and locally, braunite, appear as zones as at the Morro do Urucum, or as nodules in a kaolinitic and sandy matrix as at Morro do Rabicho. These observations permit the interpretation of the iron and manganese mineralization as the result of chemical precipitation of these elements, and the silica from exhalative (SEDEX-type) hydrothermal fluids circulating in large convection cells. These cells were related to the development of the rift, and the hydrothermal fluids permitted the leaching of basalt associated with up-welling of the upper mantle below the Corumbá Graben (Dardenne, 1998). The continuation of the ferro-magnesian sequence to the W in Bolivia is found at the Serranía de Mutún.

The Brasília Belt and the Goiás Massif

The Brasília Fold Belt (Fig. 24) runs for over 1000 km N-S along the western margin of the São Francisco Craton (Almeida, 1977). In general, the several lithostratigraphic units of the Brasilia Fold Belt show more intense progressive deformation, accompanied by increasing metamorphism from E to W, reflecting in the polarity of the belt and an eastward vergence towards the São Francisco Craton. Three distinct tectonic zones are recognized: a Cratonic Zone, an External Zone and Internal Zone (Fuck *et al.*, 1993, 1994). The Pirineus Mega-inflection (Marini *et al.*, 1984a, b; Araújo Filho, 1999), with general strike E-W, permits the division of the Brasília Fold Belt into two distinct parts: northern

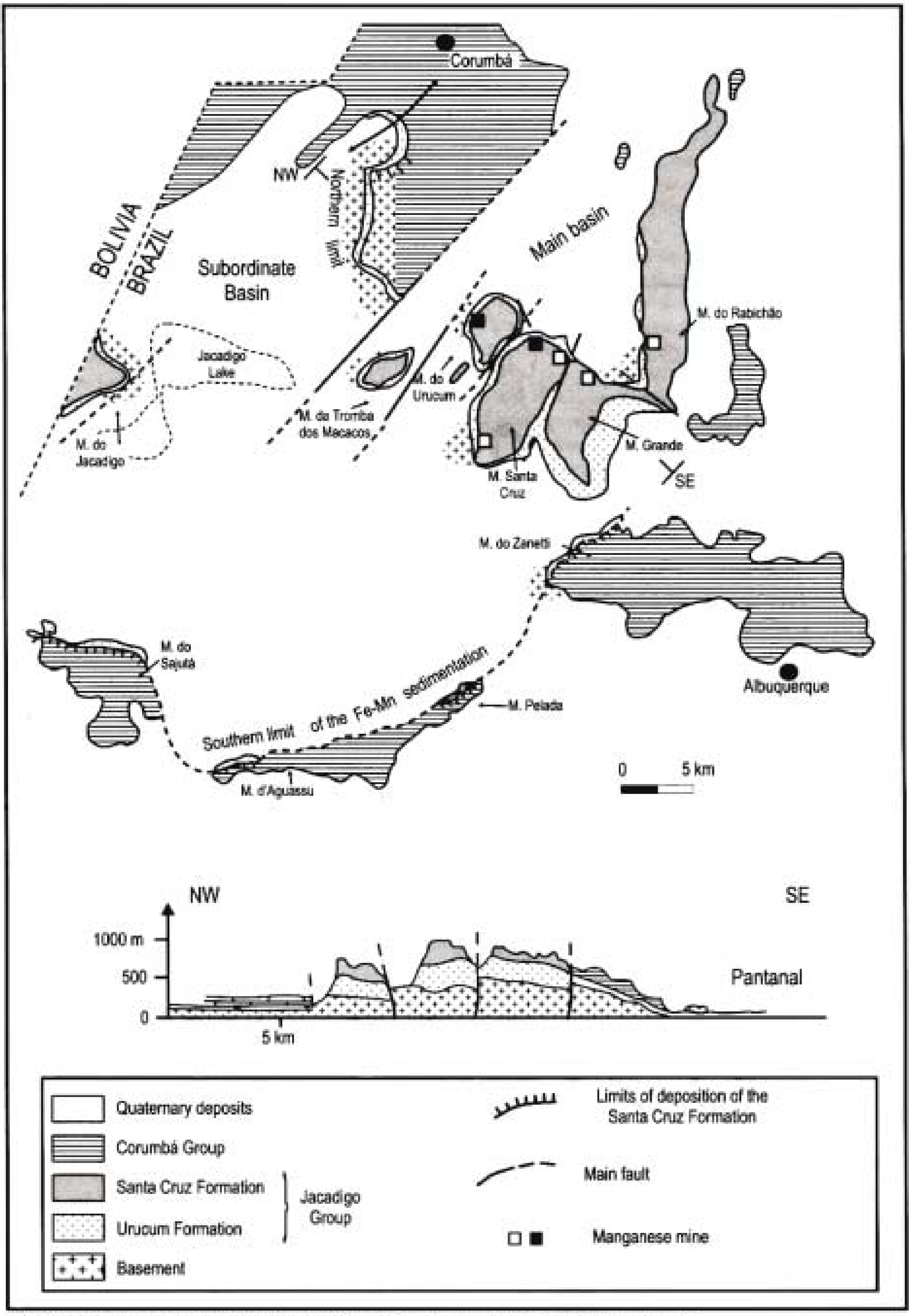


FIGURE 23 - Geological map and cross-section of the Corumbá graben system (modified after Walde et al., 1981).

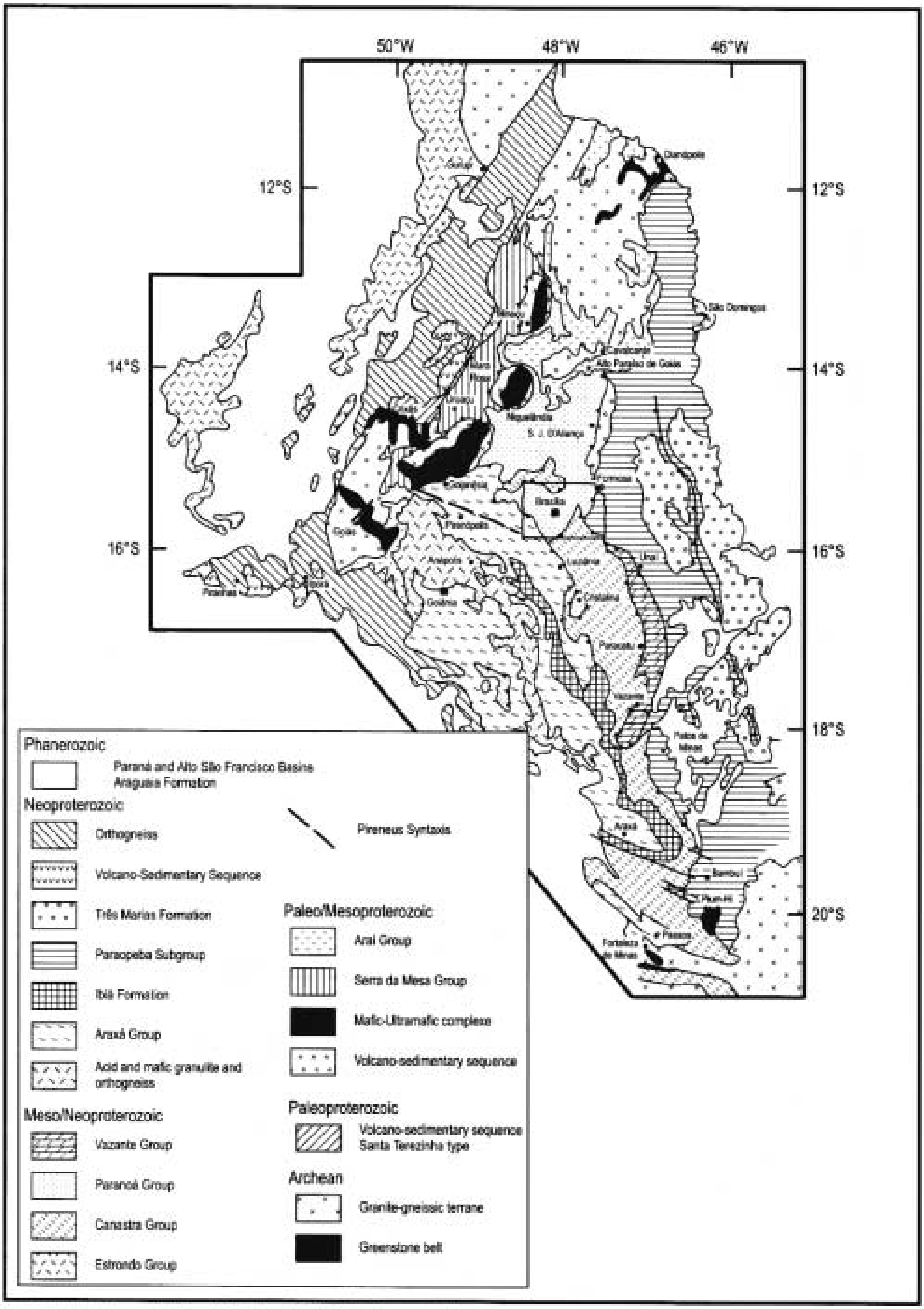


Fig. 24 - Geological map of the Brasília Belt (modified after Dardenne, 1999).

and southern, each undergoing a somewhat different tectonic evolution during the Brasiliano Cycle.

The internal zone of the Brasília Fold Belt and the oldest terranes were both very much affected by deformation and metamorphism during the Brasiliano Event. For this reason both are taken together in the discussion of the tectonic evolution of the Brasília Fold Belt.

The Goiás Massif

The Archean terranes showing ages between 3.0 and 2.5 Ga (Queiroz *et al.*, 1999) occupy an oval-shaped area in the northwestern part of Goiás, in which there occur greenstone belt sequences: Crixás, Pilar de Goiás, Guarinos e Goiás, preserved in elongate synforms in the older granite-gneiss complex: Anta, Caiamar, Hidrolina, Itaporanga and Uvá (Jost *et al.*, 2000). This older nucleus stabilized at about 2.5 Ga, and was accreted to the Santa Terezinha volcano-sedimentary sequence that was deformed and metamorphosed at about 2.0 Ga by the Transamazonian tectono-thermal event (Kuyumjian, 2000).

In the ambit of the Goiás Massif, the principal mineral resources are:

- Gold deposits associated with the Crixás, Guarinos, Pilar de Goiás and Goiás Velho greenstone belts (Carvalho, 1999); Mina III and Mina Nova in the Crixás greenstone belt (Fortes, 1996; Magalhães, 1991); Maria Lázaro and Caiamar in the Guarinos greenstone belt (Lacerda, 1991; Pulz, 1990, Pulz *et al.*, 1991); Cachoeira do Gogó in the Pilar de Goiás greenstone belt (Pulz, 1995).
- The occurrence of Ni sulphide at Boa Vista in the Crixás greenstone belt (Costa Jr. *et al.*, 1997).
- An emerald deposit in the Santa Terezinha Paleoproterozoic volcano-sedimentary sequence that has been affected by the Transamazonian Event at 2.0 Ga (Giuliani *et al.*, 1993).

The Crixás Gold Deposit

The Crixás Greenstone Belt (Fig. 25) is one of the most important from the economic point of view. Here mining is carried on a large scale by Mineração Serra Grande S.A. at Mina III and Mina Nova, in addition to which there are two occurrences known as Meia Pataca/Pompex and Mina Inglesa.

The Mina III deposit (Yamaoka and Araújo, 1988) is situated 2.5 km from the town of Crixás, and the mine has been worked since 1990. The gold mineralization occurs at the base of metasediments of the Ribeirão das Antas Formation near the contact with the mafic volcanic units of the Rio Vermelho Formation. There are three mineralized zones: the lower and upper zones show a homogeneous distribution of the gold, whereas the intermediate zone is somewhat discontinuous. In 1994, the reserves are given at about 4.792 Mt of ore at an average grade of 10.12 g/t Au (Carvalho, 1999).

The lower zone consists of quartz veins, concordant with the principal foliation and varying in thickness from 0.5 m to 5 m; carbonaceous schist with disseminated sulphide (arsenopyrite and/or pyrrhotite) near the veins (Fortes and Coelho, 1997). The mineralized bodies are discontinuous and about 500 m wide along the strike of the foliation, and length of about 1200 m along the dip of the stretching lineation. In this zone the gold occurs preferentially associated with quartz and in the carbonaceous schist. The average grade of the ore is 12 g/t Au (Carvalho, 1999). The upper zone consists of a zone of massive sulphide (pyrrhotite and/or arsenopyrite), between 0.5 m and 2 m thick, associated with Fe-dolomitic marble, quartz-chlorite-carbonate schist, pyrrhotite-magnetite-biotite schist and marble with biotite (Fortes and Coelho, 1997). The ore bodies are lenticular and 50 m to 200 m wide along the strike of the foliation, and 400 m long along the dip of the stretching lineation. In this zone there occur two ore-types: one rich in arsenopyrite, and the second rich in pyrrhotite. The mineralization is accompanied by hydrothermal alteration manifest by silicification, carbonatization and chloritization (Fortes, 1996).

The Mina Nova Mine is situated near Mina III, and has been worked by underground methods since 1996. The gold mineralization is associated with a zone of carbonaceous schist, locally at the contact with marble, chlorite-garnet schist and quartz veins. It is accompanied by a carbonate-rich alteration halo some 9 to 12 m thick. According to Portocarrero (1996) there are three ore-types: Type I consists of a zone 1.5 m to 2.8 m thick of carbonaceous schist with disseminated pyrrhotite, arsenopyrite and subordinate chalcopyrite; Type II occurs in a zone overlying the Type I ore, 0.3 m to 1.7 m thick and consisting of carbonaceous sericite schist with disseminated arsenopyrite; Type III ore occurs rarely in thin quartz veins with disseminated gold, arsenopyrite and pyrrhotite. In 1996, the reserves were given as 3 Mt of ore at 6 g/t Au (Carvalho, 1999). In 1998, the production of the Mineração Serra Grande Ltda. was about 4.5 t Au.

According to Thomson and Fyfe (1990), Fortes *et al.* (1997), the mineralization is associated with a low-angled shear zone, related to the Brasiliano Event.

The Paleo-Mesoproterozoic Intracontinental Rift

The development of the intracontinental rift during the Paleo-Mesoproterozoic (Nilson *et al.*, 1994) occurred through several stages: 1) the intrusion of the mafic-ultramafic complexes of Cana Brava, Niquelândia and Barro Alto at about 2.0 Ga (Correia *et al.*, 1996, 1997); 2) the intrusion of anorogenic tin granite at 1.77 Ga in the Rio Paranã sub-Province, and at 1.59 Ga in the Rio Tocantins sub-Province (Pimentel *et al.*, 1991; Marini and Botelho, 1986; Botelho and Moura, 1998); 3) the deposition of the Araí and Serra da Mesa sediments in the pre, syn and post-rift phases (Dardenne and Freitas Silva, 1999).

Several mineral deposits are associated with the evolution of the rift:

- Deposits of Ni laterite associated with mafic-ultramafic complexes of Niquelândia and Barro Alto;
- Promising occurrences of PGE in the transition zone between peridotite and pyroxenite in the mafic-ultramafic complexes of Niquelândia and Cana Brava (Ferreira Filho *et al.*, 1992; Ferreira Filho *et al.*, 1994; Medeiros and Ferreira Filho, 1999; Oliveira, 1993; Lima, 1997; Suita, 1996, 1998);
- Large asbestos deposit that originated as the results of Brasiliano tectonics in the mafic ultra-mafic Cana Brava

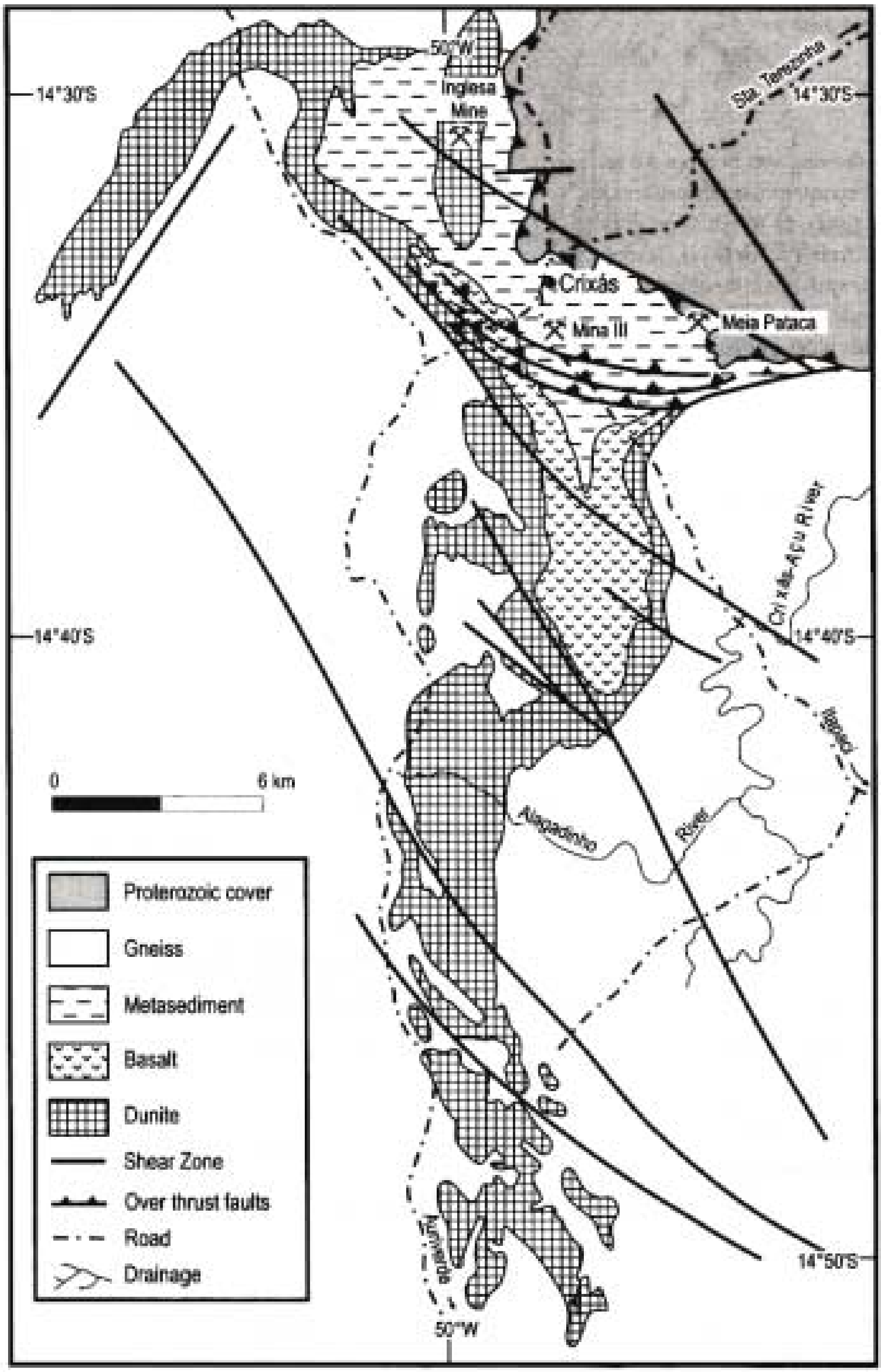


FIGURE 25 - Geological map of Crixás Greenstone Belt (modified after Magalhães, 1991).

complex (Pamplona and Nagao, 1981; Ianhez *et al.*, 1997).

Deposits of cassiterite and indium associated with greisenization and albitization related to the anorogenic granitic intrusions (Marini and Botelho, 1986; Botelho, 1992; Botelho and Moura, 1998).

The Cana Brava Asbestos Deposit

The mafic-ultramafic Cana Brava Complex is part of a high-grade terrane (Goiás Granulite Belt) that was placed over the metasediments of the Araí and Paranoá groups of the Brasília Fold Belt by Brasiliano collisional tectonics. The asbestos deposit, discovered in 1962, near the town of Minaçu, is associated with serpentinite at the base of the complex in its southeastern extremity. The mineralized belt (Ianhez *et al.*, 1997) is about 6300 m long (Pamplona and Nagao, 1981), and is essentially tabular in shape. It is about 4800 m N-S and 1500 m E-W, and it contains the A, B, C and F ore bodies, having an average thickness of between 100 m and 110 m. The dip varies between 10° and 70° to the NW and W. These bodies extent to a depth of 290 m, and they are separated by dextral transcurrent shear zones the strike of which is NE-SW. The principal foliation is mylonitic and is roughly parallel to the original banding. Metamorphism is in the greenschist to amphibolite facies, resulting in the development of type I brown serpentinite. The zones of transverse and transcurrent shearing were initially ductile (greenschist facies) and subsequently ruptile, and resulted in the development of type II green serpentinite. The most important mineral species is chrysotile. The mineralization occurs in stockworks, and is restricted to extension fractures distributed in the serpentinite mass. The long axis of the fibrous chrysotile permits the grouping of the fibers into two types: the slip fibers are oriented parallel to the walls of the fractures, whereas the cross fibers are transverse to the vein wall. The length of the fibers varies between 1 mm and 20 mm, for an average length of about 6 mm. The accessory minerals are magnetite and hematite, chlorite, carbonate and talc. In 1997, the measured reserves for the A and B bodies are given as 122.89 Mt of fibrous ore at 5.2%.

The Brasiliano Cycle

The reconstruction of the tectonic evolution permits the differentiation of some fundamental phases in the development of the Brasiliano Cycle.

The Development of the Mesoproterozoic/Neoproterozoic Passive Margin

The development of the passive margin is characterized by the deposition of metapelite, limestone and dolomite on a marine platform dominated by tides and/or storms. These sediments are assigned to the Paranoá and Canastra groups (Fuck *et al.*, 1988, 1993). The Vazante Group that occurs in the southern part of the Brasília Fold Belt may correspond to the upper carbonate and pelitic part of the Paranoá Group. To the W of the mafic-ultramafic complex, in the northern part of the Brasília Fold Belt, occurred the opening of an ocean at 1.2 Ga (Correia *et al.*, 1999) in which were deposited the volcano-sedimentary sequences of Palmeirópolis, Indaianópolis and Juscelândia. The principal mineral deposits related to the passive margin are:

- Small deposits of Pb-Zn-Ag (Cu) of the VMS-type, associated with the Palmeirópolis and Juscelândia sequences (Araújo and Nilson, 1988; Araújo, 1999).

The Goiás Magmatic Arc

The development of the Goiás Magmatic Arc, situated along the western margin of the Brasilia Fold Belt started at about 900 Ma (Pimentel and Fuck, 1991, 1992), and remained active until 600 Ma. The arc consists of gneiss, granitoid rocks and volcano-sedimentary sequences of the island arc-type. The Goiás Magmatic Arc is related to the formation of a back-arc basin in which were deposited the Ibiá and Araxá groups.

The mineral deposits are associated with the rocks of the volcano-sedimentary sequence, and are classified as being of the VMS-type: Cu-Pb-Zn-Au at Chapada (Kuyumjian, 1991, 1995, 1999; Richardson *et al.*, 1986); Au-Ag-Ba at Zacarias (Arantes *et al.*, 1981) and Cu-Au at Bom Jardim de Goiás.

The pre-Collisional Event at 790 Ma

This event was responsible for the granulitization observed at the base of the mafic-ultramafic unit, and for the generation of syn and late-tectonic granite, enriched in tin in the Ipameri region (Pimentel *et al.*, 1999). It is probable that the beginning of the evolution of the shear zones responsible for the gold mineralization date from this event.

The Foreland-type Bambuí Basin

With the uplift of the Brasília Fold Belt developed a depression in front of a mountainous chain in which began the deposition of pelite and carbonate of the Bambuí Group in a foreland-type basin. This sedimentation extended much further than the original depression, covering a large part of the São Francisco Craton to the E in the states of Bahia and Minas Gerais (Dardenne, 1978).

The saccharoidal pink dolomite belonging to the upper part of the first regressive carbonate cycle of the Bambuí Group is related to small deposits and occurrences of Pb-Zn-Ag-CaF_2 found along the valley of the São Francisco River in the vicinity of Januária, Itacarambi, Montalvânia and Serra do Ramalho. These deposits show clear indications of dissolution, substitution and secondary dolomitization by circulating connate hydrothermal fluids that permit the classification of these deposits as being of the MVT-type (Dardenne, 1978, 1979; Dardenne and Freitas-Silva, 1998, 1999).

Large deposits of phosphate, lead and zinc are found in the Vazante Group (Dardenne and Freitas Silva, 1998) that probably constitutes the transition between the Paranoá (Mesoproterozoic) and Bambuí (Neoproterozoic) groups.

The phosphate deposits of Rocinha-Lagamar

The phosphate deposits of Rocinha-Lagamar (Dardenne *et al.*, 1997; Chaves *et al.*, 1976) situated in the northwestern part of the State of Minas Gerais occur in the basal part of the Vazante Group in the external zone of the

Brasília Fold Belt. The phosphorite, associated with carbonaceous and carbonate-rich slate, dark grey in colour and intensely microfolded, occurs as phosphoarenite, phosphorudite and phospholutite. The phosphoarenite consists of intraclasts and phosphatic pellets, set in a cryptocrystalline phosphomicrite. Locally, the intraclasts are surrounded by fibrous cement of microcrystaline, limpid, prismatic apatite. The main mineral species is fluorapatite that resulted from the leaching of CO_2 of the regional carbonate-fluorapatite by fluids related to both metamorphism and weathering, culminating in the development of apatite, rich in aluminum and strontium of the wavelite-type.

The origin of the phosphate is related to the evolution of organic matter in physio-chemical conditions that were transitional between reducing and oxidizing environments in relatively deep water, probably representing a glacio-marine depositional system.

The reserves of the Lagamar Deposit are about 5 Mt at 30% to 35% P_2O_5, whereas the reserves of the Rocinha Deposit are estimated about 400 Mt ore at 10% to 12% P_2O_5.

The Morro Agudo Pb-Zn deposit

The Morro Agudo Pb-Zn deposit (Fig. 26) is associated with dolomite of the Vazante Group, situated in a back-reef facies that developed on the western flank of a stromatolitic bioherm of Morro do Calcário (Dardenne, 1978, 1979). The mineralization is essentially disseminated and consists mainly of sphalerite and galena with subordinate pyrite and barite. Breccia, dolarenitic breccia and dolarenite are the main host rocks of the main mineralized zones, denominated I, J, K and L. Zone M has a stratabound character, whereas zone N is stratiform with regular banding of chert lamina, galena, sphalerite and pyrite. Pyrite is exceptionally abundant in this zone (Romagna and Costa, 1988; Oliveira, 1998). The mineralized beds are limited by a syn-sedimentary normal fault with strike N10°W that acted as a preferential conduit for the mineralizing fluids. The disseminated mineralization shows evidence for the substitution of non-consolidated dolomitic material by sphalerite and galena that developed from the syn-diagenetic stage to the late-diagenetic stages (Dardenne, 1979; Dardenne and Freitas-Silva, 1999; Freitas-Silva and Dardenne, 1997). The data favour the comparison of the Morro Agudo Deposit with the Navan Deposit, Ireland (Hitzmann, 1995), that show the same characteristics as described by Dardenne (In: Pedrosa-Soares *et al.*, 1994), Freitas-Silva and Dardenne (1997), Dardenne and Freitas-Silva (1998, 1999), Hitzmann *et al.* (1995), and classified as the SEDEX-type by Misi *et al.* (1999). The reserves of Morro Agudo Pb-Zn deposit is about 17.5Mt ore at 5.1% Zn, 1.53% Pb and 300 ppm Cd.

The Vazante Zn deposit

The Vazante Zn deposit (Fig 27) is associated with a major tectonic structure, represented by a normal fault the attitude of which is N45°E/50° to 70°NW (Dardenne, 1979; Dardenne and Freitas-Silva, 1998, 1999). The fault zone is practically restricted to the pelitic interval occurring between dark grey dolomite at the base of the section and pink dolomite at the top, assigned to the Vazante Group (Dardenne 1979; Rigobello *et al.*, 1988; Oliveira, 1998). The mineralization contained in the fault zone is intensely sheared, and occurs in the form of lenticular and imbricate pockets of ore and dolomite (Pinho, 1990; Rigobello *et al.*, 1988; Dardenne, 1979; Dardenne and Freitas-Silva, 1998, 1999). The ore consists mainly of willemite along with hematite and zincite with subordinate franklinite, smithsonite, sphalerite and galena (Monteiro, 1997; Monteiro *et al.*, 1996). The willemite ore is extremely rich in Zn, containing 40% to 45% Zn. The mineralization is accompanied by intense silicification and sideritization of the wall rock that also display a network of fractures and veins filled with siderite/ankerite and red jasper. The data suggest that the hydrothermal mineralization originated by filling of a listric fault that had been reactivated during the Brasiliano Event. The partial reserves of the Vazante Deposit are about 8.5Mt ore at 23% Zn.

The Irecê phosphate deposit

The phosphate deposit discovered in 1985 by the CPRM (Bonfim, 1986) is situated at Fazenda Três Irmãs. It is associated with columnar stromatolites of the Jurussania Krylov-type intercalated with units of cross-bedded dolarenite assigned to the Salitre Formation of the Una Group (Neoproterozoic), equivalent to the Bambui Group in the Irecê Synclinorium. This dolomitic and phosphatic unit, approximately 18 m thick, is overlain by a sequence of silicified dolosiltstone, rich in sulphide (pyrite, sphalerite and galena with associated pyrite) and barite nodules, investigated by RioFinex in 1976 and by the CBPM. Reserves are estimated at 1.0 Mt at 8% Pb+Zn and 120 g/t Ag. Three types of primary phosphorite are present: columnar stromalitic phosphorite, laminar stromatolitic phosphorite, and intraclastic phosphorite. The highest grades are found in the columnar stromatolitic phosphorite where these may attain 20% P_2O_5. The phosphorite consists of micro and cryptocrystalline fluorapatite, associated with calcite and dolomite, in addition to detrital quartz and microcline, fluorite veins, microcrystalline quartz, pyrite, sphalerite and galena (Misi, 1992; Misi and Kyle, 1994; Kyle and Misi, 1997). The intraclastic phosphorite resulted from the erosion of the columnar and laminar stromatolitic phosphorite, forming the intercolumnar material and the beds intercalated with columnar and laminar stromatolites. The phosphatization is early, occurring in a syn-diagenetic phase prior to the dolomitization and is intimately associated with the development of cianobacteria colonies that formed the stromatolitic lamination. The time of the fluorapatite precipitation has not yet been defined. This may have occurred directly as the result of bacterial activity or, alternatively, as the early substitution of the carbonate lamina, rich in organic matter.

The secondary phosphorite resulted from the supergene alteration of primary phosphorite by the preferential leaching of carbonate, inducing a significant enrichment in phosphate. This type of ore attains grades exceeding 30% P_2O_5 (up to 38% P_2O_5). The estimated reserves of the Irecê Deposit are about 40 Mt of ore at 14% P_2O_5.

The Late Collisional Event at 630 Ma

This event is manifest in the overthrusted sheets of Araxá, Ibiá and Canastra sediments as nappes along low-angled faults over rocks of the Vazante and Bambui groups.

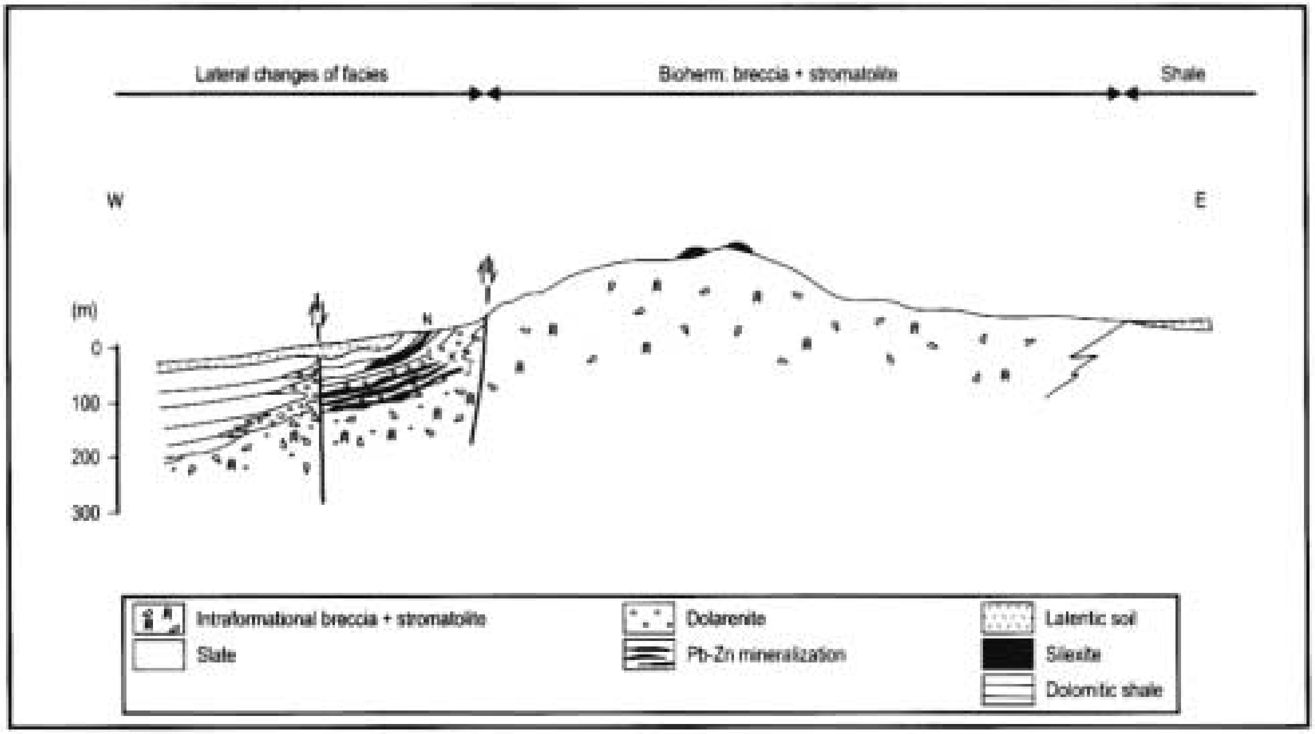


FIGURE 26 - Schematic geological section of Morro Agudo Pb-Zn Deposit (modified after Dardenne, 1978; Dardenne and Freitas Silva, 998).

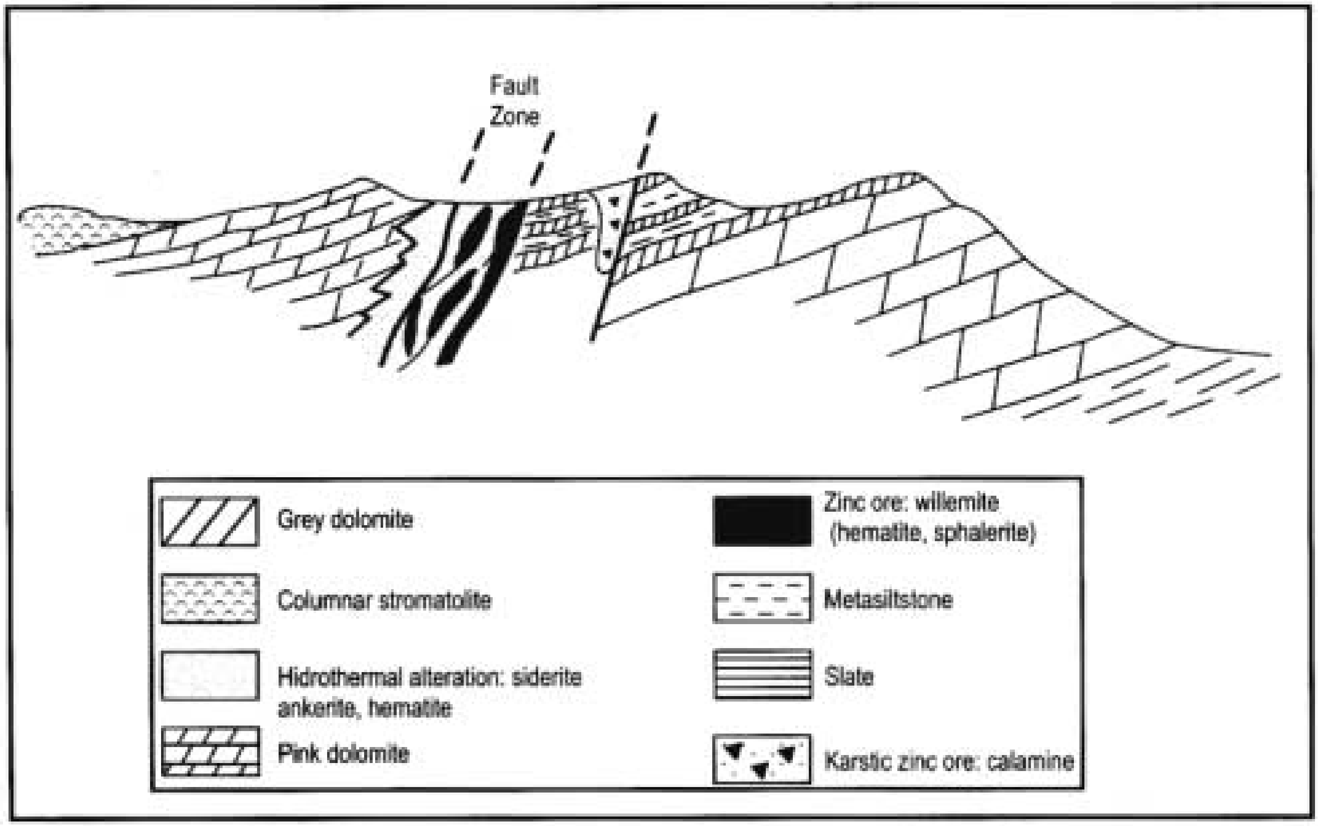


FIGURE 27 - Schematic cross-section of the Vazante Mine (modified after Dardenne, 1978; 1999).

Many gold deposits are associated with the development of high-angled shear zones: Buracão, Santa Rita, Rio do Carmo, Buraco do Ouro (D'El Rey Silva and Senna Filho, 1998; Lacerda, 1986; Giuliani *et al.*, 1993, Olivo, 1989; Olivo and Marini, 1988; Magalhães *et al.*, 1998; Magalhães and Nilson, 1996); and low-angled shear zones: Araxá, Luziânia, and Morro do Ouro in the external zone of the Brasilia Fold Belt (Hageman *et al.*, 1992; Freitas-Silva *et al.*, 1991; Freitas-Silva, 1996). In the region occupied by the Goiás Magmatic Arc, gold deposits are also associated with shear zones: a gold occurrence at Posse, for example (Arantes *et al.*, 1981; Oliveira *et al.*, 1997).

At the end of this second collisional event there occurred late and post-tectonic granitic and mafic-ultramafic intrusions. The mafic-ultramafic complex of Americano do Brasil and Mangabal I and II contain Cu-Ni-Co mineralization in the form of chalcopyrite, pentlandite and bornite (Nilson, 1981).

The Morro do Ouro deposit

The Morro do Ouro Deposit (Freitas-Silva *et al.*, 1991; Freitas-Silva, 1996) is hosted in carbonaceous phyllite (Morro do Ouro Member) of the Paracatu Formation (Canastra Group), which overthrusts the rocks of the Vazante Group in the external zone of the Brasilia Fold Belt. These rocks were deformed and metamorphosed in the greenschist facies (chlorite) and display intrafolial, isoclinal and recumbent folds related to a monocline, associated with a thrust fault (Fig. 28) of large amplitude, the attitude of which is N10°W/15°SW. This fault developed in a ductile-ruptile shear zone accompanied by mineral lineation and stretching (elongation) fabrics, mylonitic foliation and boudinage of the quartz veins. The stretching lineation is orientated S70°W/15 °. The gold is disseminated in segregations of metamorphic quartz (boudins), along with arsenopyrite, pyrite, sphalerite, galena, siderite and sericite. The gold particles are more concentrated at the margins of the boudins and in the proximity of sulphide and carbonate species. The hydrothermal alteration is restricted to the boudins, where the principal processes are pyritization, sideritization and sericitization. The gold is usually free in the quartz, although a small amount may be associated with sulphide. The average gold grade in the boudins is about 2.5 g/t Au. In the ore, the gold grade is very low, about 0.45 g/t Au. Nevertheless, the annual production is roughly 8.0 t Au, and the reserves exceed 250 t Au.

The Araçuaí Belt

The Araçuaí Fold Belt (Almeida, 1967, 1977; Almeida *et al.*, 1981) that lies along the southern and southeastern margin of the São Francisco Craton in the NE of Minas Gerais and along the divide with Bahia, is arc-shaped and concave to the SE. This belt resulted from the development of a rift that began about 800 Ma with the rupture of the continental crust, and continued opening until ocean crust began to form in the middle of the structure (Pedrosa-Soares *et al.*, 1992; 1998; Fuck *et al.*, 1993). The rift basin filled with sediments assigned to the Macaúbas Formation that were distributed in three domains: continental, transitional and internal, resulting in a sedimentary zonation of the belt (Uhlein, 1991; Pedrosa-Soares *et al.*, 1992, 1998; Uhlein *et al.*, 1999). The closure of the rift from 700 Ma that culminated in the principal orogenic phase at between 659 and 550 Ma, brought about deformation and metamorphism, resulting in a polarity varying from E-W to N-S and the development of a series of thrusts in the transitional/continental domains, anatexis and granitic intrusions in the internal domain (Pedrosa-Soares *et al.*, 1999).

Older terranes, related to the Espinhaço, Minas and Rio das Velhas supergroups, as well as younger units such as the Bambui Group at the edge of the São Francisco Craton were involved in the deformation and metamorphism of the Araçuai Belt. The main mineral resources of the Araçaui Belt are:

- The Riacho dos Machados gold deposit (Fonseca *et al.*, 1997) is hosted in a volcano-sedimentary sequence (Archean or Paleoproterozoic in age) developed in a Brasiliano shear zone that has affected the basement of the Araçuaí Belt.

- Iron deposits of the sedimentary-exhalative-type (SEDEX) present in the diamictite beds of the Macaúbas Group in the Porteirinha region (Vilela, 1986);

- Graphite deposits of the Minas Gerais-Bahia area. These deposits are very important. They occur near the towns of Pedra Azul, Salto da Divisa and Maiquinique, and they are associated with schist and gneiss cut by shear zones (Pedrosa-Soares *et al.*, 1994, 1999; Faria, 1997; Reis, 1999);

- The Western Pegmatite Province of Brazil, famous for its collector-quality mineral specimens, is associated with granitic magmatism of the internal zone of the Araçuaí Belt (Correia Neves, 1997; Pedrosa Soares *et al.*, 1994; Quémeneur and Lagache, 1999; Correia Neves *et al.*, 1986, 1987).

The Ribeira Belt

The Ribeira Fold Belt is a continuation of the Araçuaí and the Brasilia belts, along the southeastern-southern coast of Brazil. The characteristic of the Ribeira Belt is the presence of swarms of subvertical longitudinal faults representing dextral shear zones, locally with displacement of tens of kilometres. In the N of the belt, these subvertical shear faults give way to subhorizontal thrusts and overthrust. In the Cenozoic, these sequences underwent uplift of the central zone that resulted in the relief of the Serra do Mar. The dextral longitudinal faults that cut the basement as well as the metasedimentary sequences of the Ribeira Fold Belt define, in the internal part of the uplifted area, a corridor about 1000 km long and 100 km wide known as the Apiai-São Roque Fold Belt. This belt contains metasediments of the Açungui and São Roque groups with which are associated numerous Brasiliano granitic intrusions that have affected the older basement terranes and the metasediments of the Setuva Group.

In the Ribeira Belt, the principal mineral resources are found in the Vale do Ribeira (valley of the Ribeira River) (Fig. 29) that include:

- Deposits of Pb-Zn-Ag-Ba of the Perau-type in the Perau/Água Clara complexes of Mesoproterozoic age. These are generally concordant and stratiform, and are considered

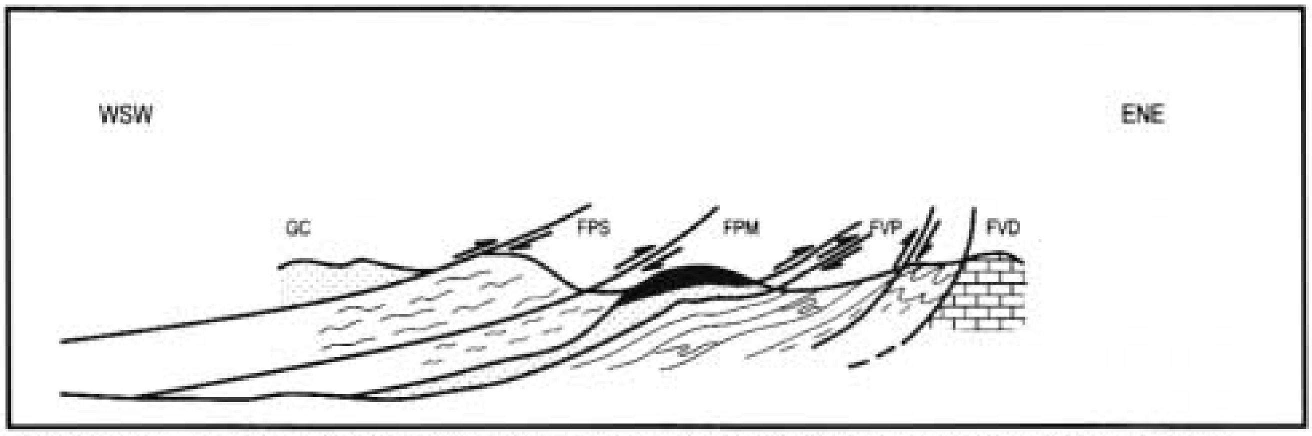


FIGURE 28a - Schematic representation of the structure in the Morro do Ouro gold mine. FP = Paracatu Formation; S = Serra da Anta Facies; FV = Vazante Formation; D = dolomite; P = psamopelite; GC = Canastra Group; black = Morro do Ouro gold deposit (modified after Freitas-Silva et al., 1991).

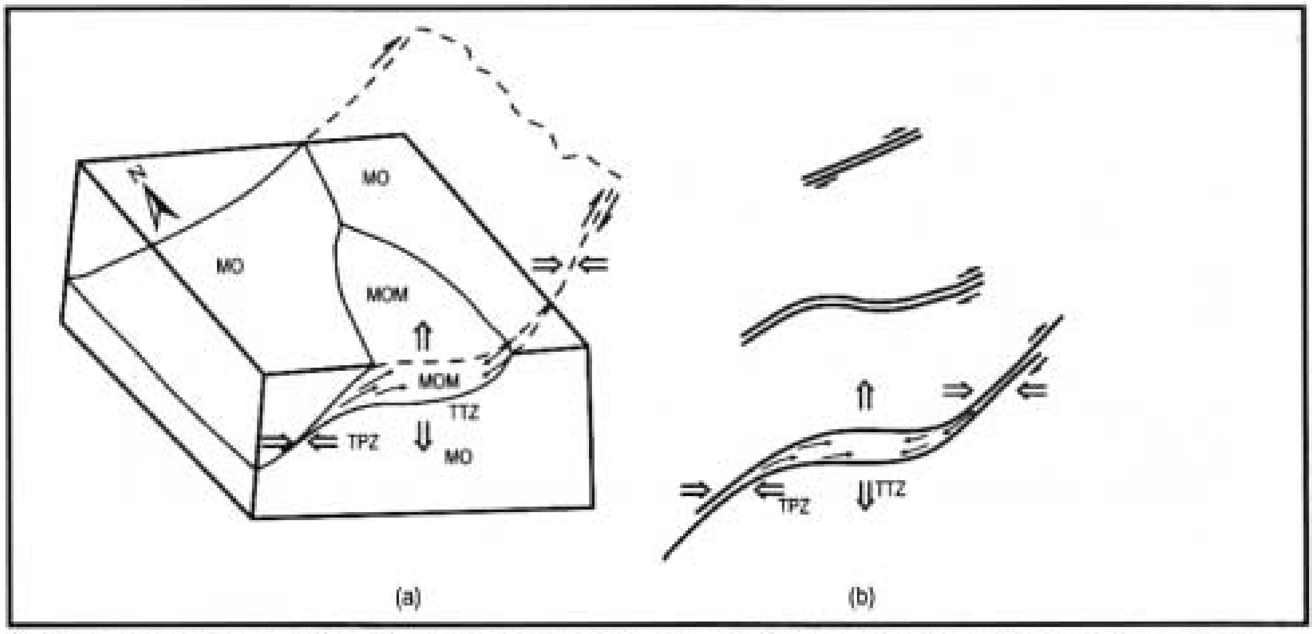


FIGURE 28b - Schematic representation of the formation of the Morro do Ouro gold deposit. (a) - MO = Morro do Ouro Facies, MOM = Morro do Ouro deposit; TPZ = transpressive zone; TTZ = transtensive zone; single tailed arrows = fluid migration direction. (b) - Evolution stages of gold concentration (modified after Freitas-Silva et al., 1991).

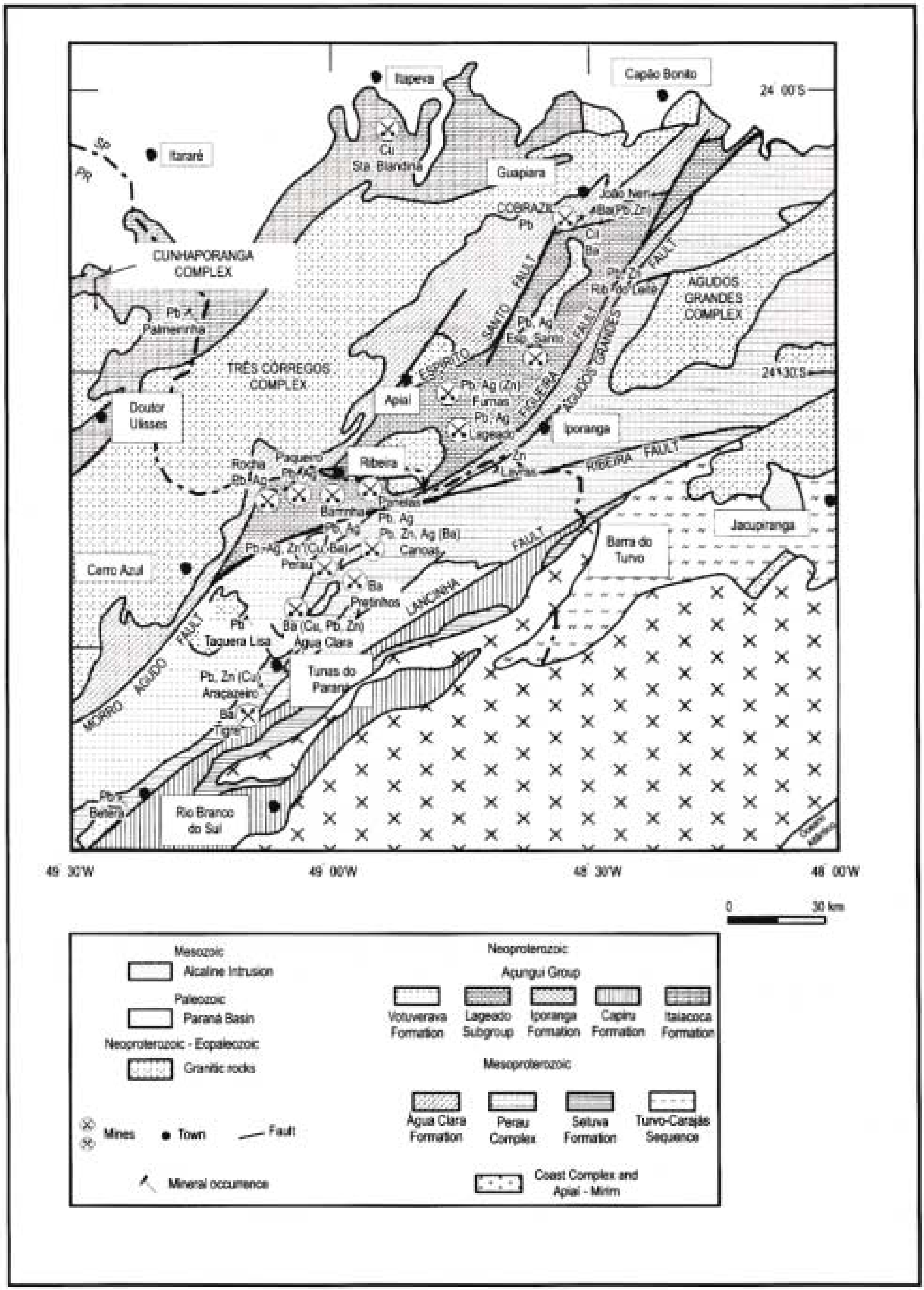


FIGURE 29 - Distribution of base metals deposits in the Vale do Ribeira region (modified after Daitx, 1998).

to be of sedimentary-exhalative origin of the SEDEX-type (Fleischer, 1976; Daitx, 1998);

- Deposits of Pb-Zn-Ag of the Panelas-type in the Lageado and Itaiacoca subgroups. These are generally stratabound in the form of vertical veins in carbonate rocks of Meso- to Neoproterozoic age (Fleischer, 1976; Daitx, 1998).

- Stratabound deposits of fluorite at Sete Barras, Volta Grande and Mato Dentro (Dardenne and Touray, 1988; Dardenne *et al.*, 1997; Ronchi *et al.*, 1993);

- Tungsten deposits associated with the Itaóca granite intrusion (Mello and Bettencourt, 1998);

- Talc deposits in the Itaiacoca Group at Abapã, controlled by vertical shear zones at the contact between quartzite and dolomite and their reworked products deposited in karst depressions (Lima, 1993; Lima and Dardenne, 1987).

The Dom Feliciano Belt and the Rio De La Plata Craton

According to Jost (1981) and Fernandes *et al.* (1995) the Dom Feliciano Fold Belt, situated in the southern extremity of Brazil developed between the Rio de La Plata Craton and the Kalahari Craton as a result of subduction to the NW of the oceanic crust and from the accretion of older magmatic arcs between 850 and 750 Ma at the margin of the Rio de La Plata Craton giving rise to the Pelotas Batholith. The opening of a back-arc basin followed by collision around 650 Ma led to the accretion of a second magmatic arc and the intrusion of calc-alkaline to alkaline granite into the Encruzilhada do Sul metasedimentary sequence with vergence to the NW. Marine and continental molasse deposits accompanied by alkaline volcanism (Bom Jardim Group) were deposited in intramontane basins formed by grabens and half-grabens.

The main mineral deposits observed in the Dom Feliciano Belt are:

- Occurrences of Pb-Zn of the VMS-type in the Vacacaí volcano-sedimentary sequence;

- Deposits of porphyry gold-type;

- Tin deposits associated with S-type granite (Franz, 1997; Franz *et al.*, 1998);

- Deposits of Cu-Pb-Zn in the Camaquã region.

Deposits of the porphyry-gold-type

Most of the gold mineralization of the porphyry-gold-type is found in the Lavras do Sul Granitic Complex, as well as in Neoproterozoic felsic volcanic rocks and fluvio-lacustrine sediments assigned to the Cerro dos Martins Formation in the State of Rio Grande do Sul. The mineralization occurs as the disseminated and vein types, and is associated with faults and intense shearing, brecciation and hydrothermalism. The mineralization is genetically related to late magmatism of the Brasiliano Event (610-580 Ma) having shoshonitic and alkaline affinities. The parageneis is of the Au-Cu-Pb-Zn-Ag type. The most important deposit is that of Volta Grande (7 t Au), and the Bloco Butiá (6.5 t Au) and Cerrito (3.5 t Au) prospects (Santos *et al.*, 1998; Andrade *et al.*, 1988; Reischel, 1980).

The San Gregorio Gold Mine of northern Uruguay (Fig. 2) may also be an example of mineralization of the porphyry-gold type. The gold mineralization followed the intrusion of the Corrales Granite, and consists of auriferous pyrite, chalcopyrite, pyrrhotite, galena, and Fe-rich sphalerite. This granite is probably of Brasiliano age and is intruded into a Paleoproterozoic granite-gneiss complex of the Rio de la Plata Craton. E-W and NW-SE shear zones cut this complex. The mineralized zones currently being mined at the San Gregorio Mine are the same as that of the wall rock. The mineralized zones are hydrothermally altered gneiss and metabasalt, in addition to a quartz vein system. (Ellis *et al.*, 1995). The gold reserves at the San Gregorio Mine are about 6.5 Mt at 2.8 g/t Au and 2.8 g/t Ag (0.65 Moz Au and 0.65 Moz Ag) (J. Spoturno, personal communication).

Camaquã District

The mineral deposits of the Camaquã District (Fig. 30), situated in the central-southern region of the State of Rio Grande do Sul, are associated with conglomerate and sandstone of the red bed-type. These red beds belong to the Vargas Member of the Arroio dos Nobres Formation of the Bom Jardim Group deposited in a system of alluvial fans at the end of the Dom Feliciano (630-600 Ma) collisional orogenesis in a molasse basin, delimited by faults striking NE-SW. This basin has been interpreted as being of the foreland-type or an intramontane basin of the strike-slip-type with rhyolitic, dacitic and andesitic volcanism of the Hilário Member at the base. The Bom Jardim Group is covered by an angular unconformity by sediments, also of the red bed-type assigned to the Guaritas Formation with which are associated the rocks of the Rodeio Velho Member, dated at 470 Ma by Hartman *et al.* (1998), implying that the basin developed between 600 and 470 Ma. To the NW of the basin occur a number of granite intrusives of calc-alkaline to shoshonitic composition, referred to as Lavras do Sul, Caçapava do Sul and São Sepé that have been dated at between 590 and 560 Ma (Remus *et al.*, 1999). In this district, three types of mineralization are recognized:

1) Vein mineralization, discovered in 1865, and intensely mined to 1996 at the Camaquã mines known as São Luiz (underground) and Uruguai (underground and open pit), which have produced about 398 Mt of ore at 1.06% Cu, 0.2 g/t Au and 8 g/t Ag (Teixeira and Gonzalez, 1988; Remus *et al.*, 1999). These deposits have been described successively by Bettencourt (1976), Ribeiro (1991), Remus *et al.* (1999) and Ronchi *et al.* (1999). The ore occurs as *amas*, veins, ribbons and stringers, locally forming stockworks with orientation parallel to the faults striking NW-SE, and is surrounded by hydrothermal alteration halos displaying chloritization, sericitization and silicification (Remus *et al.*, 1999; Ronchi *et al.*, 1999; Bettencourt, 1976). The paragenesis is pyrite-chalcopyrite and quartz, and bornite-chalcopyrite-hematite-barite-calcite.

2) Disseminated mineralization in the sandstone and conglomerate units of the Vargas Member:

- Copper mineralization around the Camaquã mines (São Luiz and Uruguai) with paragenesis of pyrite-chalcopyrite (Veigel and Dardenne, 1990).

- Lead and zinc mineralization with subordinate copper at

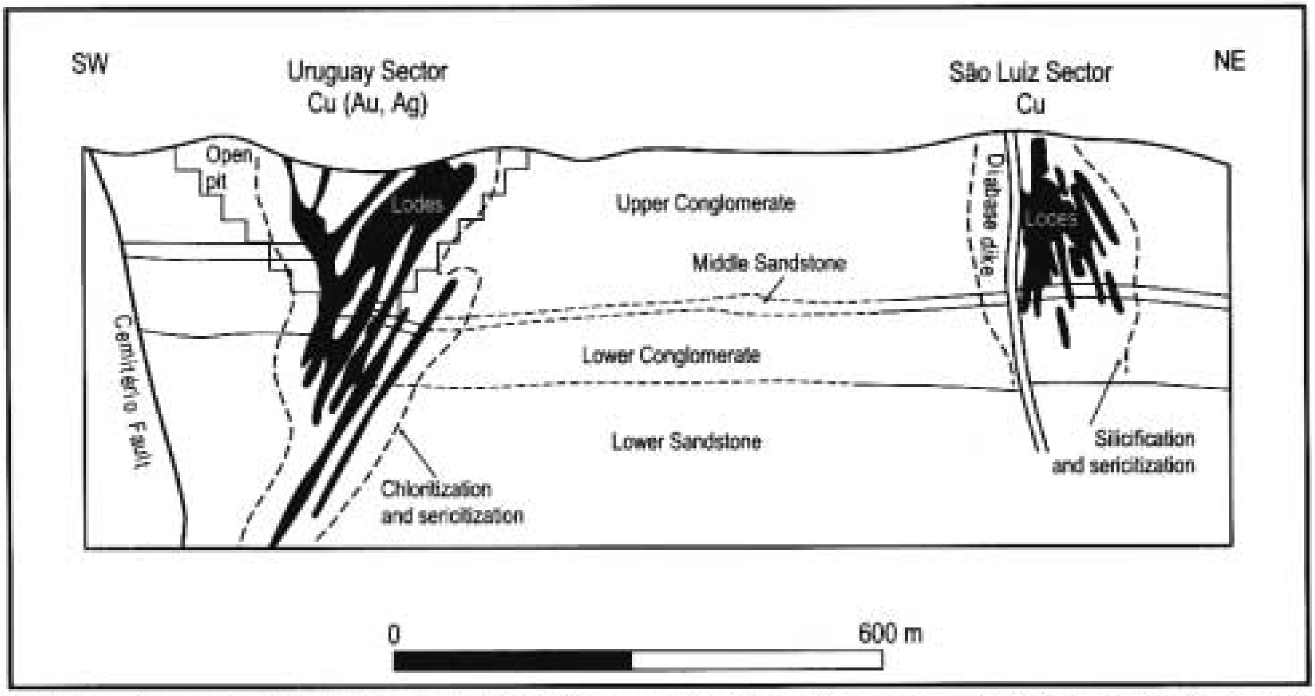


FIGURE 30 - Schematic cross-section of the Uruguay and São Luiz sectors of the Camaquã Cu (Au, Ag) Mine (modified after Teixeira and Gonzalez, 1988).

the Santa Maria Deposit with paragenesis of galena-sphalerite-chalcopyrite (Veigel and Dardenne, 1990). The reserves are given as about 33.4 Mt of ore at an average grade of 1.44% Pb, 1.08% Zn and 12 to 14 g/t Ag (Badi and Gonzalez, 1988).

3) The secondary mineralization occurs as oxide, and the cementation phases with the associated mineral species hematite-bornite-chacocite-covellite at the Camaquã mines, and hematite-bornite-chalcocite-stephanite at the Santa Maria Deposit (Veigel and Dardenne, 1990).

The origin of the Cu-Au and Pb-Zn-(Cu)-Ag mineralization of the Camaquã region has been the subject of discussion and several hypotheses for this have been proposed: a) epigenetic hydrothermal mineralization resulting from the canalization of connate waters, heated by volcanism along the NW-SE faults (Veigel and Dardenne, 1990); b) epigenetic hydrothermal mineralization without specifying the source of the mineralizing fluids (Ronchi *et al.*, 1999); c) epithermal mineralization associated with the intrusion of the Lavras do Sul and Caçapava do Sul granites (Bettencourt, 1976; Remus *et al.*,1999).

The Borborema Province

The Borborema Province or Northeastern fold region (Fig. 31), which resulted from the Brasiliano collage, is a complex mosaic of Neoproterozoic fold belts and a basement nucleus attributed to the Transamazonian collage (Van Schmus *et al.*, 1995). The main structures have a fan-shaped geometry permitting the division of the Province into five domains (Brito Neves *et al.*, 1999): Median Coreaú Domain; Northern Domain; Transversal Domain; and the Southern Domain.

The main mineral deposits found in the Borborema Province are:

- Magnesite deposits at José de Alencar in the Orós Belt, Ceará, associated with a Paleo-Mesoproterozoic volcano-sedimentary sequence (2.0-1.7 Ga), interpreted as evaporite deposits (Parente and Guillou, 1995; Parente and Arthaud, 1995; Parente, 1995);

- Copper deposits of the Martinópole volcano-sedimentary sequence (Pedra Verde) in the State of Ceará, interpreted here to be of the sedimentary-exhalative or SEDEX-type.

-Gold deposits in shear zones, associated with quartz veins and skarn with paragenesis W-Mo-Au in the Seridó Province of the states of Rio Grande do Norte and Paraíba, described by Melo and Legrand (1993); Legrand *et al.* (1993, 1996), Souza Neto *et al.* (1996), Melo *et al.* (1996). At the São Francisco Deposit, to the E of Currais Novos, the gold mineralization is associated with biotite-garnet schist and occurs as the result of successive hydrothermal phases that accompany the metamorphic evolution. This resulted in mineralization in auriferous veins, coeval with shearing. The deposits contain about 1.75 t Au (Silva and Legrand, 1996; Ferran, 1988)

- The Scheelite Province of the State of Rio Grande do Norte (Salim, 1993);

- The uranium deposit in the region of Itataia in the State of Ceará;

- The Pegmatite Province of Seridó, associated with Brasiliano Cycle granite intrusions, dated at 555 Ma (Legrand *et al.*, 1991; Legrand *et al.*, 1993). The pegmatite bodies are intruded along the foliation striking NNE-SSW in mica schist, and contain Ta, Nb, Li, Be and Sn (Silva and Dantas, 1997).

The Seridó Scheelite Province

The Seridó Scheelite Province in the states of Rio Grande do Norte and Paraíba contains a number of deposits including Brejuí, Barra Verde, Boca de Lage, Bodó, Parelhas and Bom Fim (Fig. 32). These deposits have been mined

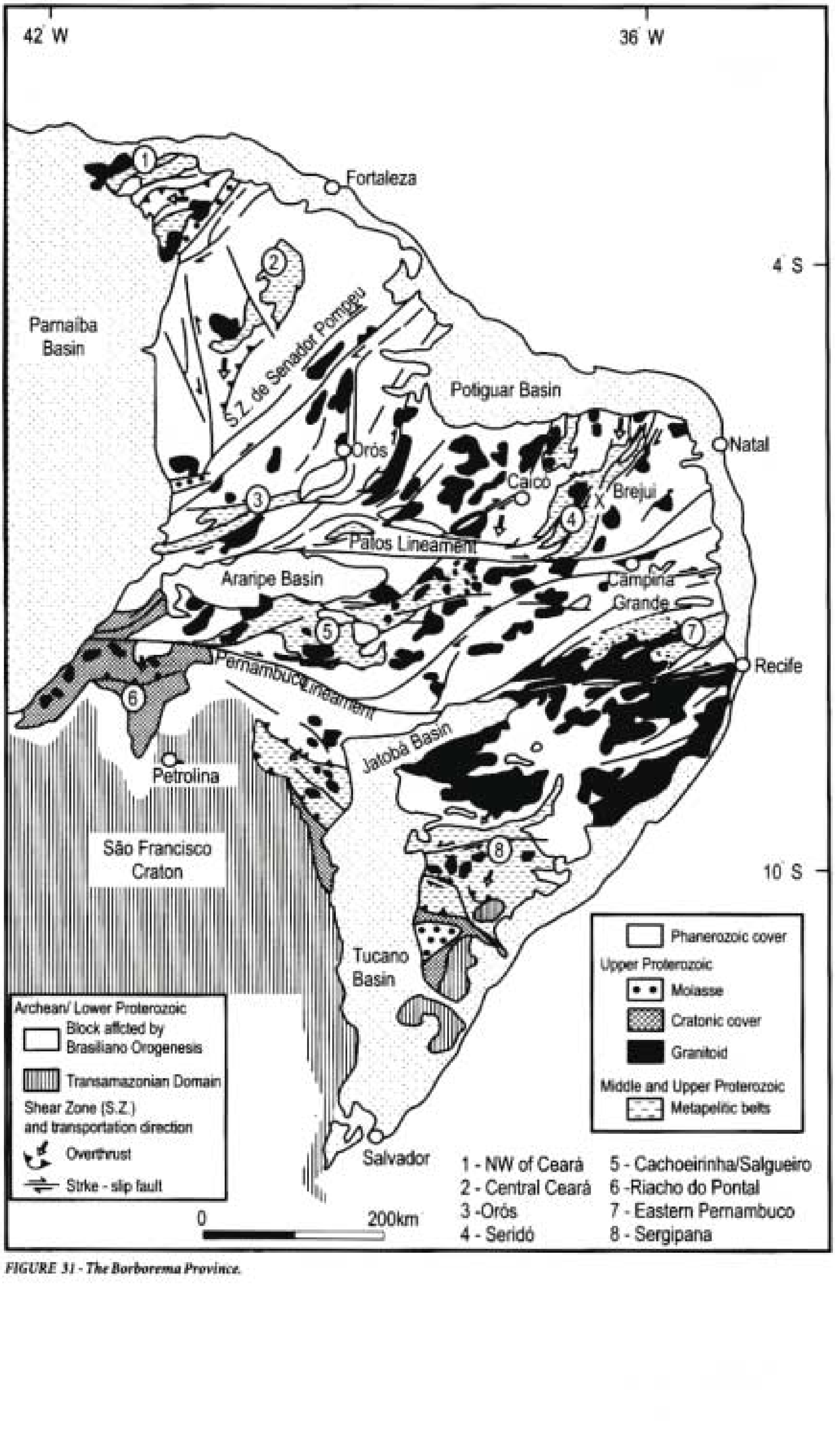


FIGURE 31 - The Borborema Province.

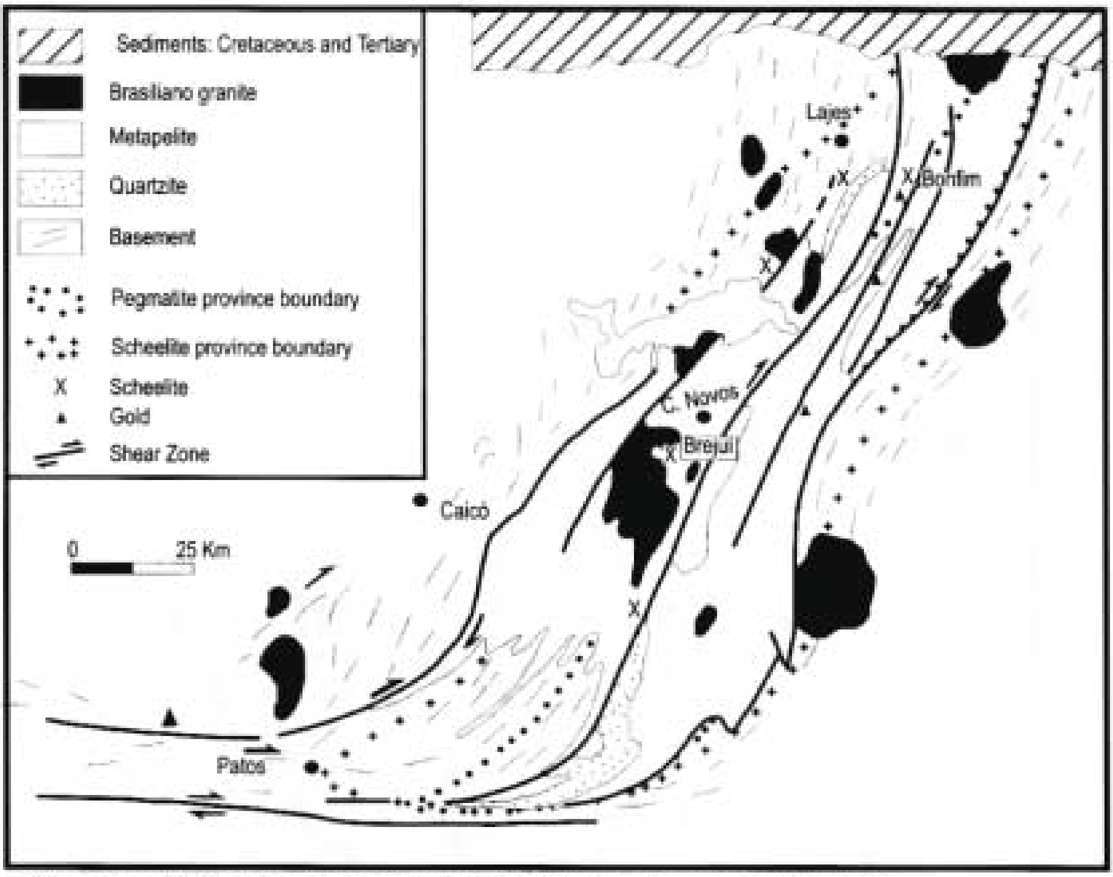


FIGURE 32 - Distribution of mineral deposits in the Seridó Belt, PB-RN (modified after Legrand et al., 1996).

intensely from the Second World War up to 1985. Production is estimated at 60 000 t of concentrate (70% WO3) from ore containing 0.7% to 1% WO3. In the region occurs a basement gneiss known as the Caicó Complex, which is generally considered to be of Archean age, reworked during the Transamazonian Cycle, and a metasedimentary sequence known as the Ceará Series and/or Seridó Series (Jardim de Sá and Salim, 1980; Archanjo and Salim, 1986). This metasedimentary sequence is divided into three formations or groups according to different authors: the Jucurutu Formation consisting mainly of finely-banded biotite gneiss, rich in epidote, intercataled with quartzite, mica schist, marble and amphibolite associated with banded iron formation units in which are found the principal scheelite deposits; the Equador Formation consisting of muscovite-rich quartzite with intercalated paragneiss, mica schist and marble; the Seridó Formation consisting of a thick sequence of aluminous mica schist with intercalated paragneiss, quartzite and marble with a unit known as the Parelhas Conglomerate at the base. The age of the metamorphism and the plutonism of the Seridó Series is dated at between 650 and 550 Ma. The intrusive rocks are related to an early, pre- to syn-tectonic dioritic suite dated at between 600 and 500 Ma (Jardim de Sá *et al.*, 1986; Legrand *et al.*, 1991; Leterrier *et al.*, 1990).

In the Seridó Province the principal scheelite mineralization is associate with skarn deposits of the Jucurutu Formation, and the most important mines (Brejuí, Barra Verde, Boca de Lages) are situated around the Acari Granite Massif (555 Ma) (Legrand *et al.*, 1991) at the contact marble/metasediment and marble/granite in the Jucurutu paragneiss. Two types of scheelite are described in the various mines in the region: fine-grained scheelite orientated along the foliation of the primary skarn deposits, generally considered of early genesis; coarse-grained scheelite, related to late hydrothermal alteration in retrograde skarnite, forming high-grade concentrations in the hinges of vertical folds, and in shear zones associated with these folds (Salim, 1993). Locally, the scheelite occurs in quartz veins and pegmatite that cut the mineralized skarnite. The scheelite is of variable colour from white to yellow. There also occurs a black variety related to the presence of thin laths of molybdenite, pyrite, chalcopyrite and bornite. The accessory minerals are magnetite, native bismuth and bismuthinite. The sulphide minerals have developed in zones of late hydrothermal alteration with minerals of zeolite paragenesis substituting the silicate minerals of the primary skarnite (plagioclase, amphibole, diopside, garnet) and secondary skarnite (scapolite, vesuvianite, epidote). Gold is often observed associated with molybdenum and tungsten (Au-Mo-W) in the skarnite of the Seridó Province (Legrand *et al.*, 1993, 1996; Melo and Legrand, 1993; Melo *et al.*, 1996; Souza *et al.*, 1996).

The Itataia Uranium Deposit

The Itataia Deposit (Forman and Waring, 1981) is situated in the Municipality of Santa Quitéria some 220 km from the city of Fortaleza in the State of Ceará. It occurs in a metamorphic sequence consisting of migmatite at the base, overlain by quartzite and gneiss and covered by crystalline carbonate known as the Itataia Group (Mendonça *et al.*, 1985), to which a Paleoproterozoic age has been assigned. Brasiliano pegmatite and granite intrude it.The uranium mineralization is associated with episyenite, the intrusion of which resulted in the sodic metasomatism of the gneiss; to massive collophanite; and to collophanite stockworks filling fractures in marble and carbonaceous breccia (Mendonça *et al.*,1985).The collophanite occurs as microcrystalline fluorapatite, limpid and in spherulites with a fibro-radial structure, intimately associated with masses of cryptocrystalline collophanite. The age of the mineralization is considered to be Brasiliano to Cambro-Ordovician. The measured reserves are 79.5 Mt of ore at 11% P_2O_5 and 1000 ppm U_3O_8. It is very similar to the deposits of Espinharas in the State of Paraíba. It may be related to ringed igneous granite intrusions of the Itaperuaba-type, dated at between 550 and 450 Ma, which show albitization accompanied by uranium mineralization (Haddad, 1981).

Phanerozoic Platform Cover and Associated Magmatism

During the Phanerozoic (Fig.33) the evolution of the South American Platform was dominated in the Paleozoic by the development of huge intracratonic syneclises represented by the Amazonas-Solimões, Parnaiba, Paraná and Chaco-Paraná basins in which the sedimentation began in the Silurio-Ordovician and ended at the close of the Permian (Milani and Zalán, 1999). In the Mesozoic there occurred the final in-filling of the Paleozoic basins. Rifting related to the opening of the North Atlantic Ocean in Triassic-Jurassic times, and the opening of the South Atlantic Ocean in the Cretaceous, led to the formation of basins on the Brazilian continental margin as well as in isolated Cretaceous basins in the northeastern region of Brazil. During the Cenozoic lateritic weathering profiles developed over the South American Platform from the beginning of the Tertiary. Finally, there occurred marine sedimentation in marginal basins along the Brazilian coast, and the deposition of fluvial continental sediments in the interior.

Paleozoic Deposits

The mineral resources of the Paleozoic basins (Fig.33) are very limited and restricted to the following: occurrences of Devonian oolitic ironstone formation units in the basal part of the Pimenteiras Formation in the Parnaiba Basin (Ribeiro and Dardenne, 1978; Ribeiro, 1984), in the Jatapu region of the Amazonas Basin (Façanha da Costa, 1966; Hennies, 1969), and in the Serra do Roncador region of the Paraná Basin, showing a Devonian metallogenetic phase for this type of deposit; potassium deposits associated with Permo-Carboniferous evaporites in the Amazonas Basin; coal and pyrobituminous schist deposits in the Permian sediments of the Paraná Basin (see Lopes and Ferreira, this volume); and the Figueira uranium deposit, likewise associated with the Permian sediments of the Paraná Basin.

The Potassium Deposits of Fazendinha and Arari of the Middle Amazonas Basin

From the Silurian to the end of the Devonian the Amazonas Basin underwent a marine transgression from E to W. Following a period of generalized flooding, there occurred a slight tilt of the basin to the W with concomitant uplift along part of the eastern margin, causing an inversion of the direction of the marine invasion, that now came from W to E with the deposition of a transgressive sequence (Monte Alegre and Itaituba formations), followed by a phase of very restricted circulation resulting in the deposition of the evaporite sequence of the Nova Olinda Formation, with which are associated the potassium deposits at Fazendinha and Arari, transitional to the continental sediments of the Indirá Formation (Upper Permian). At this time the Amazonas Basin (*sensu lato*) (including the Solimões Basin) was divided from E to W by the physical barriers of the Iquitos, Purus and Gurupá highs into the Juruá (Upper Amazonas) and Middle Amazonas sub-basins (Fig. 34). According to Sad *et al.*, 1982, 1997), the cyclic re-occurrence of high and low salinity, separated by clastic sedimentation or by less soluble chemical sediments such as limestone and anhydrite, has permitted the separation of the evaporitic sequence into 11 cycles. Cycle VII marked the period of greatest restriction of the evaporite basin with highly saline brines and the deposition of finely crystalline banded halite with high bromine content (>70 ppm Br), culminating with the precipitation of potassium and magnesium salts in the form of chloride and sulphate. Cycle VII terminated with fresh water incursion from continental sources bringing about the development of continental lacustrine conditions.

In the Fazendinha region, the mineralized beds are sub-horizontal and lie at a depth of 980 m to 1140 m below the surface. Their average thickness is 2.7 m and the KCl content varies between 14.31% and 38.69% (average 27%). The potassium-rich zone is divided into three intervals:

- A lower interval with milky white sylvinite, finely laminated that grades, transitionally, to an overlying sequence of finely banded halite beds, implying a primary origin for the chemical precipitation of the sylvinite. This interval is 1 to 1.8 m thick and contains 29.7% KCl.
- An intermediate interval denominated the sulfate zone, in function of the presence of minerals such as kainite, kieserite, leonite, langbeinite, polyhalite and anhydrite associated with halite and silvinite, implying a marine transgression bringing solutions rich in calcium and sulfate. This interval is between 0.5 m and 1.6 m thick and contains 20% KCl;
- An upper interval consisting of coarse-grained red sylvinite with irregular beds of anhydrite and discontinuous beds of halite. These features together with the absence of lamination lead to the interpretation that the original mineral was carnallite that underwent sylvinitization as the result of preferential leaching of magnesium, thus leading to the view that the silvinite is secondary. This interval is 0.80 m thick and contains 32% KCl, and is overlain by pink coarse-grained halite.

The measured *in situ* reserves of the Fazendinha Deposit exceed 520 Mt of ore at 28.8% KCl, permitting the recovery of 36 Mt of KCl, whereas, the reserves of the Arari Deposit are about 659 Mt with 17.7% KCl.

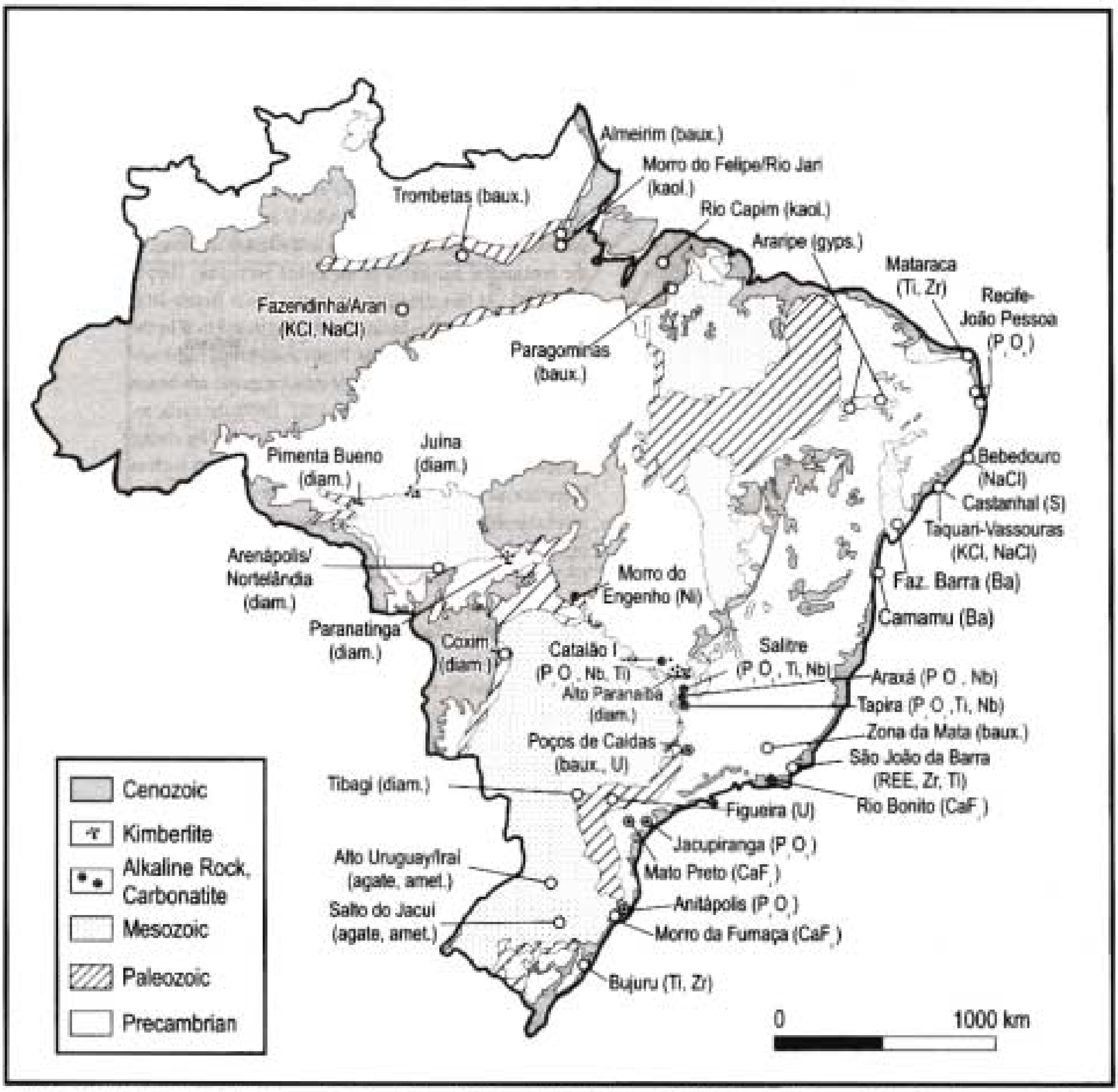


FIGURE 33 - Distribution of main Phanerozoic mineral deposits in Brazil (modified after Schobbenhaus and Campos, 1984, and other sources cited in text). Abbreviations: amet. - amethyst; baux. - bauxite; diam. - diamond; gyps. - gypsite; kaol. - kaolin; pyr.sh. - pyrobetuminous shale; REE - Rare Earths.

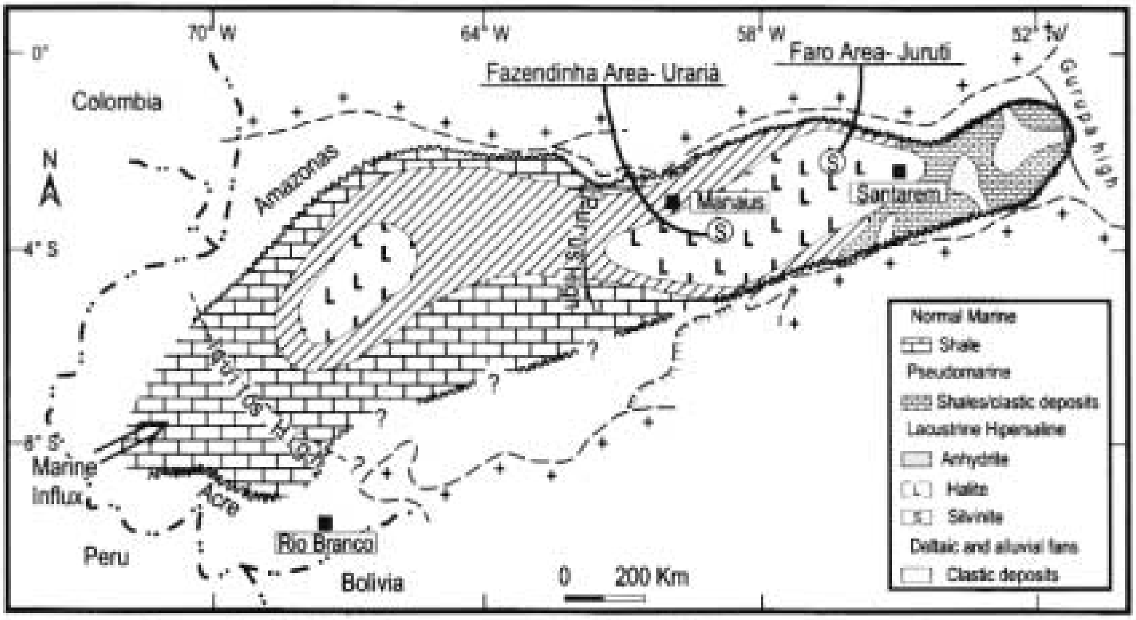


FIGURE 34 - Paleogeography of the Carboniferous Amazonas Basin showing the distribution of the evaporite facies (modified after Sad et al.,1997).

The Figueira Uranium Deposit (PR)

The uranium deposit is associated with the Triunfo Member at the base of the Rio Bonito Formation of the Paraná Basin. The deposit is confined between a coal bed at the base and a medium to coarse-grained sandstone at the top. The deposit is tabular in shape and is peneconcordant (Morrone and Daemon, 1985). In the sandstone unit the mineralization appears in the interstices of the quartz grains in the form of uraninite, intimately associated with pyrite and other sulphide species such as chalcopyrite and sphalerite. The molybdenum (average grade 0.2%) and the vanadium (200 to 500 ppm), together with selenium, nickel and germanium anomalies accompany the uranium mineralization. The reserves are estimated at about 8000 t of U_3O_8. From the paleogeographical point of view the sediments that host the Figueira Deposit and other occurrences known in the region (Telêmaco Borba and Sapopema) are associated with a system of island barrier lagoons with the development of peat deposits (Della Favera *et al.*, 1993). The evolution of the organic matter during burial led to the development of a reducing environment favourable to the fixation of uranium in its oxide form. The uraninite is associated with pyrite, and fills the secondary porosity and substitutes the wall rock sandstone. The anomalies for Mo, V, Se and Ge, associated with the uranium mineralization suggest an initial concentration of these elements together with organic matter.

Mesozoic Deposits

The mineral resources found in the Mesozoic basins and associated structures (Fig. 33) are related directly or indirectly to a global tectonic event that came about by rifting that lead to the break-up of Gondwana and the separation of the African and South American continents. The tectono-sedimentary evolution of the rifting process was divided into four stages (Ojeda, 1981, Cainelli and Mohriak, 1999): pre-rift stage; taphrogenic rift stage; transition stage of the proto-oceanic transgressive gulf during which time occurred the deposition of evaporite deposits; and the transgressive open ocean stage corresponding to a phase of thermal subsidence. These stages, which refer to the progressive opening of the South Atlantic Ocean, had important reflexes in the interior of the South American Platform as the result of successive reactivation along ancient lineaments, as well as by the appearance of new tectonic structures and the individualization of regional uplift.

The mineral deposits related to this tectono-sedimentary evolution have been classified in five categories (Dardenne, 1999): deposits related to volcanism; deposits associated with ultramafic-alkaline carbonatite complexes; deposits associated with kimberlite and lamproite intrusions; hydrothermal vein deposits; sedimentary deposits.

Deposits Associated with Volcanism

At the beginning of the Cretaceous, between 140 and 120 Ma, the pre-rift stage in the Paraná Basin is characterized by vast flows of tholeiitic basalt, basaltic andesite and minor dacite and rhyolite related to continental fissural volcanism of the Serra Geral Group. In Rio Grande do Sul, the Alto Uruguai/Araí and Salto do Jacuí amethyst and agate deposits are related to this volcanism (Fig.33). These deposits are of great economic importance, and have been intensely worked (Schmitt *et al.*, 1991; Szubert *et al.*, 1978; Cassedanne, 1991; Castro *et al.*, 1974). The origin of the silica required for geode formation is related to the dissolution of the silica in intertrap sandstone by supercritical water liberated by the crystallization of the basalt.

In the State of Piauí, the opal deposits at Pedro II are associated directly or indirectly to the circulation of hydrothermal waters originated by the intrusion of diabase sills (Orozimbo Formation) in the Paleozoic sediments of the Cabeças Formation (Devonian). These diabase sills are considered to be of Triassic-Jurassic age, corresponding to volcanic manifestations that accompanied the rifting that preceded the opening of the North Atlantic Ocean.

The mineralization occurs in fractures and siliceous breccia related to the shale and sandstone beds, and even the diabase at the base of the sill (Rosa, 1988; Samama *et al.*, 1983; Roberto and Souza, 1991; Cassedanne, 1991).

Deposits Associated with the Ultramafic-Alkaline-Carbonatite Complexes

According to Amaral *et al.* (1967), Hasui and Cordani (1968), Ulbrich and Gomes (1981), Cordani and Hasui (1968), these ultramafic-alkaline-carbonatite intrusive complexes show two groups of ages (Fig. 35) the first group occured between 130 and 120 Ma, and includes the complexes of Jacupiranga, Juquiá, Ipanema, Barra do Itapirapuã and Anitápolis, which are concentrated in the southeastern region of Brazil, and are contemporaneous with the basaltic volcanism of the Paraná Basin; and a second group, occurring between 90 and 65 Ma, including the complexes of Iporá, Santa Fé, Catalão, Serra Negra, Salitre, Tapira, Araxá, Poços de Caldas and Mato Preto, related to the re-activation along an Upper Cretaceous rift (Barbosa *et al.*, 1970). These complexes, which are distributed on the margin of the Paraná Basin along lineaments and regional uplift, occurred between the Lower and Upper Cretaceous, and are of great importance in the Brazilian mineral economy in function of the deposits associated with these. The mineral deposits are intimately related to the magmatic evolution of the complexes (Figs. 36, 37), and consist mainly of phosphate in the form of apatite, as well as magnetite, niobium, titanium, vermiculite, barite, fluorite, uranium and rare earths elements (CBMM, 1984).

In all the deposits associated with these complexes, the laterite weathering has played a fundamental role in the economics of the deposits, tending to increase the grades in the weathering profile by two mechanisms: a) relative concentration of the resistate minerals in the laterite cap, principally pyrochlore and apatite; b) neoformation of nickel minerals (silicate and oxide, enriched in Ni), alumina (gibbsite in the bauxite) and titanium (anatase).

The Diamond Deposits Associated with Kimberlite Intrusions

The principal occurrences of kimberlite in Brazil are distributed along the Transbrasiliano Lineament and a lineament with azimuth 125° (Gonzaga and Tompkins, 1991; Tompkins and Gonzaga, 1989). The Transbrasiliano Lineament is associated with the Gilbués/Picos kimberlite in the State of Piaui and the Poxoréu kimberlite in the State of Mato Grosso. To the 125° lineament is related the Cretaceous kimberlite provinces of the Paranatinga region (Batovi Kimberlite, dated at 121 Ma); of the Aripuanã region (Juína Kimberlite) in the State of Mato Grosso; of Pimenta Bueno in the State of Rondônia and of Alto Paranaiba (dated at 70 Ma) in the State of Minas Gerais (Bizzi, 1993). Only the Juína Kimberlite (Teixeira, 1998) has shown significant diamond mineralization (Fig. 38).

Hydrothermal Vein Fluorite Deposits

In southern Brazil the veined fluorite mineralization (Dardenne and Touray, 1988; Dardenne *et al.*, 1997) occurs in the State of Santa Catarina in a belt about 30 km wide mainly containing the Brasiliano granite intrusives of Pedras Grandes and Tabuleiro in addition to dykes and sub-volcanic acid rocks (Eopaleozoic) and mafic rocks (diabase of the Serra Geral Formation) and Paleozoic sediments of the Paraná Basin. The fluorite veins that cut all the rock-types mentioned above are related to ancient Brasiliano lineaments striking NNE-SSW that were reactivated during the Cretaceous as transcurrent and normal faults: the Canela Grande, Grão Pará and Armazém lineaments (Bastos Neto, 1990, Bastos Neto *et al.*, 1992). Four main phases of mineralization may be observed. The first three phases occurred between 140 and 100 Ma, whereas the last phase corresponds to a late phase and dates at about 70 Ma (Bastos Neto *et al.*, 1992). These four phases are well defined in function of the tectonic evolution of the veining and the rare earth content in the several generations of fluorite. Fluorite, associated with chalcedony, occurs in the form of coarse to fine-grained banded structures; as hydraulic breccia and as cockade ore, all very characteristic (Dardenne and Savi, 1984; Bastos Neto, 1990).

The genetic model that is generally accepted for the generation of these veins (Savi and Dardenne, 1980) involves the leaching of the fluor of the regional granitoid rocks by circulating hydrothermal waters related to the thermal anomaly associated with the rifting along the margins of the South Atlantic Ocean.

In the southeastern region some of the larger fluorite veins are associated with alkaline rocks of the Taguá Massif, dated at 65 Ma. The fluorite deposits have characteristics that are similar to those of the Santa Catarina District (Becker *et al.*, 1997).

In Uruguay, vein deposits of fluorite show the same controls as those observed in Santa Catarina.

Sedimentary Deposits

The Cretaceous sedimentary deposits may be divided into four categories (Dardenne, 1999): clastic diamondiferous deposits; evaporite deposits; phosphatic deposits; and deposits associated with the circulation of connate waters.

Clastic Diamond Deposits

During the rift stage of the Lower Cretaceous, thick sequences of fluvial clastic deposits were layed down in the marginal coastal basins and interior basins of Brazil. In the Alto São Francisco Basin, the basal conglomerate beds of the Abaeté Formation (Areado Group) contain numerous occurrences of detrital diamond (Campos *et al.* 1993), the origin of which is related to the successive reworking of older deposits, principally those of the Sopa-Brumadinho Formation of the Espinhaço Supergroup, and those alluvial deposits probably associated with diamictite of the Macaubas and Santa Fé de Minas groups (Campos, 1996). The diamond occurrences of the Abaeté Formation, which

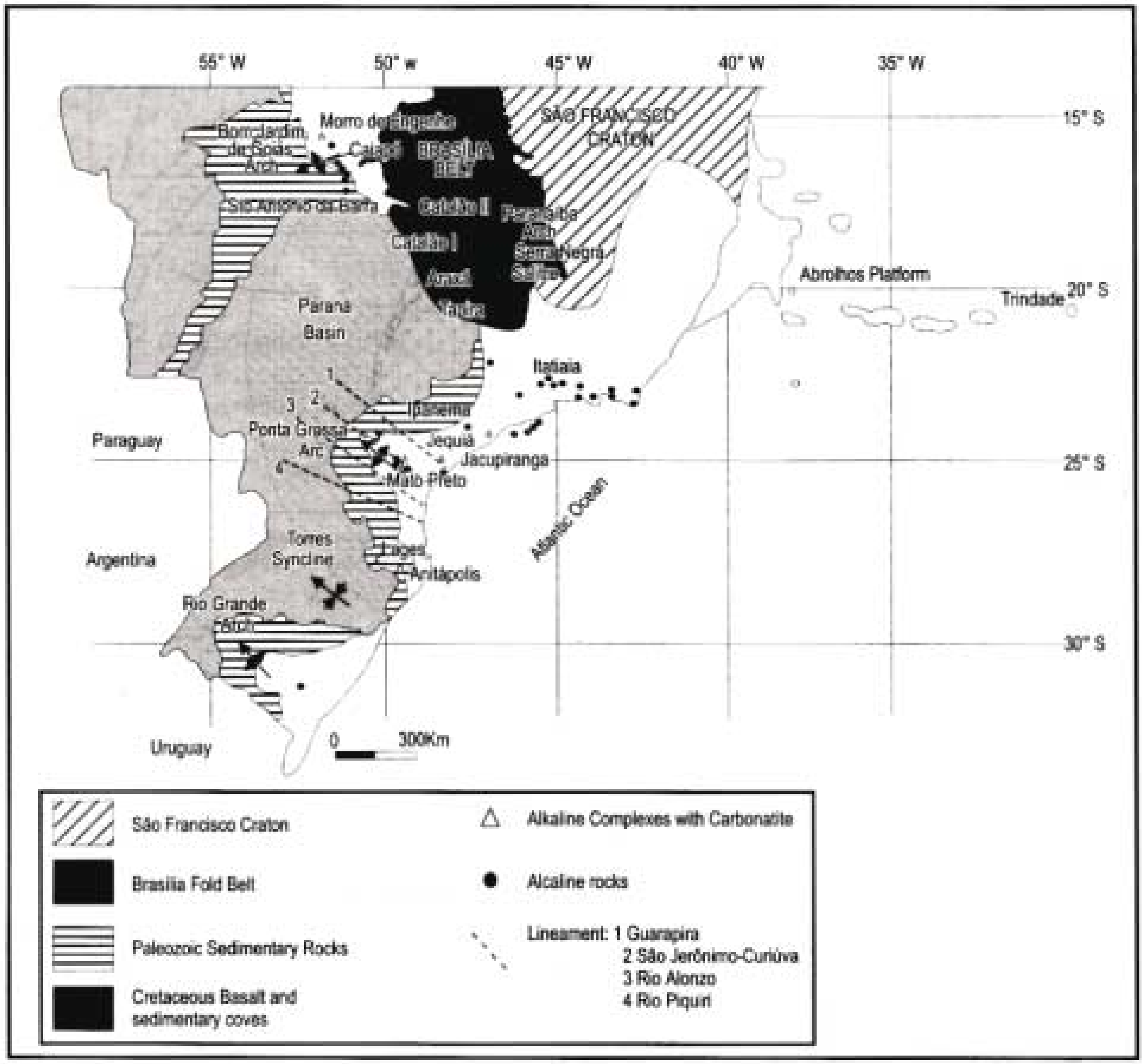


FIGURE 35 - Alkaline complexes in central - western and Southeastern Brazil.

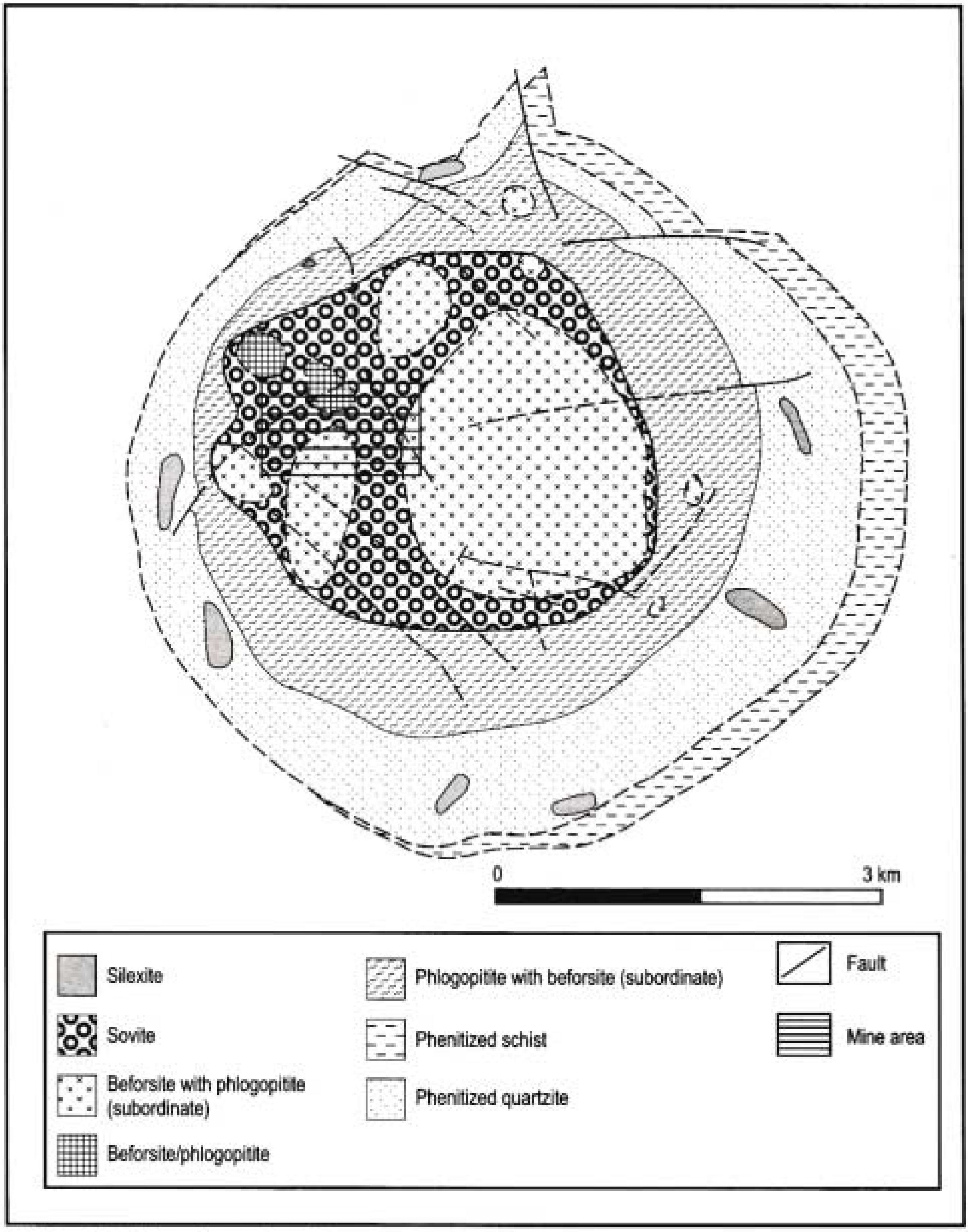


FIGURE 36 - Geological Map of the Barreiro Complex.

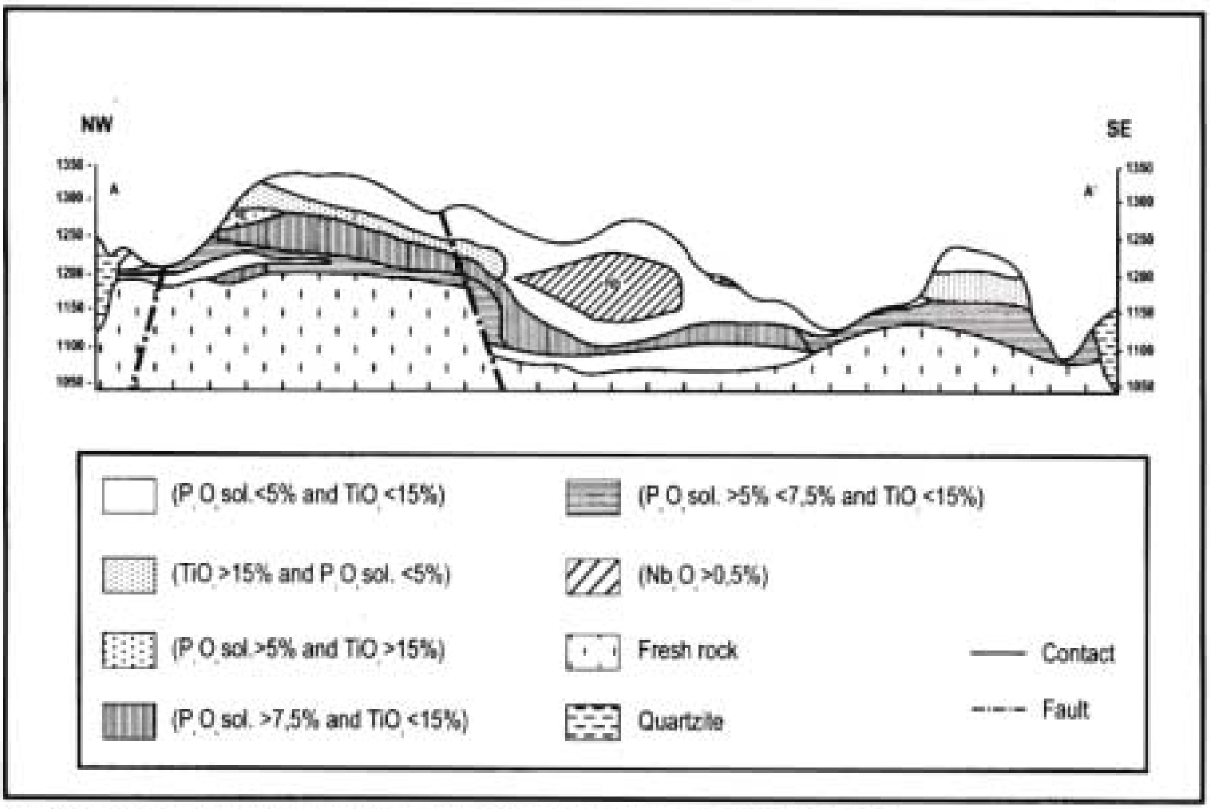


FIGURE 37 - Niobium, phosphate and titanium mineralizations of the Tapira Complex (CVRD - DOCEGEO).

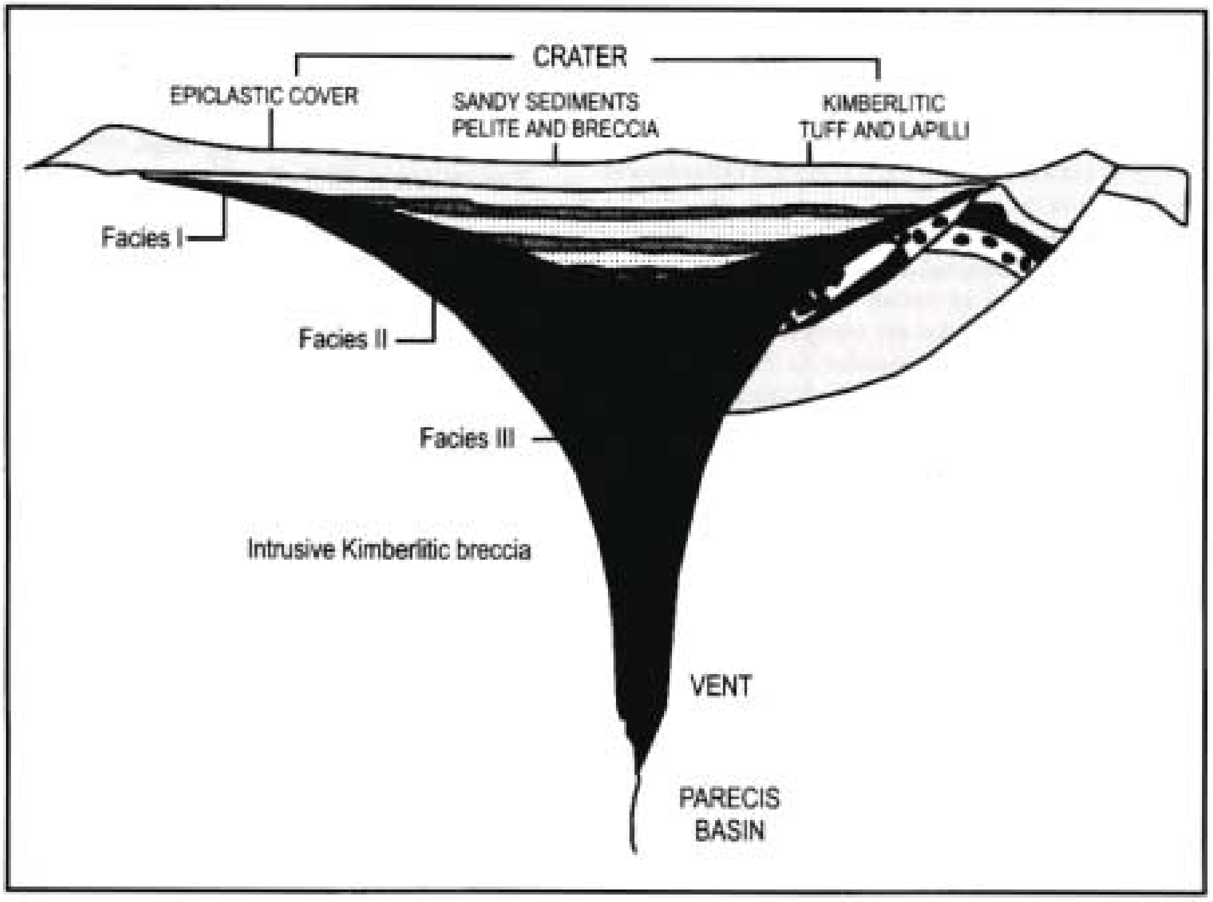


FIGURE 38 - Explosive kimberlite structure of Juina (modified after Teixeira, 1998).

have been amongst those most investigated and evaluated, are situated in the Cana Brava region of the State of Minas Gerais (Dardenne *et al.*, 1991; Campos, 1996) and the Serra do Cabral where the highest grades obtained reach a few points per m^3.

In the Brazilian Northeastern region, similar deposits are found in the Gilbués area of the State of Piaui where the diamonds are associated with the conglomeratic facies of the Pé do Morro Formation at the base of the Areado Group (Gonzaga and Tompkins, 1991).

During the re-activation of the Upper Cretaceous rift, there occurred an uplift of the Paranaiba High between the Paraná and São Francisco basins with which are related the alkaline intrusions, volcanic flows and kimberlite pipes of the Patos Formation. The erosion of these volcanic edifices resulted in the deposition of conglomerate and epiclastic sandstone units of the Uberaba Formation in the Bauru Basin, and the Capacete Formation in the Alto São Francisco Basin. To these rock-types are associated the diamonds of the Romaria Deposit (Gallo, 1991; Suguio *et al.*, 1979) and the provenance of inumerous occurrences found in Recent alluvium.

In the State of Mato Grosso the diamondiferous placers are related to the erosion of conglomerate beds of the Upper Cretaceous Parecis Formation (Arenápolis/Nortelândia diamond district; 400 000 carats at 0.02-0.04 ct/m^3) and the Bauru Group (Poxoréu diamond district; 0.04 ct/m^3) deposited, respectively, in the Parecis and Bauru basins; both separated by the Rondonópolis High (Schobbenhaus 1984; Souza, 1991; Fleischer, 1976; Weska *et al.* 1997; DNPM-Brazilian Mineral Yearbook, 1998).

Evaporite Deposits

In the northeastern region of Brazil, evaporite deposits occur in the marginal coastal basins as well as in the Cretaceous basins of the continental interior.

In the marginal coastal basins, the evaporite sedimentation of Aptian age developed during a transitional phase that also represented a proto-oceanic gulf phase, related to the evolution of a rift at the start of the opening phase of the South Atlantic Ocean. The marine transgression occurred from S to N when the oceanic waters flowed over the Walvis barrier at the latitude of Rio Grande do Sul. In this basin were deposited two evaporite sequences contained in the Muribeca Formation: the Paripueira Evaporites of Eo-aptian age, that are worked for halite by underground dissolution methods at Bebedouro near the city of Maceió in the State of Alagoas (Amaral and Melo, 1997); the Ibura Evaporites of Neoaptian age, which are associated with deposits of carnallite and sylvinite in the Santa Rosa de Lima and Taquari-Vassouras sub-basins of the Sergipe Basin (Fig. 39). Cycle VII consists of beds of halite and sylvinite with thin zones of carnallite, which are mined at the Taquari-Vassouras underground mine (Szatmari *et al.*, 1979; Cerqueira *et al.*, 1986, 1997). The lower sylvinite unit is yellow in colour and crystalline. It is 3.82 m thick, and contains 25.03% KCL. The upper sylvinite unit is reddish and whitish in colour and finely crystalline. The average thickness of this unit is 4.27 m and it contains 24.95% KCl. A halite zone having a maximum thickness of 14.62 m in the central part of the deposit separates these two sylvinite units, and locally the two units may merge to form a single sylvinite unit. The sylvinite reserves are given as about 42 Mt with 24.95% KCl. The annual production in 1991 was 275 000 t of KCl and 850 000 t of NaCl.

At the Siririzinho Anticline, which separates the two basins, there occurs the native sulphur deposit of Castanhal (Frota and Bandeira, 1997; Morelli *et al.*, 1982), situated in the lower part of the Ibura Member where it is intimately associated with the biogenic reduction of anhydrite beds in the presence of oil, water and sulphurous gas. In the interior basins of the Brazilian Northeast, the evaporite deposits consist mainly of gypsum and subordinate anhydrite, precipitated during an Aptian marine transgression. The main gypsum deposits, mined for the manufacture of Portland cement and plaster, occur in the Santana Formation on the Chapada do Araripe (Krauss and Amaral, 1997; Silva, 1988), and in the Codó Formation in the Maranhão Basin (Baquil, 1997).

The Phosphate Deposits of the Pernambuco-Paraiba Basin

In the Brazilian Northeastern Region, phosphate is associated with sedimentary sequences (Paraiba Group) of the Pernambuco-Paraiba Basin (Upper Cretaceous), extending as a narrow coastal belt some 15 to 20 km wide, N-S, for a distance of about 100 km between the cities of Recife and João Pessoa, and dipping gently towards the Atlantic Ocean. Along this coastal belt the phosphate beds indicate a marine transgression at the base of the Gramame Formation. These beds are essentially continuous and overlie the Beberibe Sandstone. The thickness of the phosphorite beds varies from a few centimetres to a maximum of 4 m, with grades between 20% and 35% P_2O_5 (Kegel, 1955; Moreira Neto and Amaral, 1997). The estimated reserves for the region are about 65 Mt of ore at 22% P_2O_5.

In the high-grade phosphorite the phosphatic material including moulds of mollusks, planktonic foraminifera, intraclasts, pellets, ooliths, cropolites, algal and coral fragments is abundant (Tinoco, 1971). The phosphorite consists essentially of fluorapatite with low CO_2 content (1.14% to 1.38%), high F/P_2O_5 ratios (0.195 to 0.146) (Boujo *et al.*, 1998; Menor *et al.*, 1977). The phosphorite has a certain amount of radioactivity, representing, according to White (1957) equivalent uranium grades of 0.018% to 0.25%.

Deposits of Barite associated with the Circulation of Connate Fluids

In the Cretaceous basins of Sergipe/Alagoas, Camamu, Recôncavo and Tucano, there are numerous occurrences of barite, galena and sphalerite related to circulation of connate fluids in rift environments (Dardenne, 1997, 1999). At the barite deposit of Fazenda Barra (Bandeira *et al.*, 1986) the barite originated by the replacement of an anhydrite cement in a sandy zone by barite carried in percolating waters rich in barium.

At the Camamu Deposit (Dardenne and Campos, 1984), genesis resulted from the replacement by leaching of an anhydrite zone of Eo-aptian age by barite derived from barium-rich solutions, in turn derived from the feldspar of acid granulite.

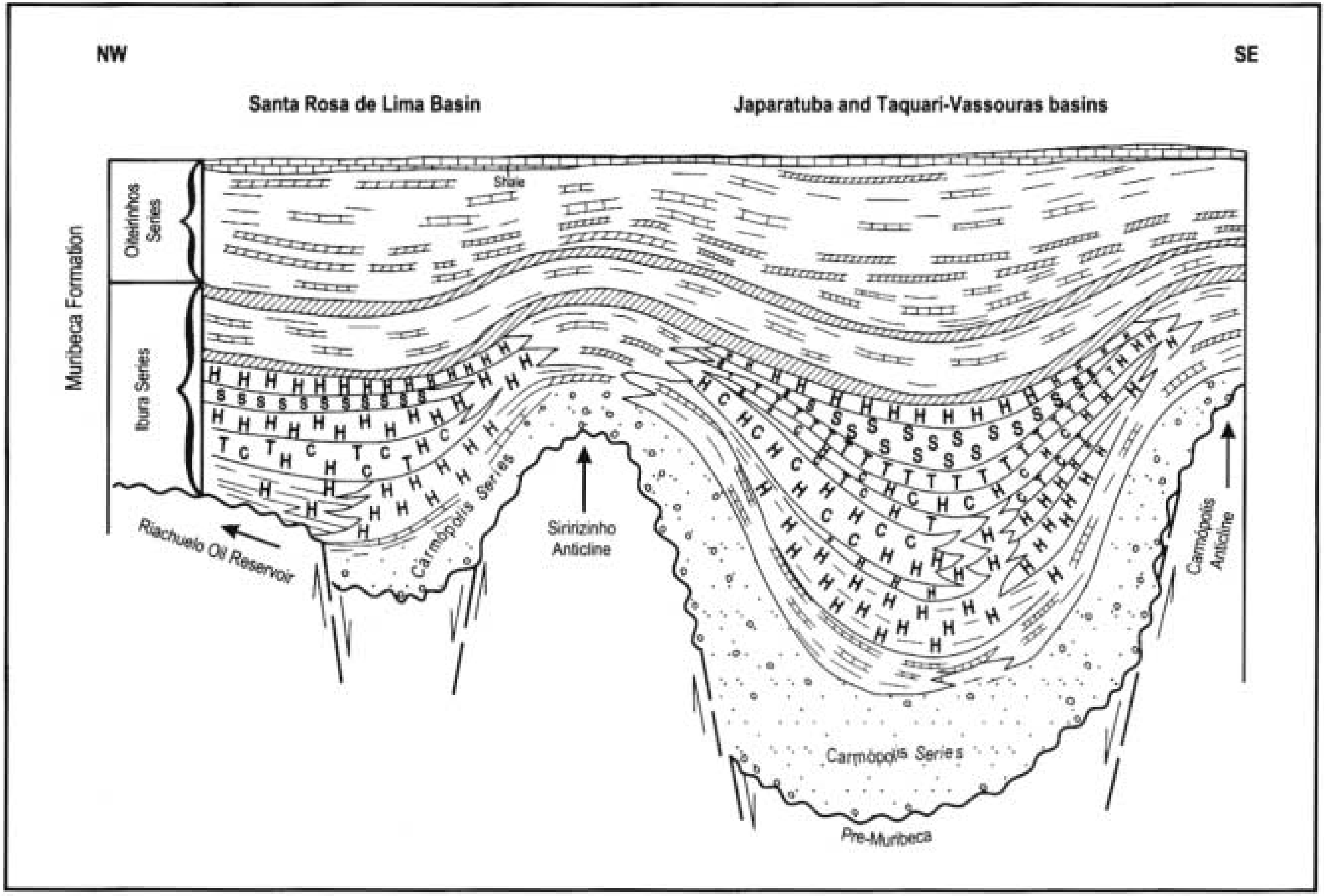


FIGURE 39 - Geological cross-section of Santa Rosa de Lima and Taquari-Vassouras basins, Sergipe-Brazil (modified after Cerqueira et al., 1997).

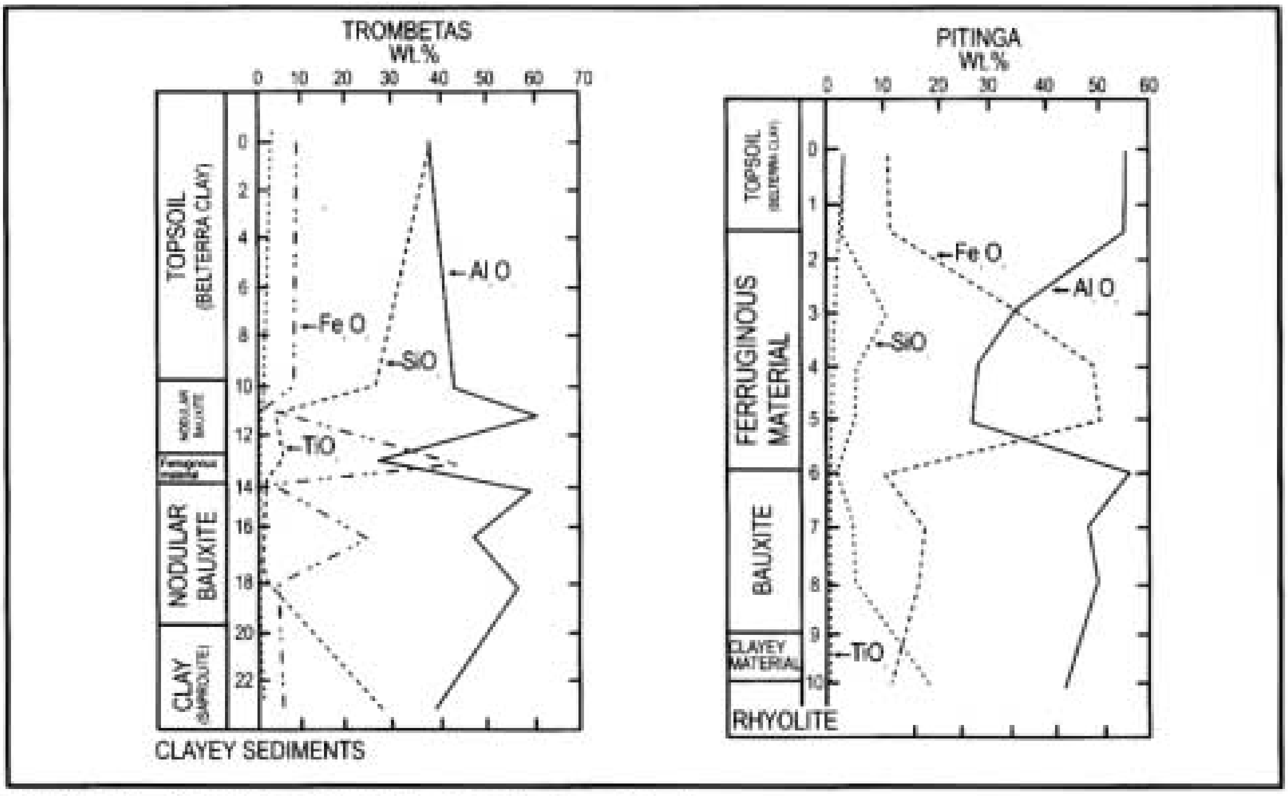


FIGURE 40 - Chemical variations in the lateritic bauxite profiles at Trombetas and Pitinga, modified after Boulange and Carvalho (1997).

Cenozoic

During the Cenozoic the principal mineral deposits have resulted from the weathering of laterite that developed on the South American Platform from the beginning of the Tertiary, and the accumulation of heavy mineral concentrates as placer deposits in alluvium, paleo-alluvium and in paleo-marine sediments along beaches at the continental margin.

Mineral Deposits of Lateritic Origin

A large part of the South American Platform is situated in the intertropical zone in which there predominate conditions favourable for the development of mechanisms of supergene alteration, leading to the development of thick lateritic cover. However, the main phase of lateritization, with which are associated the mineral deposits is related to the lower Tertiary (Eocene), and the second phase is generally attributed to the upper Tertiary (Miocene-Pliocene), which suggests a complex evolution and a fossil character for most of the mineral concentrations of lateritic origin. These minerals may be classified in two categories in function of the nature of their origin: a) lateritic deposits originating by relative concentration of their insoluble or slightly soluble chemical elements in the form of neoformed minerals. The most significant of these are aluminum, manganese, nickel, and very locally, titanium; b) lateritic deposits originating from residual accumulations of stable minerals, resistant to supergene alteration. These include hematite in itabirite, cassiterite in albitite; pyrochlore, apatite and barite in alkaline-carbonatite complexes and locally gold.

In this review, only those lateritic deposits belonging to the first category will be discussed, whereas the importance of the deposits included in the second category will be stressed before the genesis of the primary mineralization is discussed.

Bauxite Deposits

The bauxite deposits are widely dispersed in different regions of the South American Platform, and may be divided into three main provinces: a) Eastern Amazon Basin; Berbice Basin in Guyana and Suriname; c) Los Pijiguaos in Venezuela; d) Central-eastern region of Brazil; e) Southeastern region of Brazil.

The bauxite deposits of the Eastern Amazonas Basin

In the states of Amazonas and Pará, the main bauxite deposits: Trombetas, Nhamundá, Juruti, Almeirim, and Paragominas, display similar weathering profiles (Fig. 40) developed from argillaceous sediments and argillaceous-sandy sediments of the Ipixuna, Itapecuru and/or Alter do Chão formations, of Lower to Upper Cretaceous age. The distribution of the zones in the weathering profile may have a bearing on the origin of the bauxite of Amazonia, which is still somewhat controversial:

- the bauxite is overlain by a thick kaolinitic cap (up to 20 m) known as the Argila de Belterra. This deposit is considered to be allochthonous, having being deposited in a lacustrine environment (Grubb, 1979; Truckenbrodt and Kotschoubey, 1981; Kotschoubey *et al.*, 1981, 1997); or autochthonous, having developed *in situ* in the weathering profile (Lucas, 1997; Boulangé and Carvalho, 1997; Aleva, 1981).

- the presence of a nodular ferruginous zone intercalated between two bauxitic zones, suggesting polyphasic evolution, involving climatic diversity with alternating humid and dry periods (Kotschoubey *et al.*, 1997), or alternatively, the migration of iron through the weathering profile to form an intermediate ferruginous crust (Lucas, 1997; Boulangé and Carvalho, 1997; Aleva, 1981). These observations have led to two distinct models: the allochthonous model implying an evolution involving climatic diversity, and the autochthonous model implying polyphasic alteration *in situ*. proposed a similar model for the evolution of bauxite deposits on the Guiana Shield. The deposits of Eastern Amazonia contain huge reserves of bauxite.

The total reserves of the Eastern Amazonas region exceed 1.5 billion t of ore.

The bauxite deposits of the Berbice Basin

The coastal basin of Berbice in Guyana and Suriname contains the world's largest reserves of high-grade gibbsitic bauxite suitable for refractory liner requiring a very low iron content, as well as for usage in the chemical and metallurgical industries. It seems that these bauxite deposits developed in the Paleogene and some were covered by Oligocene sediments (Gibbs and Barron, 1993).

In Guyana, the deposits are situated in the Berbice and Demerara river regions, and in Suriname, at Moengo, Onvervacht and Paranam (Fig. 5).

The Los Pijiguaos bauxite deposits

The Los Pijiguaos bauxite deposits are the most important in Venezuela. These deposits are developed over the Parguaza rapakivi granite of Mesoproterozoic age. The high-grade ore occurs at an erosion level between 620 m and 690 m, and formed during an intense weathering cycle during the Upper Cretaceous and lower Tertiary. Measured and indicated bauxite reserves are 201.8 Mt of ore at 48.7 % Al_2O_3, and 10.9% SiO_2.

The Parguaza Granite is a batholith that covers an area of at least 10 000 km^2 in the State of Amazonas extending into Venezuela. The granite is intruded into the volcanic rocks of the Cuchivero Group (Uatumã Supergroup) (Sidder and Mendoza, 1995). The granite protolith contains 65% to 73% SiO_2, and 13.5% to 15% Al_2O_3. The average thickness of the Los Pijiguaos ore is about 7.5 m, and the overburden, when present, is <1 m. The main mineral is gibbsite with smaller amounts of kaolinite. Structural and physical features such as joints, fractures and the slope angle control the Al_2O_3 enrichment and the depletion of SiO_2 in the granitic rocks. The ore zones enriched in alumina and depleted in silica may be correlated with zones having the highest density of fractures. A slope with inclination between 2° and 10° is considered favourable for the formation of ore (Sidder, 1995).

The bauxite deposits in the Central-eastern region of Brazil (Zona da Mata)

The bauxite deposits in the Central-eastern region of Brazil, also known as the Zona da Mata, are found in the southern part of the State of Minas Gerais and on the elevated terrains of the Serra da Mantiqueira. The bauxite deposits have developed over different Precambrian rock-types. Of note, are the deposits associated with granulitic rocks in the region of Cataguazes in Minas Gerais, which constitute an extensive aluminous belt orientated NE-SW, between the towns of São João do Nepomuceno and Cataguazes. The total reserves exceeding 100 Mt are of great economic importance in function of their strategic situation near to the large markets of Rio de Janeiro, São Paulo and Belo Horizonte (Roeser *et al.*, 1984; Valeton and Melfi, 1988; Valeton *et al.*, 1991; Beissner *et al.*, 1997).

The bauxite deposits of the Alkaline Province of Southeastern Brazil

Bauxite deposits originated from the chemical weathering of alkaline rocks occur in the Poços de Caldas Province, the Coastal Province of Rio de Janeiro and São Paulo and the Province of Lages-Anitápolis in the State of Santa Catarina. The more important reserves are associated with the Poços de Caldas Alkaline Complex, and are estimated at about 50 Mt (Schulmann *et al.*, 1997).

The Kaolin Deposits of the Amazon Region

There are three main districts in the Amazon region known for their kaolin deposits: Rio Capim, Morro de Felipe and Manaus-Itacoatiara (Costa and Moraes, 1998). The more important deposits developed as the result of the *in situ* alteration of Cretaceous sediments of the Ipixuna-Itapecuru and Alter do Chão formations (Murray and Partridge, 1982). The thickness of the kaolinitic zone varies from 10 to 20 m (Fig. 41). The kaolin deposits are characterized by their whiteness in function of their low iron oxide and hydroxide content, and are used mainly in the paper industry. Production from several mines at Morro do Felipe and from the Rio Capim District is about 2 M tpa. According to Costa and Moraes (1998) and Kotschoubey *et al.* (1996), the kaolin deposits are related to the lower zones of the laterite profile that developed initially in the lower Tertiary and evolved progressively by desferrification and resilicification in reducing and acid environments, principally in the Oligocene-Miocene transition.

Nickel Laterite Deposits

The principal nickel laterite deposits are found in the Amazon region, and specifically in the Carajás Mineral Province. Here the deposits are associated with differentiated intrusive bodies, dated at 2.645 Ga, including Vermelho, Puma-Onça and Jacaré-Jacarezinho. In the southwestern region of Brazil there occur deposits related to the mafic-ultramafic complexes of the Niquelândia and Barro Alto, dated at between 2.0 and 1.7 Ga, and to the ultramafic-alkaline complexes of Upper Cretaceous age such as Santa Fé de Goiás, Morro do Engenho, Morro dos Macacos, Rio dos Bois and Montes Claros.

In the nickel laterite deposits of the Vermelho-type (Fig. 42) (Alves *et al.*, 1986) the weathering profiles are developed over peridotite and serpentinized dunite with pyroxenite intercalations where the nickel is concentrated in ferruginous zones with limonitic ore (1.2% Ni) as well as in coarse-grained saprolite (1.5% to 2.0% Ni) as silicate ore in the form of garnierite and smectite (1.5% to 2.0% Ni). The relative amounts of the two ore-types are approximately the same (Costa, 1997; Bernadelli *et al.*, 1983; Castro Filho and Matos, 1986). The reserves at the Vermelho deposit are estimated at 44 Mt of ore at 1.5% Ni.

At the nickel laterite deposits found at the Niquelândia

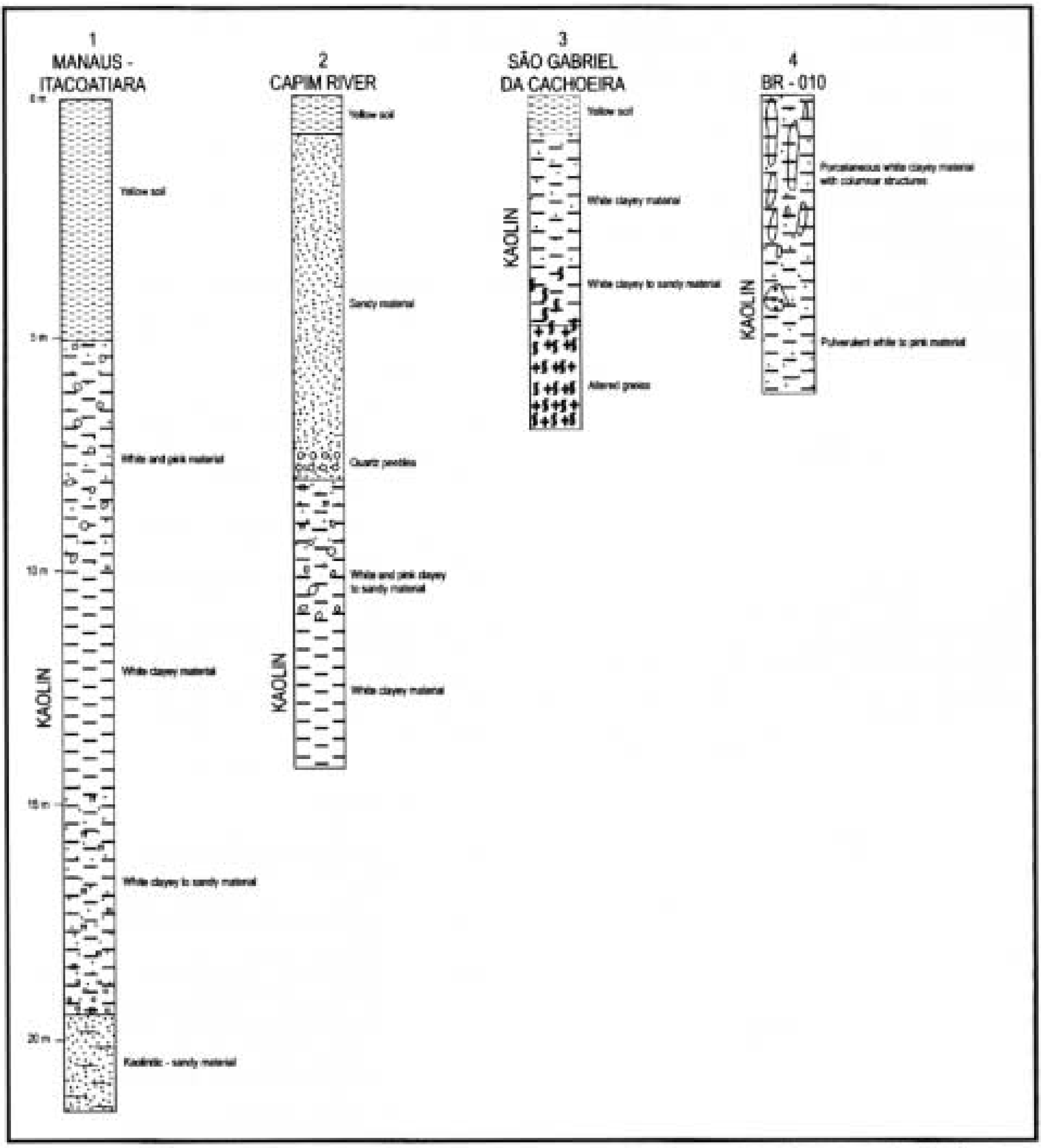


FIGURE 41 - Geological profiles of kaolin deposits (modified after Costa,1997): 1 - Manaus-Itacoatiara; 2 - Capim River; 3 - São Gabriel da Cachoeira; 4 - BR - 010 (Belém-Brasília).

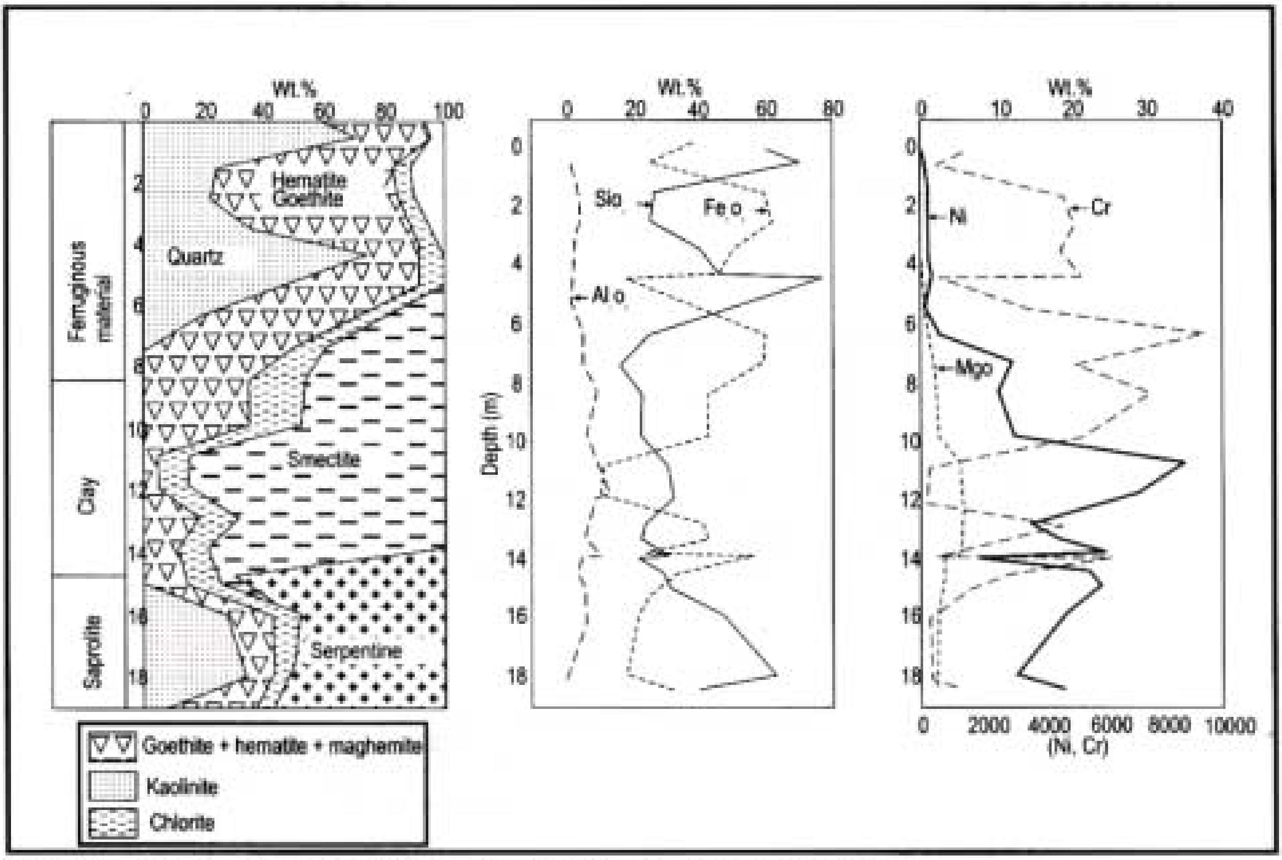


FIGURE 42 - Mineralogical and chemical variations through the nickel lateritic deposit of Vermelho/Carajás.

and Barro Alto complexes (Baeta *et al.*, 1986) the weathering profiles are developed over partially serpentinized peridotite and dunite and pyroxenite, with nickel concentration on the oxide ore, rich in goethite, and in the silicate ore, rich in smectite. The polyphasic evolution of these profiles (Fig. 43) may be observed in all the deposits of the Central-Western region of Brazil (Oliveira and Trescases, 1982; Melfi *et al.*, 1980, 1988). The reserves at the Niquelândia Complex are estimated at about 60 Mt of ore at 1.45% Ni, and have been intensely mined by Niquel Tocantins and CODEMIN, whereas the reserves at the Barro Alto Complex have been evaluated at about 72.39 Mt of ore at 1.67% Ni.

Finally, there is the Rincón Del Tigre igneous complex in Bolivia, near the frontier with Brazil. This is a differentiated mafic-ultramafic layered intrusion some 3000 to 4000 m thick, dated at 993±139 Ma. Tertiary weathering cycles produced a secondary nickel concentrate over the serpentinized dunite of the complex. Extensive resources of secondary nickel silicate ore have been proved (Litherland *et al.*, 1986).

Lateritic Gold Deposits

During lateritic weathering the gold was partially or totally remobilized in profiles developed over primary mineralization, resulting in very high gold concentrations, as can be seen at the Igarapé Bahia Deposit in the Carajás Mineral Province, and at the Cassiporé Deposit in Amapá (Fig. 44) (Costa, 1997; Costa *et al.*, 1993, 1996).

Placer Deposits

Placer deposits developed during the Cenozoic in drainage in the interior of the South American Platform as well as along the littoral regions led to the mechanical concentration of heavy minerals.

Gold and Cassiterite Placers in the Amazon Region

The economic importance of gold and cassiterite placer deposits is very great. Gold concentration along drainage occurs in alluvial and paleo-alluvial deposits in the proximity of primary deposits in the mineral provinces of Amapá, Tapajós, Rio Madeira and Alta Floresta. The cassiterite provinces are Pitinga and Rondônia. Whereas the heavy mineral concentrates in Recent alluvium are of limited economic importance, this is not the case with the terrace and buried paleo-alluvial deposits (Figs.45 and 46) that may be of economic value, locally (Bastos, 1988; Bettencourt *et al.*, 1997; Veiga *et al.*, 1988; Veiga, 1988).

Beach Placers along the Brazilian Coast

Along the Brazilian coastline from the NE to S, occur placer deposits in beach sands with monazite (REE oxide), ilmenite/rutile and zirconite. Ilmenite deposits are also found along the littoral of Argentina.

The more important Brazilian deposits of this type are found along the littoral of the states of Paraíba, Bahia, Espirito Santo and Rio de Janeiro. The largest deposit of monazitic sands occurs at São João da Barra in the State of

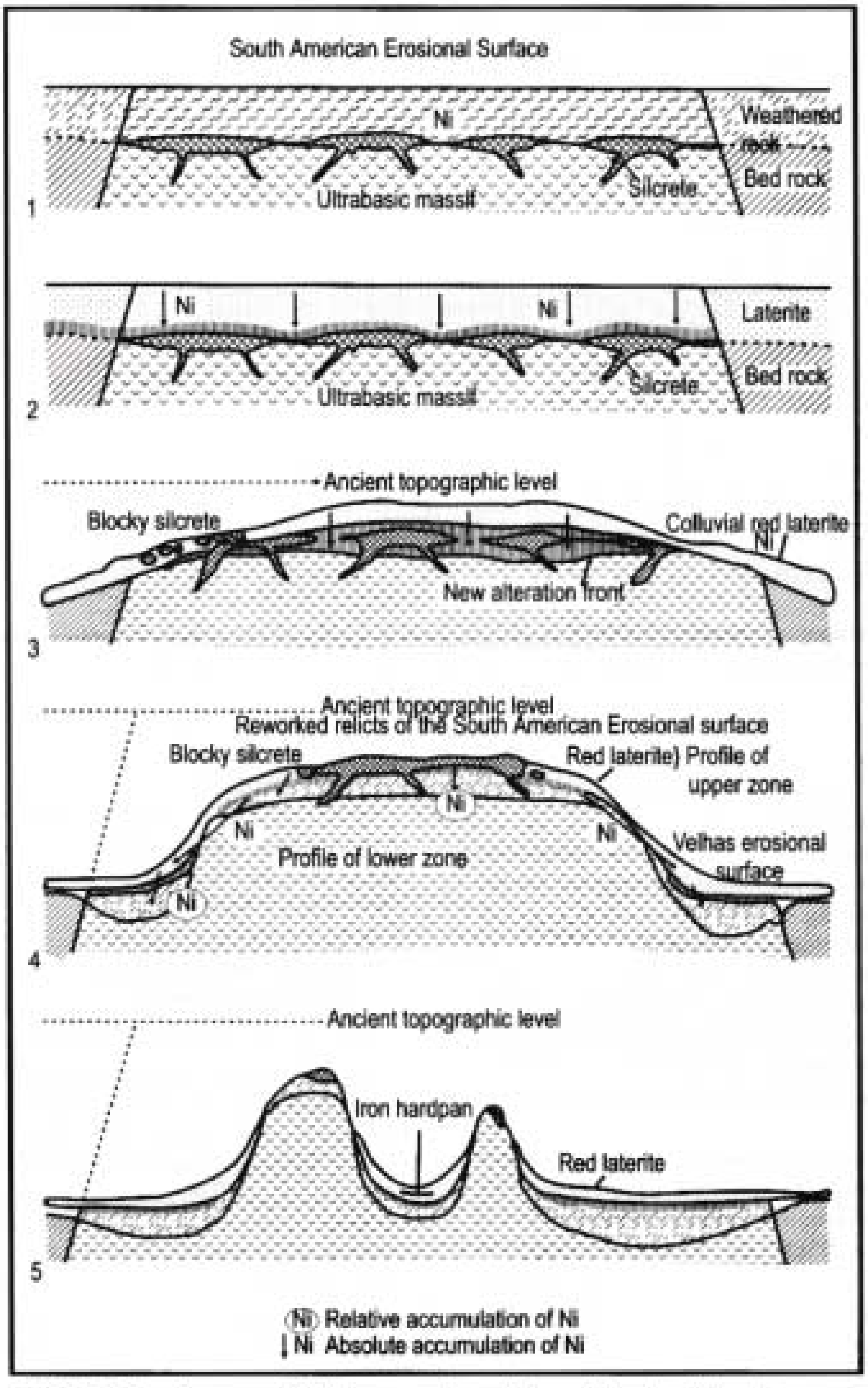


FIGURE 43 - Schematic sequence of the different morphotectonic phases during the evolution of the Ni lateritic deposits in Central Brazil (modified after Melfi et al.,1988).

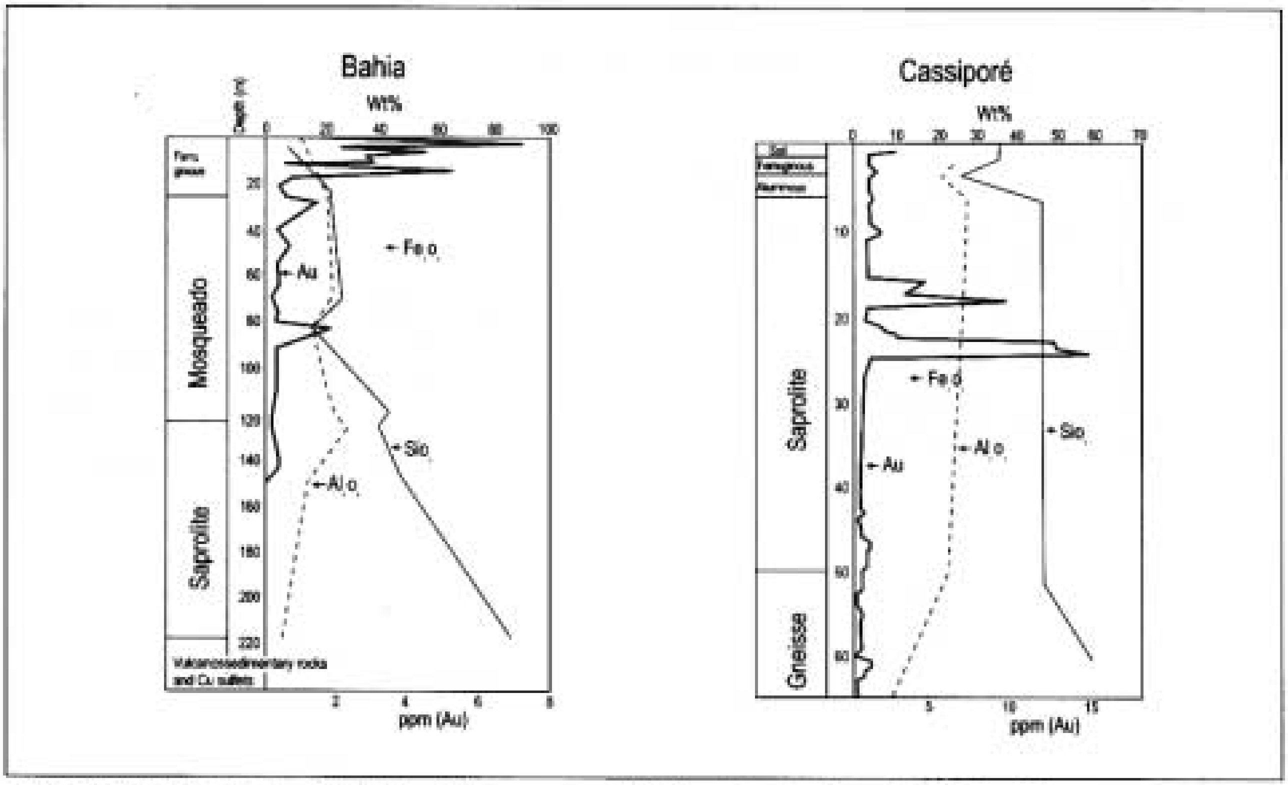


FIGURE 44 - Vertical distribution of principal chemical components and gold in lateritc deposits of Igarapé Bahia (Carajás) and gold mines of Cassiporé (modified after Zang and Fyfe 1993; Costa et al., 1993)

Rio de Janeiro. It contains about 40 000 t of monazite. Mataraca, in the State of Paraíba, has the largest deposits and is the largest Brazilian producer of ilmenite concentrate (c. 100 000 tpa) and zirconite (Source: DNPM). However, the largest recently evaluated total reserves of ilmenite occur at Bujuru in the State of Rio Grande do Sul with 10.8 Mt of ilmenite (Santos et al., 1998).

At Mataraca, the average grade of the heavy minerals varies between 3% and 5%. The measured reserves at Mataraca are 2.7 Mt of ilmenite (81.54%), rutile (2.4%) and zirconite (16.06%).

The Distribution of Mineral Deposits Throughout Geological Time on the South American Platform – Metallogenic Epochs

During the development of the South American Platform from the Archean to the Proterozoic, as well as during its tectonic evolution during the Phanerozoic, a number of mineral deposits were formed. The synthesis shown in Figure 47 gives a general view of the chronostratigraphic position of the principal mineral deposits in relation to the major tectonic events, as well as an indication of the principal metallogenic epochs occurring on the platform.

The definition of the metallogenic epochs, this is to say, the geological time interval during which the formation of the mineral concentrations of a certain metal or substance

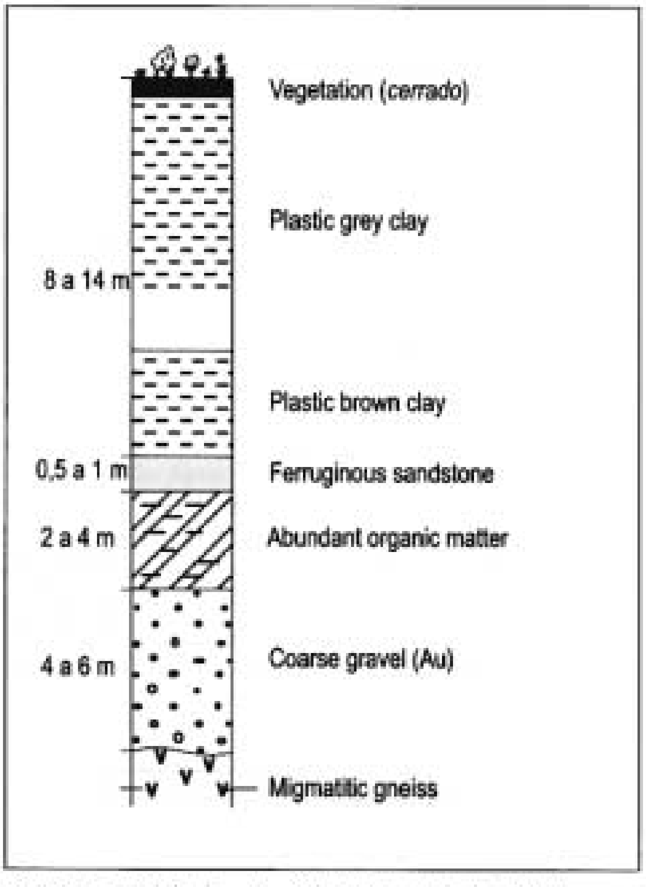


FIGURE 45 - Burried paleovalley of Rio Madeira: Periquito gold mine.

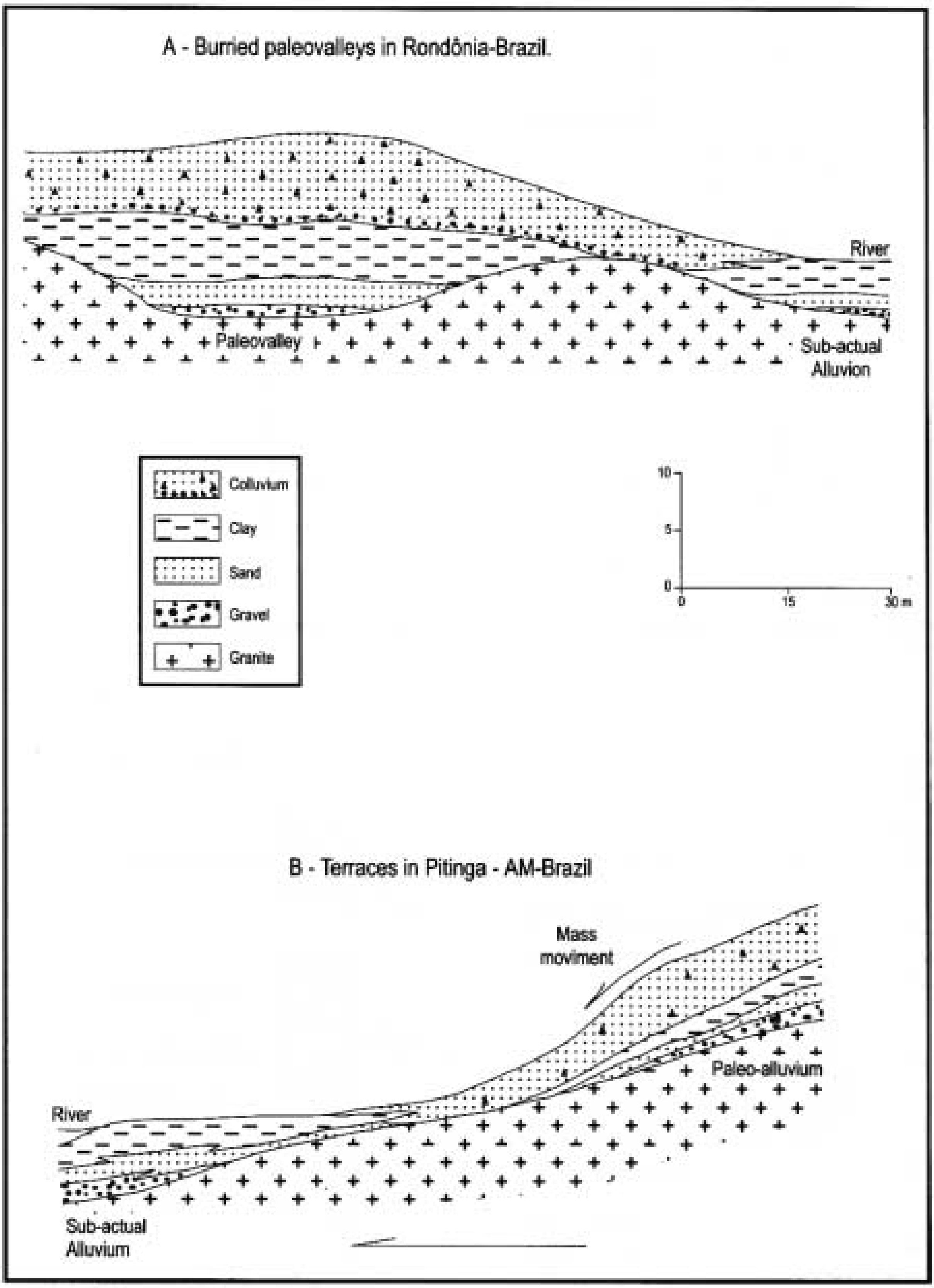


FIGURE 46 - Schematic model for Au and Sn placers (modified after Veiga et al., 1988).

was especially favourable or pronounced, remains relatively difficult for certain regions of the South American Platform, seeing that this implies the application of a time-bounded concept. The inherent difficulties lie in a more precise level of definition, not only of the mineralization, but also of the geochronological positioning of the host unit and/or generator of the mineralization. This view is especially valid for the oldest units that underwent a complex geological history.

In spite of these difficulties it is now possible to position the most important mineralizations in their geochronological context, and relate this to specific tectonic and/or magmatic events that were especially favourable for the generation of certain mineral deposits or groups of deposits. This may be attributed to the advances in recent years in metallogenic and geochronological studies in the more important mineral provinces/districts of the South American Platform.

In this setting, it can be noted that throughout time there occurred regional specialization of certain groups of mineral deposits reflecting crustal evolution and highlighting the limits between the chronostratigraphic units and emphasizing the heterogeneity of the primitive crust and mantle (Dardenne, 1982, Schobbenhaus, 1984; Schobbenhaus and Campos, 1984; Delgado *et al.*, 1994; Tassinari and Melito, 1994).

In the **ARCHEAN**, three major divisions may be defined: Paleo-Archean, Meso-Archean and Neo-Archean. In the Meso-Archean and Neo-Archean, the development of volcano-sedimentary sequences and associated plutonism was omnipresent, defining distinct metallogenic epochs in function of their age and metals content.

In the **Paleo-Archean** there occurred the individualization of the oldest continental block recognized on the South American Platform between 3.4 and 3.7 Ga that is represented by the Imataca Block in Venezuela. In the Imataca Block, rocks of this age include banded iron formation units of the Superior Province-type or more probably associated with volcanism (Algoma or Carajás type), and secondarily, of the Algoma-type, deformed and metamorphosed in the granulite and amphibolite facies between 2.8 and 2.7 Ga (Aroense Event), and between 2.15 and 2.0 Ga (Transamazonian Event). Supergene alteration of these rocks resulted in important concentrations of iron ore (e.g. Cerro Bolivar, San Isidro, El Pao). Manganese mineralization (gondite) is also found associated with the Imataca Complex.

In other areas of the South American Platform there are indications of primitive crustal continental nuclei older than 3.0 Ga. However, the related metallogenic epoch has not yet been defined.

In the **Meso-Archean**, between 3.0 and 2.8 Ga there occurred the development of the oldest granite-greenstone terranes, and the formation of continental blocks in the areas of Rio Maria (Central Brazil Shield), Crixás (Goiás Massif), and Pium-hi, Fortaleza de Minas and the Gavião Block (Atlantic Shield).

In the Rio Maria area there can be observed the importance of rifting mechanisms and the evolution of the volcanism, the composition of which varies from komatiitic to tholeiitic and calc-alkaline which presupposes the involvement of plate tectonics from the earliest Meso-Archean times. This led to the definition of a continental microplate at *c.* 3.0 to 2.9 Ga, which was affected by the deformation and metamorphism of the Aroense Event (2.8–2.7 Ga) that gave rise to large high-angled shear zones with which are associated the gold deposits of Babaçu, Lagoa Seca and Diadema.

Although gold is found in the Meso-Archean granite greenstone terranes, large deposits formed at this time are not known. With respect to the Crixás gold deposit, there exists a difference of opinion regarding the age of the mineralization. This age is defined by the Brasiliano tectonic overriding that is younger and therefore not related to the greenstone belt rocks. However, there occur mineral deposits having a more diversified content. For example, the magnesite deposits of the Serra das Éguas in the Brumado greenstone belt, the barite deposit at Itabura in the Mundo Novo greenstone belt, the Fe-Ti-V+PGE of the Jacaré and Campo Alegre de Lourdes sills, the chromite of the Pium-hi greenstone belt, and the O'Toole Deposit (Ni-Cu-Co+PGM) of the Morro de Ferro greenstone belt. In the Gavião Block the base metals anomalies are numerous.

In the **Neo-Archean**, between 2.80 and 2.50 Ga, there developed on the South American Platform two distinct nuclei showing distinct metalogenic features that are specific to each:

- *The Carajás Mineral Province*. This is a polymetallic mineral province with deposits of iron, copper, copper-gold, manganese, chrome and nickel, displaying a complex geotectonic evolution that is still not well understood. This geotectonic evolution involved specific metal deposits generated during distinct metallogenic epochs: a) About 2.76 Ga, an epoch during which iron was precipitated as jaspilite of the Carajás-type associated with the Grão-Pará volcano-sedimentary sequence can be distinguished. It also includes the Luanga mafic-ultramafic complex that hosts deposits of chrome + PGE associated. At the same time there occurs Fe oxide Cu-Au (-U-REE) mineralization related to the Igarapé Bahia-Alemão, Pojuca, and Salobo volcano-sedimentary sequences, and to the granitic intrusions of Sossego, Cristalino, and S-118 amongst others. Following a first phase of deformation, originating from the reactivation along large shear zones, there are recognized: b) An epoch of manganese deposition in the Águas Claras sedimentary sequences represented by the Azul Mn deposit; c) An epoch marked by the Serra Pelada-Serra Leste gold deposit, associated with fractures related to further reactivation of the shear zones, and mafic intrusions at 2.645 Ga. This epoch terminates with Ni and perhaps PGE mineralization related to the differentiated mafic-ultramafic complexes: Vermelho, Onça, Puma, Jacaré and Jacarezinho, indicating the stabilization of the Amazonian Craton at the end of the Archean.

- *The Quadrilátero Ferrífero Mineral Province*: The gold (Cuiabá, Morro Velho, Raposos, Lamengo, São Bento, Juca Vieira, etc.) and manganese (Lafaiete) deposits are directly related to the evolution of the Rio das Velhas greenstone belt (2.77 Ga) and to its association with BIF units of the Algoma-type. Whereas the volcano-sedimentary origin of the manganese in the form of queluzite is widely accepted, there exists a considerable doubt regarding the early volcanogenic sulphide-associated gold mineralization present in the banded iron formation units. The large gold

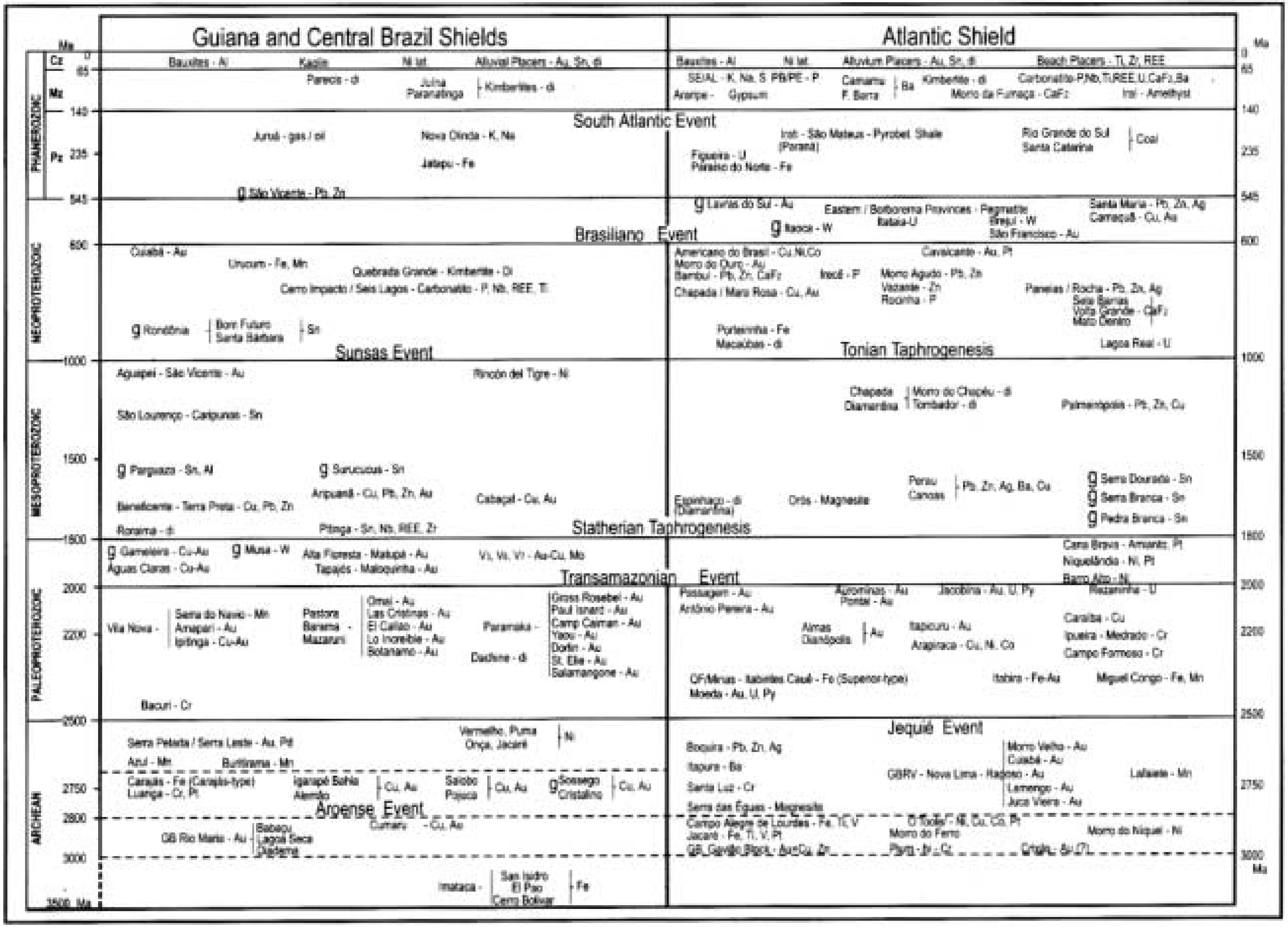


FIGURE 47 - Metallogenetic Epochs in the South American Platform.

deposits are related to low-angled shear zones that developed during Archean tectonic cycles at about 2.6 Ga, with important remobilization during the Transamazonian Event between 2.0 and 1.8 Ga. These observations lead to the definition of a gold province and an epoch of gold mineralization of fundamental importance to the evolution of the South American Platform.

In the **PALEOPROTEROZOIC** (2.5-1.8 Ga) the metallogenesis was well diversified with well-developed metallogenic epochs in the Amazonian and São Francisco cratons.

On the Guiana Shield, a gold epoch (2.2 to 2.0 Ga) is related to volcano-sedimentary sequences of the greenstone belt-type known as Pastora, Barama-Mazaruni, Paramaka and Vila Nova in which the gold mineralization is intimately associated with Transamazonian shearing (2.0 Ga). In the region of Dachine, new occurrences of diamond have been recently discovered in association with pyroclastic rocks showing a komatiitic nature. At the end of the Paleoproterozoic occured the anorogenic granite intrusions of Pitinga (1.8 Ga) with Sn-Ta-Nb-REE mineralization.

On the Central Brazil Shield a new epoch of gold mineralization between 1.9 and 1.8 Ga is developed with the definition of the Tapajós and Alta Floresta gold provinces, where the gold mineralization is associated with granitic calc-alkaline type I intrusions, and is classified as being of the porphyry-Au and epithermal Au types. At this time there can be correlated the first manifestations of anorogenic granite intrusion (1.88 Ga) with Sn-W mineralization (Musa) Granite and the Cu-Au deposits of Águas Claras (Carajás Granite) and Gameleira (Pojuca Granite).

On the Atlantic Shield the epoch of gold mineralization (2.1-2.0 Ga) is equivalent to that observed on the Guiana Shield. It is related to the volcano-sedimentary sequence of the Rio Itapicuru greenstone belt (2.2-2.1 Ga) with the gold deposits at Fazenda Brasileiro and Maria Preta, associated with shear zones formed during the Transamazonian Event. In this epoch are also included the Passagem de Mariana, Antônio Pereira gold deposits, amongst others, of the Quadrilátero Ferrífero. This period is also marked by the presence of paleoplacers of the Witwatersrand-type, assigned to the Moeda Formation of the Quadrilátero Ferrífero dated at 2.5 Ga, and the Jacobina Group in the State of Bahia dated at about 2.0 Ga; and above all the huge iron ore deposits of the Superior Province-type in the form of itabirite, occurring in the Quadrilátero Ferrífero in the State of Minas Gerais. At this time there also occurred the emplacement of the differentiated mafic-ultramafic sills with copper mineralization (Caraiba) and chrome (Rio Jacurici and Campo Formoso). Also included are the mafic-ultramafic complexes of Goiás, which mark a proto-rift system, striking approximately N-S and with which are associated the deposits of nickel (Niquelândia and Barro Alto), asbestos (Cana Brava) as well as the considerable possibility of PGE deposits (Niquelândia and Cana Brava).

In the **MESOPROTEROZOIC** (1.8-1.0 Ga) the development of intracontinental rifts has affected the stable cratonic nuclei. These rifts mark large areas of crustal weakness dominated by taphrogenesis that have as their main characteristic an association with extensive continental volcanism, anorogenic granite intrusion, and clastic sedimentary cover. The anorogenic granite intrusives are associated with tin mineralization and define a metallogenic epoch common to the Amazonian and São Francisco cratons dated at about 1.8-1.75 Ga. On the Amazonian Craton the intrusion of the anorogenic granite and associated tin mineralization migrated in time from NE to SW, together with the continental volcanism and the sedimentary cover. The principal phases of tin granite intrusion are:

- 1.88 Ga Carajás-Musa-type Granite
- 1.8 Ga Pitinga-type Granite
- 1.5 Ga Surucucus-type Granite
- 1.3 Ga São Lourenço-Caripunas Granite

-950 Ma Rondônia-type Granite (YRG) such as the deposits at Bom Futuro and Santa Bárbara.

In the State of Goiás the tin anorogenic granites occur in the Paranã (1.75 Ga) and Tocantins (1.59) sub-provinces.

To the Mesoproterozoic are also related the diamondiferous conglomerate assigned to the Roraima Group and the Espinhaço Supergroup, between 1.7 and 1.8 Ga.

Rarely, these rifts evolve to the point where oceanic crust started to develop. Exceptions are the Alto Jauru volcano-sedimentary sequence (1.75 Ga) with the Cu-Au deposit at Cabaçal (Alto Jauru District), in Mato Grosso, and the Palmeirópolis-Juscelândia volcano-sedimentary sequence (1.3 Ga) in Goiás with its associated Pb-Zn deposits.

In the Ribeira Belt a metallogenic event at *c.* 1.7 Ga, is related to stratiform deposits of Cu-Pb-Zn-Ba-Ag of the Perau-type, of sedimentary-exhalative origin. Also at *c.* 1.7 Ga it is possible to define an epoch of magnesite precipitation, of probable evaporitic origin, in the Orós Belt of the Borborema Province (Northeastern Brazil).

At the end of the Mesoproterozoic the reactivation of the Aguapeí Rift is the result of the Sunsas orogenic event at 1.0 Ga that led to the formation of a number of small gold deposits related to high-angled shear zones. These deposits define the Alto Guaporé Gold Province. To the end of the Mesoproterozoic are also related the diamondiferous conglomerate beds of the Tombador Formation and Morro de Chapéu of the Chapada Diamantina, probably deposited between 1.2 and 1.1 Ga.

In the **NEOPROTEROZOIC** there occurred the evolution between 900 and 550 Ma of fold belts and sedimentary cover that lie around the margin of the São Francisco Craton, leading to the development of mineral deposits of very variable type, reflecting the characteristics of each of the different belts.

In the **Brasília Belt** the most important deposits include a) Au and Cu-Au deposits associated with the Goiás Magmatic Arc that developed between 950 and 600 Ma; b) the Morro do Ouro gold deposit, the origin of which is attributed to tectonic overriding resulting from the Brasiliano Event (600 Ma); c) deposits of Pb-Zn-CaF_2 of the MVT-type, and the phosphate deposits at Irecê, associated with the Bambuí cover. In the external zone of the Brasília Belt, the Morro Agudo and Vazante deposits define a Neoproterozoic Pb-Zn metallogenic epoch; d) the post-tectonic intrusions (610 Ma) of the differentiated mafic-ultramafic complexes of Americano do Brasil and Mangabal with Cu-Ni-Co mineralization.

In the **Araçuaí Belt** there occur: a) iron ore deposits of the Rapitan-type in the external zone in the region of Porteirinha, representing an epoch of iron precipitation at about 900 Ma. These deposits are probably of the sedimentary-exhalative (SEDEX) type; b) graphite deposits at Pedra Azul and Salto da Divisa in the internal belt. These deposits are associated with the amphibolite and granulite facies of the metasedimentary sequences; c) the Eastern Pegmatite Province (Li, Be and gemstones), related to granite intrusions dated at *c.* 550 Ma.

In the **Ribeira Belt** the principal epochs of Neoproterozoic metallogenesis are related to a) strata-bound deposits of the Pb-Zn-Ag Panelas-type associated with limestone and dolomite beds of the Águas Claras Formation; b) granite intrusions with deposits of wolframite and gold.

In the **Dom Feliciano Belt**, the mineral deposits occur associated with a) gold-bearing porphyritic granite of the Lavras do Sul-type dated at about 570 Ma; b) the molasse sequence at Camaquã hosting deposits of Cu-Pb-Zn.

In the external zone of the **Paraguai Belt**, the graben in the Corumbá region, was filled by about 650 Ma by jaspilite intercalated with beds of manganese of sedimentary-exhalative origin. Thus the Urucum-Mutún deposits of the Rapitan-type define the last Fe-Mn epoch at the end of the Mesoproterozoic. In the internal zone of the Paraguai Belt, the gold deposits associated with phyllite of the Cuiabá Group permit the definition of a new gold province (Cuiabá-Poconé), which developed at the end of the Brasiliano Cycle.

In the **Borborema Province** the Seridó Belt contains: a) Tungsten in the form of scheelite in skarnite; b) gold associated with shear zones and; c) Pegmatite (Ta, Nb, Be, Sn) related to the Brasiliano magmatism.

On the Amazonas Craton a diamandiferous epoch at about 710 Ma can be defined with the discovery of the Quebrada Grande kimberlite, in Venezuela.

In the **PHANEROZOIC**, the South American Platform was completely stable, which permitted the development of the broad intracratonic Paleozoic syneclises of the Paraná, Parnaíba, Amazonas and other basins, at the beginning of Siluro-Ordovincian. There was an epoch of oolitic ironstone deposition in the Devonian that occurred in the three basins; and an epoch of evaporite precipitation in the Amazonas Basin (*sensu lato*) during the Permo-Carboniferous, with the formation of extensive potassium deposits. Of the diamond occurrences associated with Paleozoic sediments, the most significant are those associated with the Devonian Furnas Formation near the town of Tibagi in the State of Paraná, and the Permo-Carboniferous Aquidauana Formation near the town of Coxim on the divide between the states of Mato Grosso do Sul and Mato Grosso. Both these occurrences are within the ambit of the Paraná Basin.

The break-up of the Gondwana Supercontinent by rifting leading to the opening of the South Atlantic Ocean during the **Mesozoic** resulted in successive reactivation of the South American Platform. This in turn, led to the formation of important mineral deposits that define the South Atlantic Metallogenic Epoch. In the Lower Cretaceous a phase of extensive basaltic volcanism in the Paraná Basin is associated with important agate and amethyst deposits in southern Brazil and in Uruguay. To this epoch is related the vein fluorite deposits of Santa Catarina and the first alkaline-carbonatite complexes of Anitápolis and Jacupiranga, with apatite deposits in the southern and southeastern region of Brazil, as well as the diamondiferous kimberlite pipes of Paranatinga and Juína. On the Brazilian coast, the opening of the South Atlantic Ocean, led to the development of a gulf that provided the depositional conditions for the precipitation of Aptian evaporite beds and potassium deposits, defining thus, an evaporite epoch. Between 80 and 90 Ma, the reactivation of the rift gave rise to a second epoch of alkaline-carbonatite complex intrusion along with deposits of apatite, niobium, titanium, nickel, barite, uranium, fluorite and REE, in addition to the diamondiferous kimberlites of Alto Paranaiba. Barite was formed in coastal basins at Camamu and Fazenda Barra, and phosphorite was deposited in the Paraiba-Pernambuco Basin between the cities of Recife and João Pessoa.

Finally, the mineral deposits that originated during the **Cenozoic** are related to lateritic weathering on the South American Platform that resulted in the formation of important deposits of bauxite, kaolin, nickel, in addition to iron, gold, titanium, manganese and niobium from the lower Tertiary to Recent times. Concomitantly, placer deposits of cassiterite, gold and diamonds resulted from the mechanical concentration of heavy minerals in drainage. In like manner, placer deposits of ilmenite, rutile, zirconite and monazite have formed in beach deposits along the Brazilian coast.

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REFERENCES

Abreu, A.S., Diniz, H.B., Prado, M.G.B. and Santos, S.P. (1988). A mina de ouro de São Bento, Santa Bárbara, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S. DNPM Brasília, vol. 3, Metais básicos não ferrosos, ouro e alumínio, 393-411.

Abreu, F.A.M., Gorayeb, P.S.S. and Hasui, Y. (1994). Tectônica e inversão metamórfica no cinturão Araguaia. *Simpósio Geol. Amazônia, 4, Belém, SBG, Anais*, p. 1-4.

Abreu, P.A.A., Knauer, L.G. and Renger, F.E. (1997). A rocha-matriz dos diamantes da Formação Sopa-Brumadinho da Província de Sopa-Guinda, Serra do Espinhaço Meridional. In: *Simpósio Brasileiro de Geologia do Diamante, 2, Cuiabá, Programa, Resumos, Palestra e Roteiro de Excursão, UFMT, Publicação Especial 03/97*, 13-14.

Aleva, G.J.J. (1981). Essential Difference Between the Bauxite Deposits Along the Southern and Northern Edges of the Guiana Shield, South America. *Economic Geology*, **76**, 1142-1152.

Alkmin, F.F., Brito Neves, B.B. and Alves, J.A.C. (1993). Arcabouço tectônico do Cráton do São Francisco, uma revisão. In: *Simpósio Sobre o Cráton do São Francisco*, coords. Dominguez, J.M.L. and Misi, A. Salvador, SBG/SGM/CNPq, 45-62.

Almeida, F.F.M. (1945). Geologia do sudoeste matogrossense. *Bol. DGM/DNPM*, Rio de Janeiro, **116**, 181 p.

Almeida, F.F.M. (1964). Geologia do centro-oeste matogrossense. *Bol. DNPM/DGM*, Rio de Janeiro, **214**, 137 p.

Almeida, F.F.M. (1967). Origem e evolução da Plataforma Brasileira. *Bol. DNPM/DGM*, Rio de Janeiro, **241**, 36 p.

Almeida, F.F.M. (1977). O Cráton do São Francisco. *Revista Brasileira de Geociências*, **7**, 349-364.

Almeida, F.F.M., Hasui Y. and Fuck, R.A. (1981). Brazilian structural provinces: An introduction. *Earth Sciences Review*, **17**, 1-29.

Almeida F.F.M., Hasui, Y. and Brito Neves, B.B. (1976). The Upper Precambrian of South America. *Boletim IG-USP*, **7**, 45-80.

Almeida, F.F.M. (1978). *Tectonic map of South America 1:5.000.000*. Explanatory note, Brasília DNPM/CGMW/UNESCO, 23 p.

Althoff, A.M.R., Villas, R.M. and Giuliani, R.A. (1994). Mineralização cuprífera da área Bahia, Serra dos Carajás (PA): Evolução dos fluidos hidrotermais e modelo metalogenético. *Geochimica Brasiliensis*, **8**, 135-155.

Alvarenga, C.J.S. and Trompette, R. (1993). Evolução tectônica Brasíliana da Faixa Paraguai: a estruturação da região de Cuiabá. *Revista Brasileira de Geociências*, **23**(1), 18-30.

Alvarenga, C.J.S., Cathelineau, M. and Dubessy, J. (1990). Chronology and orientation of N_2-CH_4, CO_2-H_2O and H_2O rich fluid-inclusions traies in intrametamorphic quartz veins from the Cuiabá gold district Brazil. *Mineralogical Magazine*, **54**, 245-255.

Alves, C.A., Bernadelli, A.L. and Beisiegel, V.R. (1986). A jazida de níquel laterítico do Vermelho, Serra dos Carajás, Pará. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S. DNPM/CVRD, vol. II, 325-334.

Alves da Silva, F.C. and Matos, V. (1991). Economic geology and structural controls of the orebodies from the medium Itapicuru gold district: Rio Itapicuru greenstone belt, Bahia, Brazil: In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 629-635. Rotterdam, A.A. Balkema.

Alves da Silva, F.C., Chauvet, A. and Faure, M. (1998). General features of the gold deposits in the Rio Itapicuru greenstone belt (RIGB, Brazil): Discussion of the origin, timing and tectonic model. *Revista Brasileira de Geociências*, **28**(3), 377-390.

Amaral, A.J.R. and Melo, P.R.C. (1997). O depósito de sal-gema de Bebedouro, Maceió, Alagoas. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IV-C, 505-518, DNPM/CPRM, Brasília.

Amaral, G., Bushee, J., Cordani, U.G., Kawashita, K. and Reynolds, J.H. (1967). Potassium-argon ages of alkaline rocks from Southern Brazil. *Geochimica et Cosmochimica Acta*, **31**(2), 117-142.

Andrade, M.S., Strieder, A.J. and Gastal, M.C.P. (1988). Controles geotectônico, magmático e estrutural das ocorrências Minerais de Cu-Au na região de Lavras do Sul (RS). In: *Congresso Brasileiro de Geologia, 40, Belo Horizonte, Anais*, Sociedade Brasileira de Geologia, 132 p.

Angélica, R.S., Costa, M.L., Lenharo, S.L.R. and Pöllmann, H. (1996). Ocorrência de uraninita associada com o ouro de Igarapé Bahia, Carajás, Pará. *Simp. Geol. Amazônia, 5, Belém, SBG, Bol. Res. Expandidos*, 152-155.

Arantes, D., Buck, P.S., Osborne, G.A. and Porto, C.G. (1981). A sequência vulcano-sedimentar de Mara Rosa e mineralizações associadas. *Boletim do Centro-Oeste*, SBG, **14**, 27-40.

Arantes, D., Osborne, G.A., Buck, P.S. and Porto, C.G. (1991). The Mara Rosa volcano-sedimentary sequence and associated gold mineralization. In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 221-229. Rotterdam, A.A. Balkema.

Araújo Filho, J.O. (1999). *Structural characteristics and tectonic evolution of the Pirineus sintaxis, Central Brazil*. Ph.D. Thesis, University of Illinois, 418 p. (unpublished).

Araújo, S.M. (1999). The Palmeirópolis volcanogenic massive sulfide deposit, Tocantins State. In: *Base Metal Deposits of Brazil*, coords., Silva, M.G. and Misi, A. pp. 64-68. MME/CPRM/DNPM, Belo Horizonte.

Araújo, S.M. and Nilson, A.A. (1988). Depósito de zinco, cobre e chumbo de Palmeirópolis, Goiás. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 171-180. MME-DNPM-CVRD.

Archanjo, C.J. and Salim, J. (1986). Posição da Formação Seridó no contexto estratigráfico regional (RN-PB). *Simp. Geol. Nordeste, 12, João Pessoa, SBG, Anais*, 270-281.

Badi, W.S.R. and Gonzalez, A.P. (1988). Jazida de metais básicos de Santa Maria, Caçapava do Sul, RS. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S. DNPM, vol. III, 157-170.

Baeta Jr., J.D.A. (1986). As jazidas de níquel laterítico de Barro Alto, Goiás. In: *Principais Depósitos Minerais do Brasil, coords.* Schobbenhaus, C. and Coelho, C.E.S. vol. II, 315-323. DNPM/CVRD.

Bandeira, S.A.B., Morelli, S., Mello, C.S.B. and Moraes, R.A.V. (1986). Depósito Stratabound de barita da Fazenda Barra, bacia sedimentar de recôncavo/Tucano (BA). In: *Congresso Brasileiro de Geologia, 34, Goiânia, 1986. Anais*... Goiânia, SBG, **5**, 2229-2240.

Baquil, C.C. (1997). Depósitos de Gipsita do Grajaú, Maranhão. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVC, 165-175.

Barbosa de Deus, P. and Viana, J.S. (1982). Distrito cromitífero do Vale do Rio Jacurici. *Congr. Bras. Geol., 32, Salvador, SBG, Roteiro das Excursões*, **3**, 44-52.

Barbosa de Deus, P., Viana, J.S., Duarte, P.M. and Queiróz, J.A. (1982). Distrito cromitífero de Campo Formoso. *Congr. Bras. Geol., 32, Salvador, SBG, Roteiro das Excursões*, **3**, 52-59.

Barbosa, J.S.F. (1997). Síntese do conhecimento sobre a evolução geotectônica das rochas metamórficas arqueanas e paleoproterozóicas do embasamento do Cráton do São Francisco, na Bahia. *Revista Brasileira de Geociências*, 27(3), 241-256.

Barbosa, O., Andrade Ramos, J.R., Gomes, F.A. and Helmbold R. (1966). *Geologia estratigráfica, estrutural e econômica do Projeto Araguaia*. Rio de Janeiro. DNPM/DGM, 94 p.

Barbosa, O., Braun, O.P.G., Dyer, R.C.E. and Cunha, C.A.B.R. (1970). Geologia da região do Triângulo Mineiro. (Projeto Chaminés). *Bol. DNPM/DFPM*, Rio de Janeiro, **136**, 140 p.

Barcelos, J.P. and Büchi, J. (1986). Mina de minério de ferro de Alegria, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. II, 77-85. DNPM, Brasília.

Barcelos, J.P. and Büchi, J. (1986). Mina de minério de ferro-manganês de Miguel Congo, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. II, 87-95. DNPM, Brasília.

Barreira, C.F., Soares, A.D.V. and Ronzê, P.C. (1999). Descoberta do depósito Cu-Au Alemão, Província Mineral de Carajás. *Simp. Geol. Amazônia, 6, Manaus, SBG, Anais*, 136-140.

Bardoux, M., Voicu, G. and Lafrance, J. (1998). Geological settings of base and precious metal occurrences in Paleoproterozoic terranes of the Guyana Shield. *Mineralium Deposita* (submitted).

Bastos, J.B.S. (1988). Depósitos de ouro do Rio Madeira, Rondônia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S. DNPM, Brasília, **3**, 575-580.

Bastos Neto, A.C., Touray, J.C., Dardenne, M.A. and Charvet, J. (1992). Contrôle tectonique des minéralisations à fluorine de Santa Catarina, Brésil: filons en décrochement et en extension. *Chronique de la Recherche Miniére*, **507**, 43-52.

Bastos Neto, A.C. (1990). *Le district à fluorine de Santa Catarina: minéralisations et altérations hydrothermales dans leur cadre géodynamique*. Doctorate Thesis, Université d'Orléans, França, 420 p.

Becker, F.E., Valle, R.R. and Coelho, C.E.S. (1997). Depósito de fluorita de Tanguá, Itabraí, RJ. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IV-B, 591-611. DNPM/CPRM.

Beisiegel, V.R., Bernadelli, A.L., Drumond, N.F., Ruff, A.W. and Tremaine, J.W. (1973). Geomorfologia/Geologia e recursos minerais da Serra dos Carajás. *Revista Brasileira de Geociências*, **3**, 215-242.

Beissner, H., Carvalho, A., Lopes, M. and Valeton, I. (1997). The Cataguazes bauxite deposit. In: *Brazilian Bauxites*, eds. Carvalho, A., Boulangé, B., Melfi, A.J. and Lucas, Y. pp. 195-208. USP/FAPESP/ORSTOM, São Paulo.

Bellizzia, A., Pimentel de Bellizzia, N. and Rodrigues, S. (1981). Recursos minerales de Venezuela y su relacion a la metalogenesis. In: Minerales de Venezuela. Ministério de Energia y Minas, Dirección de Geologia, Caracas, *Bol. de Geologia, Publicação Especial*, **8**, 6-77.

Bernadelli, A., Melfi, A.J., Oliveira, S.M.B. and Trescases, J.J. (1983). The Carajás nickel deposits. In: *Lateritisation Processes, IGCP-IUGS-UNESCO*, eds. Melfi, A.J. and Carvalho, A., pp. 108-118. Proj. 129, IAGCII, Intern. Sem. on Lateritisation Processes, São Paulo.

Bernadelli, A.L. (1982). Jazida de manganês do Azul. *Simp. Geol. Amazônia, 1, Belém, SBG, Anexo aos Anais*, 47-60.

Bernadelli, A.L. and Beisiegel, V.R. (1978). Geologia econômica da jazida de manganês do Azul. *Congr. Bras. Geol., 30, Recife, SBG, Anais*, **4**, 1431-1444.

Bertoni, C.H., Bardoux, M. and O'Donnel, M. (1998). Escudo das Guianas: o contexto geológico dos depósitos de ouro. In: *Congresso Brasileiro de Geologia, 40, Belo Horizonte, Anais, Sociedade Brasileira de Geologia*, 135 p.

Bettencourt, J.S. (1976). Minéralogie, inclusions fluides et isotopes stables d'oxygéne et de soufre de la mine de cuivre de Camaquã, RS (une étude préliminaire). *Congr. Bras. Geol., 29, Ouro Preto, SBG, Anais*, **2**, 409-423.

Bettencourt, J.S., Leite Jr., W.B., Payolla, B.L., Scandolara, J.E., Muzzolon R. and Vian J.A.J. (1997). The Rapackivi granites of the Rondônia Tin Province, Northern Brazil. *ISGAM II, Salvador, SBG, Excursion Guide*, 3-31.

Bizzi, L.A. (1993). *Mesozoic alkaline volcanism and mantle evolution of the southwestern São Francisco Craton, Brazil*. Ph.D. Thesis. University of Cape Town, 240 p.

Bodenlos, A.J. (1960). Magnesite deposits in the Serra das Éguas, Brumado, Bahia, Brazil. *U.S. Geological Survey Bulletin*, 975-C, Washington DC, 87-167.

Bonfim, L.F.C. (1986). Fosfato de Irecê (BA): um exemplo de mineralização associada a estromatólitos do Precambriano Superior. *Anais 34° Cong. Bras. Geol., Goiânia, SBG*, 5, 2154-2167.

Borges, O.C. (1999). *Processo MME/DNPM 851.676/92*. Relatório Final de Pesquisa, Mineração Itajobi Ltda.

Bosma, W., Kroonenberg, S.B., Maas, K. and De Roever, E.W.F. (1983). Igneous and metamorphic complexes of the Guiana Shield in Suriname. *Geologie en Mijnbouw*, **62**, 241-254.

Botelho, N.F. (1992). *Les ensembles granitiques subalcalins à peralumineux minéralisés Cu ,Sn et In de la sous-province Paranã, état de Goiás, Brésil*. Doctorate Thesis, Univ. Paris VI, 343 p. (unpublished).

Botelho, N.F. and Moura, M.A. (1998). Granite-ore deposit relationships in Central Brazil. *Journal of South American Earth Sciences*, **11**(5), 427-438.

Botelho, N.F., Moura, M.A., Souza, M.T. and Antunes, J.A. (1997). Petrologia e potencial metalogenético de granitos da região de Peixoto de Azevedo-Alta Floresta, Mato Grosso. In: *Simpósio de Geologia do Centro-Oeste, 6, Cuiabá, SBG, Anais*, p. 40-42.

Boujo, A., Menor, E.A., Lima, F.V., Amaral, A.J. and Magat, P. (1998). Contrôles géologiques et structuraux de la minéralisation phosphatée du nordeste brésilien. Conséquences sur le mode d'altération et la concentration résiduelle du phosphate: Exemple du gisement de Congaçari (PE). *Anais da Academia Brasileira de Ciências*, 627-646.

Boulangé, B. and Carvalho, A. (1997). The bauxite of Porto Trombetas. In: *Brazilian Bauxites*, coords. Carvalho, A., Boulangé, B., Melfi, A.J. and Lucas, Y. pp. 55-73, USP/FAPESP/ORSTOM, São Paulo.

Bowring, S.A., Grotzinger, J.P., Isachsen, C.E., Knoll, A.H., Pelechaty, S.M. and Kolosov, P. (1993). Calibrating Rates of Early Cambrian Evolution. *Science*, **261**, 1293-1298.

Brenner, T.L., Teixeira, N.A., Oliveira, J.A.L., Frank, N.D. and Thompson, J.F.H. (1990). The O'Toole nickel deposit, Morro do Ferro greenstone belt, Brazil. *Economic Geology*, **85**, 904-920.

Brito Neves, B.B., Campos Neto, M.C. and Fuck, R.A. (1999). From Rodinia to western Gondwana: an approach to the Brasíliano-Pan African Cycle and orogenic collage. *Episodes*, **22**(3), 155-166.

Brito, R.S.C. (1984). Geologia do sill estratificado do Rio Jacaré, Maracás, Bahia. *Congr. Bras. Geol., 33, Rio de Janeiro, SBG, Anais*, **9**, 316-331.

Brito, R.S.C., Pimentel, M.M., Nilson, A.A. and Gioia, S.M. (1999). Samarium-neodymium and rubidium-strontium isotopic systematics of the Rio Jacaré sill, Bahia, Brazil. *Simp. South American on Isotope Geology, 2, Cordoba, Argentina, Actas*, 44-47.

Brito W., Raposo, C. and Matos, E.C. (1984). Os albititos uraníferos de Lagoa Real. *Congr. Bras. Geol., 33, Rio de Janeiro, SBG, Anais*, 1475-1488.

Brooks, W.E., Tosdal, R.M. and Nuñez, F.J. (1995). Gold and diamond resources of the Icabarú Sur Study Area, Estado Bolívar, Venezuela. In: *Geology and Mineral Deposits of the Venezuelan Guayana*, eds. Sidder, G.B., Garcia, A.E. and Stoeser, J.W., pp. L1-L2, Washington.

Buchanan, L.J. (1981). Precious metal deposits associated with volcanic environments in the southwest. *Arizona Geolical Society Digest*, **14**, 237-261.

Cainelli, C. and Mohriak, W. (1999). Some remarks on the evolution of sedimentary basins along the eastern brazilian continental margin. *Episodes*, **22**(3), 206-216.

Campos, J.E.G. (1996). *Estratigrafia, sedimentação, evolução tectônica e geologia do diamante da porção centro-norte da Bacia Sanfranciscana*. Doctorate Thesis, UnB, Brasília, 204 p.

Campos, J.E.G., Dardenne, M.A., and Gonzaga, G.M. (1993). O potencial diamantífero do conglomerado Abaeté no NW de Minas Gerais. *Simp. Geol. Diamante, 1, Cuiabá, SBG, Anais*, 101-113.

Capdevila, R., Arndt, N., Letendre, J. and Sauvage, J.-F. (1999). Diamonds in volcaniclastic komatiite from French Guiana. *Nature*, **399**(3), 456-458.

Carvalho, M.T.M. (1999). *Integração de dados geológicos, geofísicos e* *geoquímicos aplicada à prospecção de ouro nos greenstones belts de Pilar de Goiás e Guarinos, GO.* M.Sc. Thesis, UnB, Brasília, 190 p.

Carvalho, I.G., Iyer, S., Tassinari, C.C.G. and Misi, A. (1997). Lead-and-sulfur-isotope investigations of the Boquira sediment-hosted sulfide deposit, Brazil. *International Geology Review*, 39, 97-106.

Carvalho, J.M.A., Faraco, M.T.L. and Klein, E.L. (1995). *Carta Geoquímico-metalogenética do Ouro do Amapá/NW do Pará, 1:500.000.* Serviço Geológico do Brasil-CPRM.

Carvalho, M.S. and Figueiredo, A.J.A. (1982). Caracterização litoestratigráfica da bacia de sedimentação do Grupo Beneficente no Alto Rio Sucunduri, AM. In: *Simpósio de Geologia da Amazônia, 1, Belém, Anais*, Sociedade Brasileira de Geologia, 26-44.

Cassedanne, J.P. (1991). Tipologia das jazidas brasileiras de gemas. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IV-A, 17-52. DNPM, Brasília.

Castro, A.B., Neto, J.L.M., Souza, L.R.F. and Lima, M.A.T. (1997). Depósito de barita de Altamira, Itapura, Miguel Calmon, Bahia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S. vol. IV-B, 237-248. DNPM, Brasília.

Castro, E.C., Ferreira, L.A.D. and Akinaga, R.M. (1974). Ametista do Brasil. *Cong. Bras. Geol., 8, Porto Alegre, SBG, Anais*, 1, 9-27.

Castro Filho, L.W. and Mattos, S.C. (1986). Depósitos de níquel laterítico de Jacaré e Jacarezinho, município de São Félix do Xingu, Pará. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C. and Coelho C.E.S., vol. II, 369-384. DNPM/CVRD.

CBMM - Companhia Brasileira de Metalurgia e Mineração (1984). *Complexos carbonatíticos do Brasil: geologia.* São Paulo, CBMM, 44 p.

Cerqueira, R.M., Pereira, J.C. and Pessoa, A.F.C. (1986). Jazida de potássio de Santa Rosa de Lima, SE. Geologia e Avaliação de Reservas. *An. 34º Cong. Bras. Geol., Goiânia, SBG*, 5, 2168-2181.

Cerqueira, R.M., Chaves, A.P.V., Pessoa, A.F.C., Monteiro, J.L.A., Pereira, J.C. and Wanderley, M.L. (1997). Jazidas de potássio de Taquari/Vassouras, Sergipe. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S. DNPM/CPRM, vol. IVC, p. 277-312.

Channer, D.M. DeR., Cooper, R.E.C. and Kaminsky, F. (1998). The Guaniamo diamond region, Bolivar state, Venezuela: a new kimberlite province. In: *International Kimberlite Conference, 7, Cape Town, Extended Abstracts*, 144-146.

Chauvet, A. and Menezes, M.G. (1992). Evolution structurale du sud du Craton de São Francisco: Implications sur les minéralisations auriféres de la région d'Ouro Preto, Brésil. *Comptes Rendus de l'Académie des Sciences, Paris*, 315, 495-501.

Chauvet, A., Dussin, I.A., Faure, M. and Charvet, J. (1984). Mineralização aurífera de idade proterozóica superior e evolução estrutural do Quadrilátero Ferrífero, Minas Gerais, Brasil. *Revista Brasileira de Geociências*, 24(3), 150-159.

Chaves, A.G., Heineck, C.A. and Tavares, W.P. (1976). *Projeto Patos de Minas.* Belo Horizonte. CPRM, vol. 2, 78 p. (Relatório Final de Pesquisa - unpublished).

Chaves, M.L.S.C., Dupont, H., Karfunkel J. and Swisero, D.P. (1993). Depósitos diamantíferos de Minas Gerais: uma revisão. *Simp. Geol. Diamante, 1, Cuiabá, SBG, Anais*, 79-100.

Chaves, M.L.S. and Uhlein, A. (1991). Depósitos diamantíferos da região do alto/médio Rio Jequitinhonha, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IV-A, 117-138. DNPM/CPRM, Brasília

Coelho, C.E.S. (1986). Depósitos de ferro da Serra dos Carajás, Pará. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. II, 29-64. DNPM, Brasília.

Coelho, C.E.S. and Freitas-Silva, F.H. (1998). The structural control of gold deposits of the Fazenda Maria Preta gold district at Rio Itapicuru greenstone belt, Northeastern Brazil. *Revista Brasileira de Geociências*, 28(3), 367-376.

Coelho, C.E.S. and Rodrigues, O.B. (1986). Jazida de Manganês do Azul, Serra dos Carajás, Pará. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S., vol. II, 145-152. DNPM, Brasília.

Cordani, U.G. and Brito Neves, B. (1982). The geologic evolution of South America during the Archean and Early Proterozoic. *Revista Brasileira de Geologia*, 12(1-3), 77-88.

Cordani, U.G. and Sato, K. (1999). Crustal evolution of the South American Platform, based on Nd isotopic systematics on granitoids rocks. *Episodes*, 22(3), 167-173.

Cordani, U.G., Iyer, S.S., Taylor, P.N., Kawashita, K. and Sato, K. (1992). Pb-Pb, Rb-Sr and K-Sr systematics of the Lagoa Real uranium province (South-Central Bahia, Brazil) and the Espinhaço Cycle (Ca 1,5-1,0Ga). *Journal of South American Earth Sciences*, 5, 33-46.

Cordani, U.G. and Hasui, Y. (1968). Idades K-Ar de rochas alcalinas de Primeiro Planalto do Estado do Paraná. In: *Congresso Brasileiro de Geologia, 22, Belo Horizonte, SBG, Anais*, vol. 1, 57-58.

Cordani, U.G., Teixeira, W., Tassinari, C.C.G., Kawashita, K. and Sato, K. (1988). The growth of the Brazilian Shield. *Episodes*, 11(3), 163-167.

Cordeiro, A.A.C. and Silva, A.V. (1986). Depósito de wolframita da região de Pedra Preta, Pará. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus C. and Coelho C.E.S., vol. II, 409-415. DNPM, Brasília.

Coronel, N., Spoturno, J., Gómez, C., Heinzen, W., Mari, C., Roth, W., Theune, C. and Stampe, W. (1987). *Memoria de la Carta de Materias Primas Minerales No Metalicas - a Escala 1/1.000.000.* Ministerio de Industria y Energia, Direccion Nacional de Mineria y Geologia, Montevideo, 119 p.

Correia, C.T., Girardi, V.A.V., Lambert, D.D., Kinny, P.D. and Reeves, S.J. (1996). 2 Ga U-Pb (SHRIMP II) and Re-Os ages for the Niquelândia basic-ultrabasic layered intrusion, Central Goiás, Brazil. *An. 39º Cong. Bras. Geol., Salvador, SBG*, vol. 6, 187-189.

Correia, C.T., Girardi, V.A.V., Tassinari, C.C.G. and Jost, H. (1997). Rb-Sr and Sm-Nd geochronology of the Cana Brava layered mafic-ultramafic intrusion, Brazil, and considerations regarding its tectonic evolution. *Revista Brasileira de Geociências*, 27, 163-168.

Correia, C.T., Jost, H., Tassinari, C.C.G., Girardi, V.A.V. and Kinny, P.D. (1999). Ectasian Mesoproterozoic U-Pb ages (SHRIMP II) for the metavolcano-sedimentary sequences of Juscelândia and Indaianópolis and for high grade metamorphosed rocks of Barro Alto stratiform igneous complex, Goiás State, Central Brazil. *Actas 2º South American Symposium on Isotope Geology, Cordoba, Argentina*, 31-33.

Correia, C.T., Tassinari, C.C.G., Lambert, D.D., Kinny, P. and Girardi, V.A.V. (1997). U-Pb (SHRIMP), Sm-Nd and Re-Os systematics of the Cana Brava, Niquelândia and Barro Alto layered intrusions in Central Brazil, and constraints on the tectonic evolution. *South American Symp. on Isotope Geology, Campo do Jordão, SP, Ext. Abst.*, 88-89.

Correia-Neves, J.M. (1997). Província pegmatítica oriental do Brasil. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C., Queiroz E.T. and Coelho C.E.S., vol. IV-B, 343-362. DNPM, Brasília.

Correia-Neves, J.M., Pedrosa-Soares, A.C. and Marciano, V.R.P.O. (1986). A Província pegmatítica oriental à luz dos conhecimentos atuais. *Revista Brasileira de Geociências*, 16(1), 106-118.

Correia-Neves, J.M., Pedrosa-Soares, A.C., Marciano, V.R.P.O., Monteiro, R.L.B.P. and Fernandes, M.L.S. (1987). Granitoids and pegmatites from the northern of the Eastern Brazilian Pegmatite Province. *Symp. on Granites and Associated Mineralizations, 1, Salvador, SBG, Excursions Guides*, 125-144.

Costa, J.B.S., Araújo, O.J.B., Santos, A., João, X.S.J., Macambira, M.J.B. and Lafon, J.M. (1995). A Província Mineral de Carajás: aspectos tectono-estruturais, estratigráficos e geocronológicos. *Boletim do Museu Paraense Emilio Goeldi*, Série Ciências da Terra, 7, 199-235.

Costa Jr., C.M., Ferreira Filho, C.F., Osborne, G.A., Araújo, S.M. and Lopes, R.O. (1997). Geology and geochemistry of the Boa Vista nickel sulfide deposit, Crixás greenstone belt, central Brazil. *Revista Brasileira de Geociências*, 27(4), 365-376.

Costa, M.L. (1997). Lateritization as a major process of ore deposit formation in the amazon region. *Exploration & Mininig Geology*, 6(1), 79-104.

Costa, M.L. and Morães, E.L. (1998). Mineralogy, geochemistry and genesis of kaolim from the Amazon region. *Mineralium Deposita*, 33(3), 283-297.

Costa, M.L., Angélica, R.S. and Fonseca, L.R. (1996). Geochemical exploration for gold in deep weathered lateritised gossans in the Amazon region, Brazil: A case history of the Igarapé Bahia deposit. *Geochimica Brasiliensis*, 10(1), 13-26.

Costa, M.L., Costa, J.A.V. and Angélica, R.S. (1993). Gold bearing bauxitic laterite in a tropical rain forest climate: Cassiporé, Amapá, Brazil. *Chronique de la Recherche Minière*, **510**, 41-51.

Costa, L.T.R. and Carvalho, J.M.A. (1999). Tipologia de mineralizações auríferas da região sul da Província Tapajós-Pará. In: *Simpósio de Geologia da Amazônia, 6. Manaus, Boletim de Resumos Expandidos, Sociedade Brasileira de Geologia*, 114-117.

Coutinho, M.G. da N., Robert, F. and Santos, R.A. (1998). Província Mineral de Tapajós, Amazônia, Brasil: Novo enfoque geológico das mineralizações de ouro. In: *Congresso Brasileiro de Geologia, 40, Belo Horizonte, SBG, Anais*, p. 160.

Cox, D.P. (1967). Regional environment of the Jacobina auriferous conglomerate, Brazil. *Economic Geology*, 62, 773-780.

Cruz, E.L.C. (1993). *Geologia e mineralizações auríferas do terreno granitóide-greenstone de Almas-Dianópolis, Tocantins*. M.Sc. Thesis, UnB, Brasília, 152 p.

Cruz, E.L.C. (1998). *Origem dos fluidos e dos solutos em sistemas hidrotermais do tipo lode-gold, Mina Córrego Paiol, Tocantins*. Exame Qualificação, UnB, Brasília, 82 p.

Cruz, F.F., Brenner, T.L., Moreira, A.F.S., Cunha, C.A.B.R., Gallo, C.B.M., Frank, N.D. and Pimentel, R.C. (1986). Jazida de Ni-Cu-Co de Fortaleza de Minas, Minas Gerais. In: *Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S., vol. II, p.275-306. DNPM, Brasília.

D'El Rey Silva, L.J.H. and Oliveira, J.G. (1999). Geology of the Caraíba coppermine and its surroundings in the Paleoproterozoic Curaçá belt, Curaçá river valley, Bahia, Brazil. In: *Base Metal Deposits of Brazil*, eds. Silva, M.G. and Misi, A., pp. 25-32. MME/CPRM/DNPM, Belo Horizonte.

D'El Rey Silva, L.J.H. and Senna Filho, V. (1998). Ouro em sericita-quartzo hidrotermalitos controlados por cisalhamentos conjugados Brasilianos, na região de Cavalcante (GO), Brasil Central. *Revista Brasileira de Geociências*, 28(3), 405-408.

Da Rocha Araujo, P.R., Flicoteaux, R., Parron C. and Trompette, R. (1992). Phosphorites of Rocinha mine - Patos de Minas (Minas Gerais, Brasil): Genesis and evolution of a Middle Proterozoic Deposit tectonized by the Brasiliano Orogeny. *Economic Geology*, 87, 332-351.

Daitx, E.C. (1998). Os depósitos de zinco e chumbo de Perau e Canoas e o potencial do vale do Ribeira. *Workshop, Depósitos Minerais Brasileiros de Metais Base, Salvador*, CAPES/PADCT/UFBA/ADIMB, 68-74.

Dall'Agnol, R., Bittencourt, J.S., João, X.S.J., Medeiros, H., Costi, H.T., Macambira, M.J.B. (1987). Granitogenesis in the northern Brazilian region: A review. *Revista Brasileira de Geociências*, 17, 382-403.

Dall'Agnol, R., Lafon, J.M. and Macambira, M.J.B. (1994). Proterozoic anorogenic magmatism in the central Amazonian Province, Amazonian Craton. Geochronological, Petrological and Geochemical aspects. *Mineralogy and Petrology*, 50, 113-138.

Daoud, W.E.K. (1988). *Granitos estaníferos de Pitinga, Amazonas: contexto geológico e depósitos minerais associados*. M.Sc. Thesis, UnB, Brasília, 194 p.

Dardenne, M.A. (1978). Geologia da região de Morro Agudo (MG). *Bol. Núcleo Centro-Oeste, SBG*, **7/8**, 68-94.

Dardenne, M.A. (1979). *Les minéralisations de plomb, zinc, fluor du Protérozoïque Supérieur dans le Brésil Central*. Doctorate Thesis, Univ. Paris VI, 251 p. (unpublished).

Dardenne, M.A. (1982). Chrono-stratigraphie et metallogenie du Précambrien dans le Brésil Central. *5° Congreso Latinoamericano de Geologia, Argentina, Actas*, III, 65-77.

Dardenne, M.A. (1988). Fluorite-vein deposits of the Santa Catarina district in Brazil. *7th Symposium International Association of Geologists of Ore Deposits* (IAGOD) 269-274. Ed. Schweizerbart'sche.

Dardenne, M.A. (1988). Geologia do chumbo e zinco. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C. and Coelho C.E.S., vol. III, p. 83-90. DNPM.

Dardenne, M.A. (1997). Geologia da barita. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C., Queiroz E.T. and Coelho, C.E.S., vol. IVB, 215-223. DNPM, Brasília.

Dardenne, M.A. (1998). Modelo hidrotermal sedimentar-exalativo para os depósitos Fe-Mn da região de Corumbá, Mato Grosso do Sul. *Cong. Bras. Geol., 40, Belo Horizonte, SBG, Anais*, p.152.

Dardenne, M.A. and Campos, E.G. (1984). Geologia e geoquímica do depósito de barita de Camamu, Bahia. In: *Congresso Brasileiro de Geologia, 33, Rio de Janeiro, 1984. Anais*... Rio de Janeiro, SBG, 1144-1161.

Dardenne, M.A. and Freitas-Silva, F.H. (1998). Modelos genéticos dos depósitos Pb-Zn nos grupos Bambuí e Vazante. *Workshop: Depósitos Minerais Brasileiros de Metais-Base*. Salvador, CAPES-PADCT-ADIMB, 86-93.

Dardenne, M.A. and Freitas-Silva, F.H. (1999). Pb-Zn ore deposits of Bambuí and Vazante groups in São Francisco Craton and Brasília Fold Belt. In: *Base Metal Deposits of Brazil*, eds. Silva M.G. and Misi A., pp.75-83. MME/CPRM/DNPM, Belo Horizonte.

Dardenne, M.A. and Savi, C.M. (1984)- Geologia e geoquímica dos filões de fluorita Segunda Linha Torrens e Cocal, SC. *Revista Brasileira Geociências*, 14, 120-127.

Dardenne, M.A. and Touray, J.C. (1988). La fluorine du Brésil. Gisements filoniens traditionnels et nouveaux types de minéralisations. *Chronique de la Recherche Minière*, **490**, 35-46.

Dardenne, M.A., Ferreira Filho, C.F. and Meirelles, M.R. (1988). The role of shoshonitic and calc-alkaline suites in the tectonic evolution of the Carajás district, Brazil. *Journal of South American Earth Sciences*, 1(4), 363-372.

Dardenne, M.A., Freitas-Silva, F.H., Nogueira, G.M.S. and Souza, J.F.C. (1997). Depósitos de fosfato de Rocinha e Lagamar, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C., Queiroz E.T. and Coelho C.E.S., vol. IVC, 113-122. DNPM/CPRM.

Dardenne, M.A., Gonzaga, G.M., Campos, J.E.G. (1991). The diamond bearing Cretaceous conglomerates of the Canabrava area, Minas Gerais, Brazil. In: *Field Guide Book*, eds. Leonardos, O.H., Meyer, H.O.A. and Gaspar, J.C., pp. 83-88. Fifth International Kimberlite Conference. CPRM, Special Publication 3/91, Brasília.

Dardenne, M.A., Ronchi, L.H., Bastos Neto, A.C. and Touray, J.C. (1997). Geologia da fluorita. Os distritos de fluorita brasileiros. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVB, 479-507. DNPM/CPRM.

Dardenne, M.A. (1999). Os recursos minerais do Cretáceo no Brasil.

Simp. Cretáceo Brasil, 5, Serra Negra, SP, UNESP, p. 243-254.

Davison, I., Teixeira, J.B.G., Silva, M.G., Neto, M.B.R. and Mato, F.M.V. (1988). The Rio Itapicuru greenstone belt, Bahia, Brazil: structure and stratigraphical outline. *Precambrian Research*, 42, 1-18.

De Witt, E., Landis, G.P., Zartman, R.E., Garayp, E., Martins Pereira, S.L., Prado, M.G.B., Vieira, F.W.R. and Thorman, C.H. (1994). Isotopic and fluid inclusion data on the age and origin of the São Bento and Morro Velho Gold deposits, Minas Gerais, Brazil. In: *United States Geol. Sur. Res. on Mineral Deposits, Parte A, Programs and Abstracts*, 9th V.E. McKelvey Forum on Energy and Mineral Resources, eds. Thorman, C.H. and Lane, P.E., circular 1103-A, 27-29.

Delgado, I.M., Pereira, A.J. and Thormann, C.H. (1994). Geology and mineral resources of Brazil: a review. *International Geology Review*, 36, 503-544.

Della Fávera, J.C., Chaves, H.A.F., Pereira, E., Bergamaschi, S., Reis, C.C., Lima Filho, M.F. and Pereira, S.D. (1993). Geologia da área de Figueira-Sapopema - Convênio UERJ/CPRM, UERJ/DGG/LABCG, 2° Curso de Análise de Bacias, Rio de Janeiro, 89 p.

Della Fávera, J.C., Chaves, H.A.F., Pereira, E., Câmara, F.O.L.M. and Medeiros, M.A.M. 1992. *Geologia da área de Candiota, Bacia do Paraná, RS.* Convênio UERJ-CPRM, UERJ/DGG/LABCG, 1° Curso de Análise de Bacias, Rio de Janeiro, 68 p.

Dias, G.S., Macambira, M.J.B., Dall'Agnol, R., Soares, A.D.V. and Barros, C.E.M. (1996). Datação de zircões de sill de metagabro: comprovação de idade arqueana da Formação Águas Claras, Carajás-Pará. *Simp. Geol. Amazônia, 5, Belém, SBG, Bol. Res. Expandidos*, p. 376-379.

DOCEGEO (1988). Revisão litoestratigráfica da Província Mineral de Carajás. In: *Cong. Bras. Geol., 35, Belém, 1988. Anexo aos Anais...* Belém, SBG, p. 1-54.

Dorr, J.V.M. (1969). Physiographic, stratigraphic and structural development of the Quadrilátero Ferrífero, Minas Gerais, Brazil. USGS, Washington, Prof. Paper, 641A, 110 p.

Dorr, J.V.M. (1945). Manganese and iron deposits of Morro do Urucum, Mato Grosso, Brazil. *U.S. Geological Survey Bulletin*, 946A, 47p.

Dreher, A.M., Almeida, M.E., Ferreira, A.L., Brito, M.F., Popini, M.V. and Monteiro, M.A. (1999). Veios e brechas hidrotermais da Província Aurífera do Tapajós: aspectos texturais e implicações para a exploração de ouro primário. In: *Simpósio de Geologia da Amazônia, 6, Boletim de Resumos Expandidos, Sociedade Brasileira de Geologia*, p.114-117.

Dreher, A.M., Vlach, S.F.R. and Martini, S.L. (1998). Adularia associated with epithermal gold veins in the Tapajós mineral province, Pará State, Northern Brazil. *Revista Brasileira de Geociências*, 28(3), 397-404.

Duarte, B.P. and Pires, F.R.M. (1996). On the origin of tourmaline in the Passagem gold-tourmaline deposits, Quadrilátero Ferrífero, MG. *Cong. Bras. Geol., 39, Salvador, SBG, Anais*, vol. 7, 193-194.

Ellis, J., Fesefeldt, K., Fontboté, L., Lenzi, J.A., Camargo Souto, J., Gindri, M. and Mücke, A. (1995). Mineralizaciones auríferas em lãs cercanias de Minas de Corrales, "Isla Cristalina de Rivera", N-Uruguay. In: *Simpósio Nacional de Estudos Tectônicos, 5, Gramado, Sociedade Brasileira de Geologia*, 343-346.

Espourteille, F. and Fleischer, R. (1988). Minas de chumbo de Boquira, Bahia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 91-99. DNPM, Brasília, vol. III, 91-99.

Façanha da Costa, H. (1966). Novo Distrito Ferrífero do Brasil, Rio Jatapu, Amazonas. Eng. Min. Met., Rio de Janeiro, vol. 18, 113-116 and 209-212.

Faraco, M.T.L., Carvalho, J.M.A. and Klein, E.L. (1996). *Carta Metalogenética da Província de Carajás, SE do Estado do Pard, Folha Araguaia (SB-22).* Nota Explicativa, Belém, CPRM, 28 p.

Faraco, M.T.L. and Carvalho, J.M.A. (1994). *Carta Metalogenética Previsional do Pará e Amapá, escala 1:1.000.000*, CPRM, Belém.

Faraco, M.T.L. and McReath, I. (1998). Mineralizações da suíte Vila Nova na serra do Ipitinga (noroeste do estado do Pará). In: *Congresso Brasileiro de Geologia, 40, Belo Horizonte, Anais, Sociedade Brasileira de Geologia*, p. 149.

Faraco, M.T.L. (1990). *Evolução petrológico-geoquímica das rochas da Suíte Metamórfica Vila Nova na Serra do Ipitinga (NW do Pará).* M.Sc. Thesis, UFPA, Belém, 346 p.

Faraco, M.T.L. (1997). *Evolução petroquímica e metalogenética das rochas e mineralizações associadas à Suíte Vila Nova na Serra do Ipitinga (NW do Pará).* Doctorate Thesis, Universidade Federal do Pará, Belém, 193 p.

Faria, L.F. (1997). *Controle e tipologia de mineralizações de grafita flake no nordeste de Minas Gerais e sul da Bahia.* M.Sc. Thesis, UFMG, Belo Horizonte, 102 p.

Farias, N.F. and Saueressig, R. 1982. Pesquisa geológica na jazida de cobre Salobo-3A. *Simp. Geol. Amazônia, 1, Belém, Anais*, 2, 39-45.

Farias, N.F., Santos, A.B.S., Biagini, D.O., Vieira, E.A.P., Martins, L.P.B. and Saueressig, R. (1984). Jazidas Cu-Zn da área Pojuca, Serra dos Carajás, PA. *Congr. Bras. Geol., 33, Rio de Janeiro, SBG, Anais*, 8, 3658-3668.

Fernandes, L.A.D., Meneget, R., Costa, A.F.U., Koester, E., Porcher, C.C., Tommasi, A., Kroemer, E., Rangrab, G.E. and Cammazato, E. (1995). Evolução tectônica do cinturão Dom Feliciano no escudo sul-rio-grandense: Parte I - uma contribuição a partir do registro geológico. *Revista Brasileira de Geociências*, 25(4), 351-374.

Ferran, A. (1988). Depósito de ouro de Salamangone e Mutum, Calçoene, Amapá. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S, vol. III, 581-588. DNPM/CVRD.

Ferran, A. (1988). Mina de ouro de São Francisco, Currais Novos, Rio Grande do Norte. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 589-595. DNPM/CVRD, Brasília.

Ferreira Filho, C.F. (1985). *Geologia e mineralizações sulfetadas do Prospecto Bahia, Província Mineral de Carajás.* M.Sc. Thesis, IG/UnB, Brasília, 112 p.

Ferreira Filho, C.F. and Danni, J.C.M. (1985). Petrologia e mineralizações sulfetadas do Prospecto Bahia, Carajás. *Simp. Geol. Amazônia, 2, Belém, SBG, Anais*, vol. 3, 34-47.

Ferreira Filho, C.F., Kamo, S.L., Fuck, R.A., Krogh, T.E. and Naldrett, A.J. (1994). Zircon and rutile U-Pb geochronology of the Niquelândia layered mafic and ultramafic intrusion, Brazil: constraints for the timing of the magmatism and high grade metamorphism. *Precambrian Research*, 68, 241-255.

Ferreira Filho, C.F., Nilson, A.A. and Naldrett, A.J. (1992). The Niquelândia mafic-ultramafic complex, Goiás, Brazil: a contribution to the ophiolite ' stratiform controversy based on new geological and structural data. *Precambrian Research*, 57, 1-19.

Figueiredo, B.R. (1992). Metamorphism of the polymetallic Serrote da Laje deposit, northeastern Brazil. *Proceedings 48th IAGOD Symposium*, 491-504.

Fleischer, R. (1976). A pesquisa de chumbo no Brasil. *Congr. Bras. Geol., 29, Ouro Preto, SBG, Anais*, 1, 19-32.

Fleischer, R. (1998). A rift model for the sedimentary diamond deposits of Brazil. *Mineralium Deposita*, 33, 238-254.

Fleischer, R. and Espourteille, F.S. (1999). The Boquira lead-zinc mine in Central Bahia, Brazil. In: *Base Metal Deposits of Brazil*, eds. Silva, M.G. and Misi, A., pp. 44-53. MME/CPRM/DNPM, Belo Horizonte.

Fleischer, R. and Routhier, P. (1973). The co-sanguineous origin of a tourmaline-bearing gold deposit: Passagem de Mariana, Brazil. *Economic Geology*, 68, 11-22.

Fonseca, E., Lobato, L.M. and Baars, F.J. (1997). Petrochemistry of auriferous volcano-sedimentary Riacho dos Machados Group, Central-Eastern Brazil: Geotectonic implications for shear-hosted gold mineralizations. *Journal of South American Earth Science*, 10, 423-443.

Force, E.R. and Maynard, J.B. (1991). Manganese: Syngenetic deposits on the margins of anoxic basins. In: Sedimentary and diagenetic mineral deposits: a basin analysis approach to exploration, eds. Force E.R., Eidel J.J. and Maynard J.B. *Reviews in Economic Geology*, 5, 147-157.

Forman, A.J.M. and Waring, M.H. (1981). L'uranium en Amérique du Sud et plus spécialement dans la province uranifére brésilienne. *Chronique de la Recherche Minière*, 6, 5-49.

Fortes, P.T.F.O. (1996). *Metalogenia dos depósitos auríferos Mina III, Mina Nova e Mina Inglesa, Greenstone Belt de Crixás, Goiás*. Doctorate Thesis, UnB, Brasília, 177 p.

Fortes, P.T.F.O. and Coelho, R.F. (1997). Caracterização do minério e do rejeito das jazidas auríferas Mina III, Mina Nova e Mina Inglesa, Greenstone Belt de Crixás, Goiás. In: *Caracterização de Minérios e Rejeitos de Depósitos Minerais Brasileiros*, ed. Marini, O.J., pp. 30-33. DNPM, Brasília.

Fortes, P.T.F.O., Cheilletz, A., Giuliani, G. and Ferrand, G. (1997). A Brasiliano age (500±5Ma) for the Mina III gold deposit, Crixás Greenstone Belt, Central Brazil. *International Geology Review*, 39, 449-460.

Franz, J.C. (1997). *Petrologia e hidrotermalismo dos granitóides estaníferos do Rio Grande do Sul*. Doctorate Thesis, IG/UnB, Brasília, 264 p.

Franz, J.C., Coelho, C.E.S. and Botelho, N.F. (1998). Gênese e evolução dos fluidos envolvidos na alteração hidrotermal e na mineralização de estanho em granitos do Rio Grande do Sul. *Revista Brasileira de Geociências*, 28(3), 301-306.

Freitas-Silva, F.H., Dardenne, M.A. and Jost, H. (1991). Lithostructural control of the Morro do Ouro, Paracatu, Minas Gerais, gold deposit. In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 681-683. Belo Horizonte, 1991. A.A. Balkema, Rotterdam.

Freitas-Silva, F.H. (1996). *Metalogênese do depósito do Morro do Ouro, Paracatu, MG*. Doctorate Thesis, IG/UnB, Brasília, 337 p. (unpublished).

Freitas-Silva, F.H. (1999). Geologia da Serra Leste. In: Projeto Agrícola Serra Leste, Curionópolis, PA, Companhia de Promoção Agrícola – CAMPO, 1, 171-208.

Freitas-Silva, F.H. and Dardenne, M.A. (1997). Pb/Pb Isotopic patterns of galenas from Morro do Ouro (Paracatu Formation), Morro Agudo/Vazante (Vazante Formation) and Bambuí Group Deposits. *South-American Symp. on Isotope Geol., Campos do Jordão, SP, Ext. Abst.*, 118-120.

Frota, G.B. and Bandeira, S.A.B. (1997). Depósito de Enxofre de Castanhal, Sergipe. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C., Queiroz E.T. and Coelho C.E.S., vol. IV-B, 303-316. DNPM/CPRM.

Fuck, R.A., Jardim de Sá, E.F., Pimentel, M.M., Dardenne, M.A. and Pedrosa-Soares, A.C. (1993). As faixas de dobramentos marginais do Cráton do São Francisco: síntese dos conhecimentos. In: *O Cráton do São Francisco*, eds. Dominguez, J.M.L. and Misi, A., pp. 165-181. Salvador, SBG/SGM/CNPq.

Fuck, R.A., Marini, O.J., Dardenne, M.A. and Figueiredo, A.N. (1988). Coberturas metassedimentares do Proterozóico Médio: os grupos Araí e Paranoá na região de Niquelândia-Colinas, Goiás. *Revista Brasileira de Geociências*, 18, 54-62.

Fuck, R.A., Pimentel, M.M. and Silva, J.H.D. (1994). Compartimentação tectônica na porção oriental da Província Tocantins. *Anais 38° Cong. Bras. Geol.*, Camboriú, SBG, 1, 215-216.

Galbiatti, H.F. (1999). *Natureza e controle estrutural de mineralização aurífera (Jacutinga) na Mina de Cauê, Itabira, MG*. M.Sc. Thesis, UFOP, Ouro Preto, 204 p.

Galbiatti, H.F., Pereira, M.C. and Fonseca, M.A. (1997). Estruturação dos corpos auríferos (jacutingas) na mina de Cauê, Itabira, MG. *Simp. Geol. Minas Gerais, 9, Ouro Preto, Bol.* 14, 60-62.

Galbiatti, H.F., Pereira, M.C. and Fonseca, M.A. (1999). Natureza e controle estrutural de mineralização aurífera (Jacutinga) na Mina de Cauê, Itabira, MG. *Simp. Geol. Centro-Oeste, 7, Brasília, SBG, Bol. Resumos*, p. 74.

Gallo, M.B.M. (1991). The Romaria diamond-bearing cretaceous conglomerate. In: *Field Guide Book, Fifth International Kimberlite Conference*, eds. Leonardos, O.H., Meyer, H.O.A. and Gaspar, J.C. CPRM, Special Publication 3/91, Brasília, 37-43.

Galvão, C.F., Vianna, I.A., Nonato, I.F.B.P. and Brito, R.S.C. (1986). Depósito de magnetita vanadífera da Fazenda Gulçari, Maracás, Bahia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C. and Coelho C.E.S., vol. II, 493-501. DNPM, Brasília.

Geisel Sobrinho, E., Raposo, C., Prates, S.P., Matos, E.C. and Alves, J.V. (1980). Jazidas uraníferas de Lagoa Real, Bahia. *Congr. Bras. Geol., 31, Camboriú, SBG, Anais*, 3, 1499-1512.

Geraldes, M.R., Toledo, F.H., Figueiredo, B.R. and Tassinari, C.C.G. (1996). Contribuição a geocronologia do sudoeste do Cráton Amazônico. *Cong. Bras. Geociências, 39, Salvador, SBG, Anais*, 2, 554-557.

Gibbs, A.K., Wirth, K.R, Hirata, W.K. and Olszewski, Jr. W.J. (1986). Age and composition of the Grão Pará Group volcanics, Serra dos Carajás. *Revista Brasileira de Geociências*, 16, 201-211.

Gibbs, A.K. and Barron, C.N. (1983). The Guiana Shield Reviewed. *Episodes*, 1983(2), 7-14.

Gibbs, A.K. and Barron, C.N. (1993). *The Geology of the Guiana Shield*. Oxford University Press, New York, 245 p.

Giuliani, G., Olivo, G.R., Marini, O.J. and Michel, D. (1993). The Santa Rita Gold deposit in the Proterozoic Paranoá Group, Goiás, Brazil: an example of fluid mixing during ore deposition. *Ore Geology Review*, 8, 503-523.

Gomes, J.C.M. (1986). As minas de Águas Claras, Mutuca e Pico e outros depósitos de minério de ferro no Quadrilátero Ferrífero, Minas Gerais: In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. II, 65-75. DNPM, Brasília.

Gonzaga, G.M. and Tompkins, L.A. (1991). Geologia do Diamante. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVA, 53-116. DNPM/CPRM, Brasília.

Gorayeb, P.S.S. (1989). Corpos serpentiníticos da Faixa Araguaia na região de Araguacema-Pequizeiro-Conceição do Araguaia (Goiás-Pará). *Revista Brasileira Geociências*, 19(1), 51-62.

Gray, F. and Orris, G.J. (1993). Placer diamond. In: USGS and Corporation Venezuelano de Guyana, ed. Técnica Minera, C.A., *Geology and Mineral Assessment of Venezuelan Guyana Shield*, pp. 86-88. USGS 2026, Washington.

Gray, F., Cox, D.P., Orris, G.J., Page, N.J., Wynn, J.C., Brooks, W.E. and Bliss, J.D. (1993). Mineral Resource Assessment of the Venezuelan Guayana Shield. In: *Geology and Mineral Resource Assessment of the Venezuelan Guayana Shield*, pp. 55-98. U.S. Geological Survey/ Corporación Venezolana de Guayana, Técnica Minera, C.A., United States Geological Survey Bulletin 2062, Washington.

Grubb, P.L. (1979). Genesis of Bauxite Deposits in Lower Amazonian Basin and Guianas Coastal Plains. *Economic Geology*, 74, 735-750.

Haddad, R. (1981). *Mineralizações auríferas no complexo anelar de Taperinaba, CE*. M.Sc. Thesis, UnB, Brasília, p.

Hagemann, S., Brown, P.E. and Walde, D.H.G. (1992). Thin-skinned thrust mineralization in the Brasília Fold Belt: the example of the Luziânia gold deposit. *Mineralium Deposita*, 27(4), 293-303.

Haralyi, N.L.E. and Walde, D.H.G. (1986). Os minérios de ferro e manganês da região de Urucum, Corumbá, Mato Grosso. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. II, 127-144. DNPM, Brasília.

Hartmann, L.A., Silva, L.C., Remus, M.V.D., Leite, J.A. and Philipp, R.P.

(1998). Evolução geotectônica do sul do Brasil e Uruguai entre 3,3Ga e 470Ma. *Cong. Urug. Geol., 2, Punta Del Este*, 277-284.

Hasui, Y. and Cordani, U.G. (1968). Idades potássio-argônio de rochas eruptivas mesozóicas do Oeste Mineiro e Sul de Goiás. In: *Congresso Brasileiro de Geologia, 22, Belo Horizonte, 1968. Anais*... Belo Horizonte, SBG, 139-143.

Hasui, Y. and Costa, J.B.S. (1990). O cinturão Araguaia: um novo enfoque estrutural-estratigráfico. *Cong. Bras. Geol., 36, Natal, SBG, Anais*, 6, 2535-2545.

Hasui, Y. and Haralyi, N.L.E. (1991). Aspectos litoestruturais e geofísicos do soerguimento do Alto Paranaíba. *Geociências*, **10**, 57-77.

Hasui, Y., Costa, J.B.S. and Haralyi, N.L.E. (1994). Estrutura em Quilha no Brasil Central, uma feição fundamental na geologia de Goiás e Tocantins. *Geociências*, **13**(2), 463-497.

Heald, S.G., Foley, N.K. and Hayba, D.O. (1987). Comparative antomy of volcanic-hosted epithermal depsits: acid-sulphate and adularia-sericite types. *Economic Geology*, **82**, 1-26.

Hennies, W.T. (1969). Minério de ferro oolítico no Brasil. *Anais 23º Cong. Bras. Geologia, Salvador, SBG*, 177-182.

Hippert, J.F. and Massucatto, A.J. (1998). Phyllonitization and development of kilometer-size extension gashes in a continental-scale strike-slip shear zone, North Goiás, Central Brazil. *Journal of Structural Geology*, 20(4), 433-445.

Hirata, W.K., Rigon, J.C., Kadekaru, K., Cordeiro, A.A.C. and Meireles, E.M.A. (1982). Geologia regional da Província Mineral de Carajás. *An. 1º Simp. Geol. Amazônia, Belém, SBG*, **1**, 100-110.

Hitzmann, M.W. (1995). Mineralization in the Irish Zn-Pb(Ba-As) ore field. In: Irish Carbonate-Hosted Zn-Pb Deposits, eds. Anderson, K., Ashton, J., Earls, G., Hitzmann, M. and Tear, S. *SEG Guide book*, 21, 132-137.

Hitzmann, M.W., Oreskas, N.E and Einandi, M.T. (1992). Geological characteristics and tectonic setting of Proterozoic iron-oxide (Cu-U-Au-REE) deposits. *Precambrian Research*, **58**, 241-287.

Hitzmann, M.W., Thormann, C.H., Romagna, G., Oliveira, T.F., Dardenne, M.A. and Drew, L.J. (1995). The Morro Agudo Zn-Pb deposit, Minas Gerais, Brazil: a Proterozoic Irish-type carbonate-hosted SEDEX-replacement deposit. *Annual Meeting, New Orleans, ESA, Abstracts*, 408 p.

Hoppe, A. and Schobbenhaus, C. (1990). Geology and mineral resources of Amazonia. *Zentralblatt für Geologie und Paläontologie Teil I*, **12**, 1787-1837.

Horbach, R. and Marimon, M.P.C. (1980). O depósito de cobre de Serrote da Laje, Arapiraca, Alagoas. *Cong. Bras. Geol., 35, Belém, SBG, Anais*, 1, 1-15.

Horbach, R. and Marimon, M.P.C. (1988). Depósito de cobre do Serrote da Laje, Arapiraca, Alagoas. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 61-70. DNPM, Brasília.

Huhn, S.R.B. (1991). Controle estrutural dos depósitos e ocorrências auríferas no terreno granito-greenstone da região de Rio Maria. *Simp. Geol. Amazônia, 3, Belém, SBG, Anais*, 211-219.

Huhn, S.R.B. (1992). *Geologia, controle estrutural e gênese do depósito aurífero Babaçu, região de Rio Maria, sul do Pará*. M.Sc. Thesis, IG/UnB, Brasília, 169 p.

Huhn, S.R.B. and Nascimento, J.A.S. (1997). São os depósitos cupríferos de Carajás do tipo Cu-Au-U-ETR? In: *Contribuição à Geologia da Amazônia*, eds. Costa, M.L. and Angélica, R.S. SBG, Belém, 143-160.

Huhn, S.R.B., Souza, C.I.J., Albuquerque, M.C., Leal, E.D. and Drustolin, V. (1999). Descoberta do depósito Cu (Au) Cristalino: geologia e mineralizações associadas, região da Serra do Rabo, Carajás PA. *Simp. Geol. Amazônia, Manaus, SBG, Anais*, 140-143.

Ianhez, A.C., Ribeiro, D.T. and Pamplona, R.I. (1997). Depósito de amianto de Cana Brava, Minaçu, Goiás. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVB, 47-62. DNPM, Brasília.

Iwanuch, W. (1999). Evolução geológica com base em dados geocronológicos na porção sudeste do Estado do Amazonas e do trato contíguo do norte do Estado de Mato Grosso, na região do baixo e médio Rio Juruena. In: *Simp. Geol. Amaz., 6, Manaus, SBG, Bol. Res. Exp.*, 467-470.

Jacobi, P. (1999). The discovery of epithermal Au-Cu-Mo Proterozoic deposits in the Tapajós province, Brazil. *Revista Brasileira de Geociências*, **29**(2), 277-279.

Jardim de Sá, E.F. and Salim, J. (1980). Reavaliação dos conceitos estratigráficos na região Seridó (RN-PB). *Mineração e Metalurgia*, **421**, 16-28.

Jardim de Sá, E.F., Legrand, J.M., Galindo, A.C., Martins de Sá, J. and Hachspacher, P.C. (1986). Granitogênese Brasíliana no Seridó: maciço de Acari. *Revista Brasileira de Geociências*, 16, 95-105.

JICA/MMAJ (1999). *Report on the mineral exploration in the Alta Floresta area, Federative Republic of Brazil (Phase I)*. Japan International Cooperation Agency/Metal Mining Agency of Japan (unpublished report).

JICA/MMAJ (2000). *Report on the mineral exploration in the Alta Floresta area, Federative Republic of Brazil (Phase II)*. Japan International Cooperation Agency/Metal Mining Agency of Japan (unpublished report).

João, X.S.J., Frizzo, S.J., Marinho, P.A.C., Carvalho, J.M.A., Silva Neto, C.S., Souza, A.N. and Guimarães, L.R. (1978). Projeto Sudoeste do Amapá. *Geologia Básica*, 7, 125 p.

João, X.S.J. and Marinho, P.A. (1982). Catametamorfitos arqueanos da região centro-leste do Território Federal do Amapá. In: *Simpósio de Geologia da Amazônia, 1, Belém,, Anais, Sociedade Brasileira de Geologia*, 2, 207-228.

Jost, H. (1981). *Geology and metallogeny of the Santana da Boa Vista region, South Brazil*. Ph.D. Thesis, University of Georgia, Athens, USA.

Jost, H., Resende M.G., Kuyumjian R.M., Queiroz C.L., Osborne G.A., Blum M.L.B., Pires A.C.B. and Morães R.A. (2000). Terrenos arqueanos de Goiás. *Revista Brasileira de Geociências* (in press).

Justo, L.J.E.Q. and Souza, M.M. (1986). Jazida de nióbio do Morro dos Seis Lagos, Amazonas. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho C.E.S., vol. II, 463-468. DNPM, Brasília.

Kaminsky, F.V., Zakharchenko, O.D., Channer, D.M.D.R Blinova, G.K. and Maltsev, K.A. (1998). Diamonds from the Guaniamo area. In: *International Kimberlite Conference, 7, Cape town, Extended Abstracts*, 395-397.

Kegel, W. (1955). Geologia do fosfato de Pernambuco. *Boletim Divisão de Geologia e Mineralogia*, **157**, 54 p.

Kishida, A. And Riccio, L. (1980). Chemostratigraphy of lava sequence from the Rio Itapicuru greenstone belt, Bahia, Brazil. *Precambrian Research*, **11**, 161-178.

Kishida, A., Sena, F.O., and Silva, F.C.A. (1991). Rio Itapicuru greenstone belt: geology and gold mineralization. In: *Brazil Gold' 91*, ed. Ladeira, E.A., pp. 231-234, Rotterdam, A.A. Balkema.

Kotschoubey, B. (1988). Geologia do alumínio. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 599-619. DNPM, Brasília.

Kotschoubey, B. and Truckenbrodt, W. (1981). Evolução poligenética das bauxitas do Distrito de Paragominas- Açailândia (Estados do Pará e Maranhão). *Revista Brasileira de Geociências*, **11**, 193-202.

Kotschoubey, B., Hieronymus, B., Rodrigues, O.B. and Amaral, R.T. (1996). Basaltos e serpentinitos da área da Serra do Tapa (PA). Prováveis testemunhos de um complexo ofiolítico pouco evoluído e desmembrado. *Cong. Bras. Geol., 39, Salvador, SBG, Anais*, 6, 25-29.

Kotschoubey, B., Truckenbrodt, W. and Hieronymus, B. (1997). Bauxite deposits of Paragominas. In: *Brazilian Bauxites*, eds. Coelho, A., Boulangé, B., Melfi, A. and Lucas, Y., pp. 75-106. USP/FAPESP/ ORSTOM.

Krauss, L.A.A. and Amaral, A.J.R. (1997). Depósitos de Gipsita da Casa de Pedra, Pernambuco. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVC, 159-167. DNPM, Brasília.

Kuyumjian, R.M. (1995). Diversity of fluids in the origin of the Chapada Cu-Au deposit, Goiás. *Revista Brasileira de Geociências*, **25**(3), 203-205.

Kuyumjian, R.M. (2000). The magmatic arc of western Goiás: A promising exploration target. In: *Base Metal Deposits of Brasil*, eds. Silva, M.G. and Misi, A., pp. 69-74. MME/CPRM/DNPM, Belo Horizonte.

Kuyumjian, R.M., Campos, J.E.G., Oliveira, C.G. and Queiroz, C.L. (1999). Registros da evolução transamazônica na província estrutural do Tocantins: exemplo da região de Campinorte-Alto Paraíso (GO). *Revista Brasileira de Geociências* (in press).

Kuyumjian, R.M. (1991). A suggested hydrothermal exalative origin for the Chapada copper-gold deposit, Brazil. In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 231-234. Proceedings.

Kyle, J.R. and Misi, A. (1997). Origin of Zn-Pb-Ag sulfide mineralization within Upper Proterozoic phosphate-rich carbonate strata, Irecê Basin, Bahia, Brazil. *International Geology Review*, **39**, 383-399.

Lacerda, H. (1986). Tipologia das mineralizações auríferas da área do Rio do Carmo, Cavalcante, Goiás. *Anais 34º Cong. Bras. Geol.*, SBG, 5, 1946-1955.

Lacerda, H. (1991). Gold in Central Brazil: Types of deposits, their economic significance and regional distribution. In: *Brazil Gold'91*, eds. Ladeira, E.A., pp. 195-202, Balkema.

Ladeira, E.A. (1980). *Metallogenesis of gold at the Morro Velho Mine and in Nova Lima, Quadrilátero Ferrífero, Minas Gerais, Brazil.* Ph.D. Thesis, University of Western Ontario, London, 272 p.

Ladeira, E.A. (1988). Metalogenia dos depósitos de ouro do Quadrilátro Ferrífero, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 301-375. DNPM, Brasília.

Ladeira, E.A. (1991). Genesis of gold in Quadrilétro Ferrífero: a remarkable case of permanent recycling and inheritance. In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 11-30. Rotterdam, A.A. Balkema.

Lafon, J.M. and Macambira, J.B. (1990). Age Archéen de la granodiorite Cumaru (Serra dos Gradaús, Pará, Brésil). *Comptes Rendus de l'Académie des Sciences, Paris*, **310**, 1641-1653.

Lafon, J.M., Scheller, T. (1994). Geocronologia Pb/Pb em zircão do granodiorito Cumaru, Serra dos Gradaús, Pará. *Simp. Geol. Amazônia, 4, Belém, SBG, Bol. Resumos*, 321-323.

Lafon, J.M., Rossi, Ph., Delor, C., Avelar, V. and Faraco, M.T.L. (1998). Novas testemunhas de relíquias arqueanas na crosta continental paleoproterozóica da província Maroni-Itacaiúnas (sudeste do Escudo da Guianas). In: *Congresso Brasileiro de Geologia, 40, Belo Horizonte, Anais, Sociedade Brasileira de Geologia*, p. 64.

Lafrance, J., Bardoux, M., Voicu, G., Stevenson, R. and Machado, N. (1999). Geological and Metallogenic Environments of Gold Deposits of the Guiana Shield: A Comparative Study between St-Élie (French Guiana) and Omai (Guyana). *Exploration Mining Geology* (in press).

Ledru, P. and Bouchot, V. (1993). Revue des minéralisation auriféres du Craton précambrien de São Francisco (Brésil) et discussion sur leurs contrôles structuraux. *Chronique de la Recherche Miniére*, S11, 5-20.

Legrand, J.M., Deutsch, S. and Souza, C.L. (1991). Datação U/Pb e granitogênese do macicó de Acari (RN). *Simp. Geol. Nordeste, 14, Natal, SBG, Anais*, 172-174.

Legrand, J.M., Melo Jr., G., Archanjo, C.J., Salim, J., Souza, L.C. and Maia, H.M. (1993). Mineralizações da Faixa Seridó: um processo hidrotermal do fenômeno tectono-magmático Brasíliano. *Simp. Geol. Nordeste, 15, Natal, SBG, Anais*, 185-188.

Legrand, J.M., Melo Jr., G., Silva, W.L. and Souza Neto, J.A. (1996). Origin and classification of gold mineralization in the Seridó fold belt and basement rocks, northeast Brazil. *Cong. Bras. Geol., 38, Salvador, SBG, Anais*, 221-224.

Lenharo, S.L.R. (1998). *Evolução magmática e modelo metalogenético dos granitos mineralizados da região de Pitinga, Amazonas, Brasil.* Doctorate Thesis, USP, São Paulo, 290 p.

Lenz, G.R. and Ramos, B.W. (1985). Combustíveis Fósseis Sólidos no Brasil: Carvão, Linhito, Turfa e Rochas Oleígenas. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. 1, 3-37. DNPM/CVRD, Brasília.

Leonardos, O.H., Santos, M.D., Giuliani, G. and Araújo, L.R. (1991). The Cumaru mesothermal granodiorite-hosted gold mineralization, Amazon Craton, Brazil. In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 557-562. Proceedings... Rotterdam, Balkema.

Leterrier, J., Jardim de Sá, E.F. and Macedo, M.H.F. (1990). Magmatic and geodynamic signature of the Brasíliano Cycle plutonism in the Seridó Belt. *Cong. Bras. Geol., 36, Natal, SBG, Anais*, **4**, 1600-1655.

Lima, M.I.C. (1984). Províncias geológicas do Cráton Amazônico, em território brasileiro. In: *Simp. Amazônico, 2, Manaus, 1984. Anais*... Manaus, DNPM, 9-23.

Lima, R.E. (1993). *Evolução geológica e controles dos depósitos de talco da região de Itaiacoca-Abapã, PR.* M.Sc. Thesis, UnB, Brasília, 139 p.

Lima, R.E. and Dardenne, M.A (1987). Geologia e controle da Mina Grande da Costalco, Itaiacoca, PR. *Simp. Sul-Brasileiro Geologia, 3, Curitiba, SBG, Anais*, 37-38.

Lima, T.M. (1997). *Geologia, estratigrafia e petrologia da porção sul do complexo máfico-ultramáfico de Cana Brava, Goiás.* M.Sc. Diss, IG/UnB, Brasília, 312 p. (unpublished)

Lima, M.I.C., Montalvão, R.M.G., Issler, R.S., Oliveira, A., Basei, M.A.S., Araujo, J.F.V. and Silva, G.G. (1974). *Geologia da folhas NA/NB.22-Macapá.* Rio de Janeiro, Projeto RADAM/DNPM, 1-120.

Lindenmayer, Z.G. (1981). Geological evolution of Vale do Rio Curaçá and of copper mineralized mafic-ultramafic bodies. *Geol. Recur. Miner. Bahia, Textos Básicos*, **1**, 73-10.

Lindenmayer, Z.G. (1990). *Salobo, Carajás, Brazil: Geology, geochemistry and metamorphism.* Ph.D. Thesis. University of Western Ontario, London, Canada, 407 p.

Lindenmayer, Z.G. (1998). O depósito Cu (Au-Ag-Mo) do Salobo, Serra dos Carajás, revisitado. *Workshop, Depósitos Minerais Brasileiros de Metais-Base*, Salvador, CAPES-PADCT, CPGG-UFBA, ADIMB, p. 29-37.

Lindenmayer, Z.G. (1999). Ore genesis at the Salobo copper deposit, Serra dos Carajás. In: *Base Metal Deposits of Brazil*, eds. Silva, M.G. and Misis, A., pp. 33-43. MME/CPRM/DNPM, Belo Horizonte.

Lindenmayer, Z.G. and Fyfe, W.S. (1994). The Salobo Cu (Au,Ag,Mo) deposit, Serra dos Carajás, Brazil: *Cong. Geol. Chileno, 7, Concepción, Chile, Atas*, **2**, 840-842.

Lindenmayer, Z.G., Laux, J.H. and Vieira, A.C. (1995). O papel da alteração hidrotermal nas rochas da bacia Carajás. *Bol. Museu Paraense Emílio Goeldi*, Série Ciências da Terra, **7**, 125-145.

Lindenmayer, Z.G., Ronchi, L.H. and Laux, J.H. (1998). Geologia e geoquímica de Cu-Au primária da mina de Au do Igarapé Bahia, Serra dos Carajás. *Revista Brasileira de Geociências*, **28**(3), 257-268.

Litherland, M., Annels, R.N., Appleton, J.D., Berrangé, J.P., Bloomfield, K., Darbyshire, D.P.F., Fletcher, C.J.N., Hawkins, M.P., Klinck, B.A., Mitchell, W.I., O'Connor, E.A., Pittfield, P.E.J., Power, G. and Webb, B.C. (1986). The geology and mineral resources of the Bolivian Precambrian shield. *Overseas Memoir British Geological Survey* **9**, 153 p.

Lobato, L.M. and Fyfe, W.S. (1990). Metamorphism, metasomatism and mineralization at Lagoa Real, Bahia, Brazil. *Economic Geology*, **85**, 968-989.

Lobato, L.M. and Vieira, F.W.R. (1998). Styles of hydrothermal alteration and gold mineralization associated with the Nova Lima Group of the Quadrilátero Ferrífero: Part II, The archean mesothermal gold-bearing hydrothermal system. *Revista Brasileira de Geociências*, **28**(3), 355-366.

Lobato, L.M., Baars, F.J. and Jost, H. (1999). The potential for VMS deposits in the greenstone belts in and around the southern portion of the São Francisco Craton, Brazil. In: *Base Metal Deposits of Brazil*, eds. Silva, M.G. and Misi, A., pp. 100-108. MME/CPRM/DNPM, Belo Horizonte.

Lobato, L.M., Forman, J.M.A., Fuzikawa, K., Fyfe, W.S. and Kerrich, R. (1983). Uranium in overthrust Archean basement, Bahia, Brazil. *Canadian Mineralogist*, **21**, 647-654.

Lobato, L.M., Vieira, F.W.R., Ribeiro-Rodrigues, L.C., Pereira, L.M.M., Menezes, M.G., Junqueira, P.A. and Pereira, S.L.N. (1998). Styles of hydrothermal alteration and gold mineralizations associated with the Nova Lima Group of the Quadrilátero Ferrífero: Part I, Description of selected gold deposits. *Revista Brasileira de Geociências*, **28**(3), 339-354.

Lucas, Y. (1997). The Bauxite of Juriti. In: *Brazilian Bauxites*, eds. Carvalho, A., Boulangé, B., Melfi, A.J. and Lucas, Y., pp. 107-133. USP/FAPESP/ORSTOM, São Paulo.

Macambira, M.J.B. and Lafon, J.M. (1995). Geocronologia da Província Mineral de Carajás: síntese dos dados e novos desafios. *Boletim do Museu Paraense Emílio Goeldi*, Série Ciências da Terra, 7, 263-288.

Macambira, M.J.B. and Lancelot, J.R. (1992). Idade U-Pb em zircões de metavulcânica do greenstone do Supergrupo Andorinhas, delimitantes da estratigrafia arqueana de Carajás, estado do Pará. In: *Cong. Bras. Geol., 17, São Paulo, 1992, Boletim de Resumos... São Paulo, SBG*, 2, 188-189.

Macambira, M.J.B. and Silva, V.F. (1995). Estudo petrológico, mineralógico e caracterização das estruturas sedimentares e diagenéticas preservadas na Formação Carajás, Estado do Pará. *Boletim do Museu Paraense Emílio Goeldi*, Série Ciências da Terra, 7, 363-387.

Macambira, M.J.B., Kotschoubey, B., Santos, M.D., Moura, C.A.V. and Ramos, J.F.F. (1986). Estratigrafia e mineralizações primárias de ouro da aba sul do sinclinório de Gradaús – sul do Pará. In: *Cong. Bras. Geol., 34, Goiânia, 1986. Anais*... Goiânia, SBG, **5**, 1956-1968.

Macambira, M.J.B., Lafon, J.M., Dall'Agnol, R., João, X.S.J. and Costi, H.T. (1990). Geocronologia da granitogênese da Província Amazônia central brasileira: Uma revisão. *Revista Brasileira de Geociências*, **20**, 258-266.

Machado, N. and Carneiro, M.A. (1992). U-Pb evidence of Late Archean tectonothermal activity in the southern São Francisco shield, Brazil. *Canadian Journal of Earth Sciences*, **29**, 2341-2346.

Machado, N., Lindenmayer, Z., Krogh, T.E. and Lindenmayer, D. (1991). U-Pb geochronology of Archean magmatism and basement reactivation in the Carajás area, Amazon Shield, Brazil. *Precambrian Research*, **49**, 1-26.

Magalhães, L.F. (1991). *Cinturão de cisalhamento de empurrão Córrego Geral Meia Pataca: geologia, deformação, alteração hidrotermal e mineralizações auríferas associadas (Crixás, Goiás)*. M.Sc. Thesis, UnB, Brasília.

Magalhães, L.F. and Nilson, A.A. (1996). Mineralização aurífera nos grupos Araí e Paranoá na região de Cavalcante – GO e Paranã – TO. *An. 39º Cong. Bras. Geol., Salvador, SBG*, **3**, 284-286.

Magalhães, L.F., Freitas-Silva, F.H., Nilson, A.A. and Coelho, C.E.S. (1998). Estudo de inclusões fluidas em veios de quartzo auríferos do Grupo Paranoá na região de Cavalcante, Goiás. *Cong. Bras. Geol., 40, Belo Horizonte, SBG, Anais*, 167.

Marinho, M.M. (1991). *La séquence volcano-sédimentaire de Contendas-Mirante et la bordure occidentale du Bloc Jequié (Craton du São Francisco, Brésil): un exemple de transition Archeén-Protérozoïque*. Doctorate Thesis, Univ. Chermont-Ferrand, França.

Marini, O.J. and Botelho, N.F. (1986). A província de granitos estaníferos de Goiás. *Revista Brasileira de Geociências*, **16**(1), 19-131.

Marini, O.J., Fuck, R.A., Danni, J.C.M., Dardenne, M.A., Loguércio, S.O. and Ramalho, R. (1984a). As faixas de dobramento Brasília, Uruaçu, Paraguai-Araguaia e o Maciço Mediano de Goiás. In: *Geologia do Brasil*, coords. Schobbenhaus, C., Campos, D.A., Derze, G.R. and Asmus, H.E., pp. 251-303. Brasília, DNPM.

Marini, O.J., Fuck, R.A., Dardenne, M.A. and Danni, J.C.M. (1984b). Província Tocantins, Setores Central e Sudeste. In: *O Pré-Cambriano do Brasil*, coords. Almeida, F.F.M. and Hasui, Y., pp. 205-264. São Paulo, Edgar Blücher.

Marini, O.J. and Queiroz, E.T. (1991). Main geologic-metallogenetic environments and mineral exploration in Brazil. *Ciência e Cultura*, **43**(2), 153-161.

Marques, J.C. (1999). *Petrologia e metalogênese dos corpos máfico-ultramáficos cromitíferos do vale do Rio Jacurici-BA*. Exame Qualificação, UnB, Brasília, 71 p.

Marshak, S. and Alkmim, F.F. (1989). Proterozoic contraction/extension tectonics of southern São Francisco region, Minas Gerais, Brazil. *Tectonics*, **8**(3), 555-571.

Marshak, S., Alkmim, F.F. and Jordt Evangelista, H. (1992). Proterozoic crustal extension and the generation of dome-and-kneel structure in Archean granite-greenstone terrane. *Nature*, **357**, 491-453.

Martin-Bellizzia, C. (1972). Paleotectonica del Escudo de Guyana. In: *Conferencia Geologica Interguianas, 9, Puerto Ordaz, Memória, Ministerio de Minas y Hidrocarburos, Bol. Geol., Publ. Especial*, **6**, 251-305.

Martini, S.L. (1998). An overview of main auriferous regions of Brazil. *Revista Brasileira de Geociências*, **28**(3), 307-314.

Mascarenhas, J.F. and Silva, M.G. (1994). Greenstone belt de Mundo Novo: caracterização e implicações metalogenéticas e geotectônicas no Cráton do São Francisco. *CBPM, Salvador, Série Arquivos Abertos*, **5**, 32 p.

Matos, A.A., Spier, C.A. and Soares, J.W. (1992). Depósitos de cromita da região do Rio Vila Nova, Estado do Amapá, Pará. *Congr. Bras. Geol., 37, São Paulo, SBG, Anais*, 246-247.

Medeiros, E.S. and Ferreira Filho, C.F. (1999). Caracterização geológica e estratigráfica de um PGE reef no complexo máfico-ultramáfico de Niquelândia, Goiás. *Simp. Geol. Centro-Oeste, 7, Brasília, SBG, Bol. Resumos*, p. 35.

Medeiros Filho, C.A. and Meireles, E.M. (1985). Dados preliminares sobre a ocorrência de cromita na área de Luanga. In: *Simp. Geol. Amaz., 2, Belém, SBG, Anais*, **3**, 90-96.

Medeiros Neto, F.A. (1986). Mineralizações auríferas da área Pojuca: extração, transporte e deposição a partir de fluídos hidrotermais salinos. *Cong. Bras. Geologia, 34, Goiânia, SBG, Anais*, **5**, 1969-1981.

Medeiros Neto, F.A. and Villas, R.N. (1985). Geologia da jazida de Cu + Au do corpo 4E/Pojuca, Serra dos Carajás. *Simp. Geol. Amazônia, 2, Belém, SBG, Anais*, **3**, 97-112.

Meireles, E.M. and Silva, A.R.B. (1988). Depósitos de ouro de Serra Pelada, Marabá, Pará. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 547-557, DNPM, Brasília.

Meireles, E.M., Teixeira, J.T., Lourenço, R.S. and Medeiros Filho, C.A. (1982). Geologia, estrutura e mineralização aurífera de Serra Pelada. *Cong. Bras. Geol., 32, Salvador, SBG, Anais*, **3**, 900-911.

Meirelles, M.R. (1986). *Geoquímica e metalogênese dos jaspilitos e rochas vulcânicas associadas, Grupo Grão Pará, Serra dos Carajás*. M.Sc. Thesis, UnB, Brasília 150 p.

Meirelles, M.R. and Dardenne, M.A. (1991). Vulcanismo basáltico de afinidade shoshonítica em ambiente de arco arqueano, Grupo Grão Pará, Serra dos Carajás, Pará. *Revista Brasileira de*

Geociências, **21**, 41-50.

Meirelles, M.R. and Dardenne, M.A. (1993). Geoquímica e gênese dos jaspilites arqueanos da Serra dos Carajás, Pará. *Conj. Bras. Geoquímica, 4, Brasília, SBGq, Anais, vol. Res. Expandidos*, 131-132.

Melfi, A.J., Trescases, J.J. and Oliveira, S.M.B. (1980). Les latérites nickéliféres du Brésil. *Cahiers Orstom, Géologie*, Sér. Géol., **XI**(1), 15-42.

Melfi, A.J., Trescases, J.J., Carvalho, A., Oliveira, S.M.B., Filho, E.R. and Formoso, M.L. (1988). The lateritic ore deposits of Brazil. *U.S. Geological Survey Bulletin*, **41**, 5-36.

Mello, C.H.M.P., Durão, G., Viana, J.S. and Carvalho, C.T.C. (1986). Depósitos de cromita das fazendas Medrado e Ipueira, Município de Senhor do Bonfim, Bahia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.ES., vol. II, 215-234. DNPM, Brasília.

Mello, J.S.C. and Bettencourt, J.S. (1998). Geologia e gênese das mineralizações associadas ao maciço Itaoca, Vale do Ribeira, SP e PR. *Revista Brasileira de Geociências*, **28**(3), 269-284.

Melo Jr., G. and Legrand, J.M. (1993). Mineralizações auríferas em rochas calciossilicáticas: caso da Província scheelitífera da Borborema (RN-PB). *Simp. Geol. Nordeste, 15, Natal, SBG, Anais*, 189-191.

Melo Jr., G., Legrand, J.M. and Almeida, H.L. (1996). Gold mineralization in the Cachoeirinha-Salgueiro fold belt, northeast Brazil: from local features to a regional approach. *Cong. Bras. Geol., 39, Salvador, SBG, Anais*, 210-213.

Melo, M.T.V., Borba, R.R. and Coelho, W.A. (1986). O distrito ferrífero de Itabira: minas de Cauê, Conceição, Dois Córregos, Periquito, Onça, Chacrinha e Esmeril. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhans, C. and Coelho, C.E.S., vol. II, 7-28. DNPM, Brasília.

Mendonça, J.C.G.S., Campos, M., Braga, A.P.G., Souza, E.M., Favali, J.C. and Leal, J.R.L.V. 1985. Jazida de urânio de Itataia, Ceará. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus C. and Coelho C.E.S., vol. I, 121-131. DNPM, Brasília.

Menor, E.A., Dantas, J.C.A. and Sobrinho, A.C.P.L. (1977). Sedimentação fosfática em Pernambuco e Paraíba: Revisão e novos estudos. *Atas 8º Simp. Geol. Nordeste, Campina Grande, SBG*, 1-27.

Milani, E.J. and Zalán, P.V. (1999). An outline of the geology and petroleum systems of the Paleozoic interior basins of South America. *Episodes*, **22**(3), 199-205.

Milési, J.P., Ledru, P., Johan, V., Marcouse, E., Mougeot, R., Lerouge, C., Respaut, J.P. and Sabaté, P. (1996). Hydrothermal and metamorphic events related to the gold mineralizations hosted within detrital sediments in the Jacobina basin (Bahia, Brazil). *Anais 39º Cong. Bras. Geol., Salvador, SBG*, **7**, 218-220.

Milési, J.-P., Egal, E., Ledru, P., Vernhet, Y., Thiéblemont, D., Cocherir, A., Tegyey, M., Martel-Jautin, B. and Lagny, P. (1995). Les minéralisations du Nord de la Guyane Française dans leur cadre géologique. *Chronique de la Recherche Minière*, **518**, 5-58.

Minter, W.E.L., Renger, F.E. and Siegers, A. (1990). Early Proterozoic gold placers of the Moeda Formation within the Gandarela Syncline, Minas Gerais, Brazil. *Economic Geology*, **85**(5), 943-951.

Misi, A. (1992). Geologia e gênese da fosforita de Irecê, Bahia. *Revista Brasileira de Geociências*, **22**(4), 399-406.

Misi, A. and Kyle, J.R. (1994). Upper proterozoic carbonate stratigraphy, diagenesis and stromatolitic phosphorite formation, Irecê Basin, Bahia, Brazil. *Journal of Sedimentalogy Research*, **A64**(2), 299-310.

Misi, A., and Silva, M.G. (1996). Chapada Diamantina Oriental-Bahia, Geologia e Depósitos Minerais. Superintendência de Geologia e Recursos Minerais-SGM, *Série Roteiros Geológicos*, Salvador, Bahia, 194 p.

Misi, A., Iyer, S.S. and Tassinari, C.C.G. (1996). Boquira (2,5Ga) and Morro Agudo (0,65Ga) lead-zinc deposits, Brazil: New SEDEX Subtypes? *Anais 39º Cong. Bras. Geol., Salvador, SBG*, **7**, 251-253.

Misi, A., Iyer, S.S., Tassinari, C.C.G., Coelho, C.E.S., Kyle, J.R., Franca-Rocha, W.J.S. and Gomes, A.S.R. (1999). Integrated studies and metallogenic evolution of the Proterozoic sediment-hosted Pb-Zn-Ag sulfide deposits of the São Francisco Craton, Brazil. In: *Base Metal Deposits of Brazil*, coords. Silva, M.G. and Misi, A., pp. 84-91. MME/CPRM/DNPM, Belo Horizonte.

Misi, A. and Silva, M.G. (1996). *Chapada Diamantina Oriental-Bahia, Geologia e Depósitos Minerais*. Superintendência de Geologia e Recursos Minerais-SGM, Série Roteiros Geológicos, Salvador, Bahia, 194 p.

Molinari, L. (1982). Mineralizações auríferas em Jacobina, Bahia. *Simp. Ouro, 1, Salvador*, 586 p.

Molinari, L. and Scarpelli, W. (1988). Depósitos de ouro de Jacobina, Bahia. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 463-478. DNPM/CPRM, Brasília.

Monteiro, L.V.S. (1997). *Contribuição à gênese das mineralizações de Zn da mina de Vazante, MG*. M.Sc. Thesis., USP, 159 p. (Inedited).

Monteiro, L.V.S., Bettencourt, J.S. and Grasa, R. (1996). Contribuição à gênese das mineralizações de Zn e Pb da mina de Vazante (MG). *Anais 39º Cong. Bras. Geol., Salvador, SBG*, **5**, 501-503.

Monteiro, H., Macedo, P.M., Moraes, A.A., Marchetto, C.M.L., Fanton, J.J. and Magalhães, C.C. (1988). Depósito de ouro Cabaçal I, Mato Grosso. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 535-545. Departamento Nacional de Produção Mineral, Brasília.

Moreira Neto, A.M. and Amaral, A.J.R. (1997). Depósitos de fosfato do Nordeste oriental do Brasil. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVC, 131-151. DNPM/CPRM, Brasília.

Morelli, B., Barreto, L.A., Frota, G.B., Palhano, S.G. and Andrade, E.S. (1982). O Primeiro Depósito Brasileiro de Enxofre Nativo. *An. 32º Cong. Bras. Geol., Salvador, SBG*, **3**, 1073-1085.

Morrone, M. and Daemon, R.F. (1985). Jazida de urânio de Figueira, Paraná. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S., vol. I, 133-142. DNPM/CVRD.

Moura, M.A. (1998). *O maciço granítico Matupá no depósitos de ouro Serrinha (MT): petrologia, alteração hidrotermal e metalogenia*. Doctorate Thesis, UnB, Brasília, 238 p.

Murray, H.H. and Partridge, R. (1982). Genesis of Rio Jari Kaolin. In: Proceedings of the 4th International Clay Conference, Amsterdam, eds. Van Olphen, H. and Venide, F. *Developments in Sedimentology*, Elsevier, **35**, 279-291.

Nascimento, J.A.S. and Biagini, D.O. (1988). Conhecimento atual da jazida de ouro de Lagoa Seca, sul do Pará. In: *Cong. Bras. Geol., 35, Belém, Província Mineral de Carajás, CVRD/SBG 1988. Anexo aos Anais. Belém, SBG*, 143-155.

Neder, R.D., Collins, C., Figueiredo, B.R. and Leite, J.A.D. (1998). O depósito polimetálico de Aripuanã, Mato Grosso, Brasil. *Cong. Bras. Geol., 40, Belo Horizonte, SBG, Anais*, p. 153.

Nilson, A.A. (1981). *The nature of the Americano do Brasil mafic-ultramafic complex and associated sulfide mineralization*. Ph.D. Thesis, University of Western Ontario, 460 p. (inedited).

Nilson, A.A., Botelho, N.F. and Ferreira Filho, C.F. (1994). Rifteamento mesoproterozóico no Centro-Oeste de Goiás. *Resumos 38º Cong. Bras. Geol., Camboriú, SBG*, 258-259.

Nixon, P.H., Davies, G.R., Rex, D.C. and Gray, A. (1992). Venezuela kimberlites. *Journal of Volcanology and Geothermal Research*, **50**, 101-115.

Noce, C.M. (1995). *Geocronologia dos eventos magmáticos, sedimentares e metamórficos na região do Quadrilátero Ferrífero, Minas Gerais*. Ph.D. Thesis, USP, São Paulo, 128 p.

Noce, C.M., Machado, N. and Teixeira, W. (1998). U-Pb geochronology of gnaisses and granitoids in the Quadrilátero Ferrífero (Southern São Francisco Craton): Age constraints for archean and paleoproterozoic magmatism and metamorphism. *Revista Brasileira de Geociências*, **28**(1), 95-102.

Nogueira, A.C.R., Truckenbrodt, W. and Pinheiro, R.V.L. (1995). Formação Águas Claras, Pré-Cambriano da Serra do Carajás: redescrição e redefinição litoestratigráfica. *Boletim do Museu Paraense Emílio Goeldi*, Série Ciências da Terra, 7, 177-197.

Oelofsen, B.W. and Araujo, D.C. (1983). Paleoecological Implications of the Distribution of Mesosaurid Reptiles in the Permian Irati Sea (Paraná Basin), South America. *Revista Brasileira de Geociências*, **13**(1), 1-6.

Oelofsen, B.W. and Araujo, D.C. (1987). Mesosaurus tennidens and Stereosternum tumidum from the Permian Gondwana of Both Southern Africa and South America. *South African Journal of Science*, **83**, 370-372.

Ojeda, H.A.O. (1981). Estrutura, estratigrafia e evolução das bacias marginais brasileiras. *Revista Brasileira de Geociências*, **11**(4), 257-273.

Oliveira, A.G., Fuzikawa, K., Moura, L.A.M. and Raposo, C. (1985). Província uranífera de Lagoa Real, Bahia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. I, 105-120. DNPM, Brasília.

Oliveira, A.M. (1993). *Petrografia, estratigrafia, petroquímica e potencialidade para elementos do Grupo da Platina (EGP) do Complexo Barro Alto, na região de Goianésia, Goiás*. M.Sc. Thesis, IG/UnB, Brasília, 86 p. (unpublished).

Oliveira, C.G., Tazava, E., Tallarico, F., Santos, R.V. and Gomes, C. (1998). Gênese do depósito de Au-Cu-(U-ETR) de Igarapé Bahia, Província Mineral de Carajás. *Cong. Bras. Geol., 40, Belo Horizonte, SBG, Anais*, p. 137.

Oliveira, C.G. and Leonardos, O.H. (1990). Gold mineralization in the Diadema shear belt, northern Brazil. *Economic Geology*, **85**, 1034-1043.

Oliveira, C.G., Santos, R.V. and Leonardos, O.H. (1995). Geologia e mineralização aurífera do greenstone belt Sapucaia, sudeste do Pará. *Boletim do Museu Paraense Emílio Goeldi*, Série Ciências da Terra, 7, 61-91.

Oliveira, C.G., Sintia, A.V. and Barbosa, I.O. (1997). Influência da deformação transcorrente NS na mineralização aurífera na sequência vulcano-sedimentar de Mara Rosa. *An. 6º Simp. Geol. Centro-Oeste, Cuiabá, SBG*, 59-61.

Oliveira, S.M.B. and Trescases, J.J. (1980). Geoquímica da alteração supérgena das rochas ultramáficas de Santa Fé, Goiás. *Revista Brasileira de Geociências*, **10**(4), 243-257.

Oliveira, S.M.B. and Trescases, J.J. (1982). Estudo mineralógico e geoquímico da laterita niquelífera de Niquelândia, GO. *An. 32º Cong. Bras. Geologia, Salvador, Bahia, SBG*, **3**, 1183-1190.

Oliveira, T.F. (1998). As minas de Vazante e Morro Agudo, Minas Gerais. *Workshop: Depósitos Minerais Brasileiros de Metais-Base, Salvador, CAPES-PADCT-ADIMB*, 48-57.

Oliveira, V.P., Fragomeni, L.F.P. and Bandeira, C.A. (1997). Depósitos de magnesita de Serra das Éguas, Brumado, Bahia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVC, 219-234. DNPM, Brasília.

Olivo, G.R. (1989). *Controle litoestratigráfico e gênese das ocorrências auríferas da sequência psamo-pelito-carbonática do Grupo Paranoá, Goiás*. M.Sc. Thesis, IG/UnB, Brasília, 296 p. (unpublished).

Olivo, G.R. and Marini, O.J. (1988). Ouro no Grupo Paranoá. *Anais 35º Cong. Bras. Geol., Belém, SBG*, **1**, 93-106.

Olivo, G.R., Gauthier, M., Bardoux, M., Sá, E.L., Fonseca, J.T.F. and Carbonari, F. (1995). Palladium-bearing gold deposit hosted by Proterozoic Lake Superior-type-iron formation at the Cauê mine, Itabira District, southern São Francisco craton, Brazil: Geologic and structural controls. *Economic Geology*, **90**(1), 118-134.

Padula, V.T. and Porto Alegre, H.K. (1986). Xisto: Reservas e Recursos da Formação Irati. *An. 3º Cong. Brasileiro de Petróleo*, Curitiba, **1**, 1-8.

Paes de Barros, A.J., Laet, S.M. and Resende, W.M. (1999). Províncias auríferas do norte do Estado de Mato Grosso. In: *Simpósio de Geologia da Amazônia, 6, Boletim de Resumos Expandidos, Manaus, Sociedade Brasileira de Geologia*, 124-127.

Paes de Barros, A.J. (1994). *Contribuição à geologia e controle das mineralizações auríferas da região de Peixoto de Azevedo, MT*. M.Sc. Thesis, USP, 145 p.

Pamplona, R.I. and Nagao, M. (1981). Jazimentos de amianto crisotila da mina de Cana Brava, Goiás. In: *Principais Depósitos Minerais da Região Centro-Oeste*, DNPM, Goiânia, 76-137.

Parente, C.V. (1995). *Géologie et paléogeographie d´une plataforme à evaporites et magnésite d´âge protérozoïque (2Ga): le cadre géotectonique initial de la ceinture mobile Orós dans la région d´Alencar (Ceará, Brésil)*. Doctorate Thesis, Univ. Nantes, França, 306 p.

Parente, C.V. and Arthaud, M.H. (1995). O sistema Orós-Jaguaribe no Ceará, NE do Brasil. *Revista Brasileira de Geociências*, **25**(4), 297-306.

Parente, C.V. and Guillou, J.J. (1995). Geologia e paleogeografia dos depósitos de magnesita de idade proterozóica (2Ga) da região de Alencar (Ceará). *Simp. Geol. Nordeste, 16, Recife, SBG, Anais*, **2**, 428-432.

Pedrosa Soares, A.C., Noce, C.M., Vidal, P., Monteiro, R.L.B.P. and Leonardos, O.H. (1992). Toward a new tectonic model for the Late Proterozoic Araçuaí (SE Brazil) West Congolian (SW Africa) belt. *Journal of South American Earth Sciences*, **6**, 33-47.

Pedrosa Soares, A.C., Wiedemann, C.M., Fernandes, M.L.S., Faria, L.F. and Ferreira, J.C.H. (1999). Geotectonic significance of the neoproterozoic granitic magmatism in the Araçuaí belt, eastern Brazil: a model and pertinent questions. *Revista Brasileira de Geociências*, **29**(1), 59-66.

Pedrosa Soares, A.C., Vidal, P., Leonardos, O.H. and Brito Neves, B.B. (1998). Neoproterozoic oceanic remnants in eastern Brazil: further evidence and refutation of an exclusively ensialic evolution for the Araçuaí-West Congo orogen. *Geology*, **26**, 519-522.

Pedrosa-Soares, A.C., Dardenne, M.A., Hasui, Y., Castro, F.D.C. and Carvalho, M.V.A. (1994). *Nota explicativa dos mapas geológico, metalogenético e de ocorrências minerais do Estado de Minas Gerais, 1/1000.000*. COMIG, Belo Horizonte, 97 p.

Pereira, M.C., Galbiatti, H.F. and Fonseca, M.A. (1999). Mineralização aurífera (Jacutinga) associada a fraturas em zonas transcorrentes, Mina Conceição, Itabira, MG. *Simp. Geol. Centro-Oeste, 7, Brasília, SBG, Bol. Resumos*, p. 75.

Pimentel, M.M. and Fuck, R.A. (1991). Origin of orthogneiss and metavolcanic rock units in western Goiás: Neoproterozoic crustal accretion. *Geochimica Brasiliensis*, **5**, 133-152.

Pimentel, M.M. and Fuck, R.A. (1992). Neoproterozoic crustal accretion in Central Brazil. *Geology*, **20**(4), 375-379.

Pimentel, M.M. and Machado, N. (1984). Geocronologia U-Pb dos terrenos granito greenstone de Rio Maria, Pará. In: *Cong. Bras. Geol., 38, Camboriú, 1984. Boletim de Resumos... Camboriú, SBG*, **2**, 390-391.

Pimentel, M.M., Fuck, R.A. and Botelho, N.F. (1999). Granites and the geodynamic history of the neoproterozoic Brasília belt, Central Brazil: a review. *Lithos*, **46**, 463-483.

Pimentel, M.M., Heaman, L., Fuck, R.A. and Marini, O.J. (1991). U-Pb zircon geochronology of Precambrian tin-bearing continental-type acid magmatism in central Brazil. *Precambrian Research*, **52**, 321-335.

Pimentel, M.M., Machado, M. and Lobato, L.M. (1994). U-Pb geochronology of the Lagoa Real uranium district, Brazil: Implications for the age of the uranium mineralization. 8 p.

(unpublished report).

Pinheiro, R.V.L. and Holdsworth, R.E. (1997). Reactivation of Archean strike-slip fault system, Amazon region, Brazil. *Journal of the Geological Society of London*, **154**, 99-103.

Pinho, F.E.C. (1996). *The origin of Cabaçal Cu-Au deposit, Alto Jauru Greenstone Belt, Brazil*. Ph.D. Thesis, Univ. Western Ontario, 211 p.

Pinho, F.E.C., Fyfe, W.S. and Pinho, M.A.S.B. (1997). Early Proterozoic evolution of the Alto Jauru Greenstone Belt, Southern Amazonian Craton, Brazil. *International Geology Review*, **39**, 220-229.

Pinho, J.M.M. (1990). *Evolução tectônica da mineralização de zinco de Vazante*. M.Sc. Thesis, IG/UnB, Brasília, 115 p. (unpublished).

Pinho, M.A.S.B., Lima, E.F. and Chemale, Jr. F. (1999). Geologia da região do Moriru: dados preliminares da Formação Iriri, Aripuanã, Mato Grosso. *Simp. Geol. Centro-Oeste, 7, Brasília, SBG, Bol. Resumos*, p. 14.

Pires, F.R.M. (1977). *Geologia do distrito manganesífero de Conselheiro Lafaiete, MG*. M.Sc. Thesis, UFRJ, Rio de Janeiro.

Pires, F.R.M. (1983). Manganese mineral parageneses at the Lafaiete District, MG, Brazil. *Anais da Academia Brasileira de Ciências*, 55(3), 272-285.

Portocarrero, J.L.T. (1996). *Geologia da jazida aurífera Mina Nova, greenstone belt de Crixás, Goiás, Brasília*. M.Sc. Thesis, UnB, Brasília, 102 p.

Pulz, G.M. (1990). *Geologia do depósito aurífero tipo Maria Lázara (Guarinos, Goiás)*. M.Sc. Thesis, UnB, Brasília, 190 p.

Pulz, G.M. (1995). *Modelos prospectivos para ouro em greenstone belts: exemplo dos depósitos Maria Lázara e Ogó na região de Guarinos e Pilar de Goiás, Goiás*. Doctorate Thesis, UnB, Brasília, 190 p.

Pulz, G.M., Jost, H., Michel, D. and Giuliani, G. (1991). The Archean Maria Lázara gold deposit, Goiás, Brazil: example of Au-Bi-Te-S metallogeny related to shear zones intruded by synkinematic granitoids. In: *Brazil Gold'91*, ed. Ladeira, E.A., 385-387.

Queiroz, C.L., Jost, H. and Mc Naughton, N.J. (1999). U-Pb Shrimp ages of Crixás granite-greenstone belt terranes: from Archean to Neoproterozoic. *Anais 7° Simp. Nac. Est. Tect., Lençóis-Bahia, SBG-ABGP*, Sessão 1, p. 35-37.

Quemeneur, J. and Lagache, M. (1999). Comparative study of two pegmatitic fields from Minas Gerais, using the Rb and Cs contents of micas and feldspars. *Revista Brasileira de Geociências*, **29**(1), 27-32.

Ramos, V.A. (1988). Late Proterozoic-Early Paleozoic of South América – a Collisional History. *Episodes*, **11**(3), 168-174.

Ramos, V.A. (1988). The tectonics of the central Andes: 30°S latitude. *Geological Society of America, Special Paper*, **218**, 31-54.

Ramos, V.A. (1999). Plate tectonic setting of the Andean Cordillera. *Episodes*, **22**(3), 183-190.

Raposo, C., Matos, E.C. and Brito, W. (1984). Zoneamento Cálcio-sódico nas rochas da província uranífera de Lagoa Real. *Cong. Bras. Geol., 33, Rio de Janeiro, SBG, Anais*, p. 1489-1502.

Reinhardt, M.C. and Davison, I. (1990). Structural and lithologic controls on gold deposition in the shear hosted Fazenda Brasileiro mine, Bahia State, Northeast Brazil. *Economic Geology*, **85**, 952-967.

Reis, L.B. (1999). *Estudo de mineralização de grafita no extremo nordeste de Minas Gerais*. M.Sc. Thesis, UFMG, Belo Horizonte, 87 p.

Reis, N.J. and Fraga, L.M.B. (1996). Vulcanismo Surumu: caracterização de seu comportamento químico à luz de novos dados. In: *Congresso Brasileiro de Geologia, 39, Salvador, Anais, Sociedade Brasileira de Geologia*, **2**, 88-91.

Reischel, J.L. (1980). Mineralizações auríferas associadas ao Complexo Granítico de Lavras do Sul, RS. *Cong. Bras. Geol., 31, Camboriú, SBG, Anais*, **3**, 1700-1712.

Remus, M.V.D., Hartmann, L.A, McNaughton, M.J., Groves, D.I., Reischl, J.L., Dorneles, H.T. 1999. The Camaquã Cu (Au, Ag) and Santa Maria Pb-Zn (Cu-Ag) mines of Rio Grande do Sul, Southern Brazil. In: *Base Metal deposits of Brazil*, eds. Silva, M.G. and Misi, A., pp. 54-63. MME/CPRM/DNPM, Belo Horizonte.

Renger, F.E., Noce, C.M. and Romano, A.W. (1994). Evolução sedimentar do Supergrupo Minas (500Ma de registro geológico), Quadrilátero Ferrífero, Minas Gerais, Brasil. *Geonomos*, **2**, 1-11.

Renger, F.E., Silva, R.M.P. da and Suckam, V.E. (1988). Ouro nos conglomerados da Formação Moeda, Sinclinal de Gandarela, Quadrilátero Ferrífero, Minas Gerais. *35° Cong. Bras. Geologia, Belém, Pará, Anais*, 1, 4-57.

Réquia, R.C.M., Xavier, R.P. and Figueiredo, B.R. (1997). Evolução paragenética, textural e das fases fluidas no depósito polimetálico de Salobo, Província Mineral de Carajás, Pará. *Boletim do Museu Paraense Emílio Goeldi*, Série Ciências da Terra, 7, 27-39.

Ribeiro, C.C. (1984). *Caractérisation Sédimentologique et Géochimique d'un Milieu Sédimentaire. Cas du Dévonien Moyen e Supérieur de la Région de Paraiso do Norte-Miranorte (Bassin de Maranhão-Goiás-Brésil)*. Doctorate Thesis 3°Cycle, Nancy, 230 p.

Ribeiro, C.C. and Dardenne, M.A. (1978). O Minério de Ferro da Formação Pimenteiras na Borda Sul da Bacia do Maranhão, Goiás. *Anais 30° Cong. Bras. Geologia, Recife, SBG*, **4**, 1583-1595.

Ribeiro, R.K. (1998). *Mineralogia, geoquímica e gênese das ocorrências auríferas no flanco norte do anticlinal de Mariana, Quadrilátero Ferrífero: uma nova tipologia de minério denominada Bugre*. M.Sc. Thesis, UnB, Brasília, 115 p.

Ribeiro-Rodrigues, L.C. (1998). *Gold in archean banded iron-formation of the Quadrilátero Ferrífero, Minas Gerais, Brazil. The Cuiabá Mine*. Ph.D. Thesis, Aachen University, 264 p.

Ribeiro-Rodrigues, L.C.R., Friedrich, G., Oliveira, C.G., Vieira, F.W.R., Biasi, E.E., Callegari, L.A. (1996). The BIF hosted Cuiabá gold deposit, Iron Quadrangle, Minas Gerais, Brazil: characteristics, control and genesis. *Cong. Bras. Geol., 39, Salvador, SBG, Anais*, 7, 224-228.

Richardson, S.V., Kesler, S.I. and Essene, E.J. (1986). Origin and geochemistry of the Chapada Cu-Au deposit, Goiás, Brazil: A metamorphosed wall-rock porphyry copper deposit. *Economic Geology*, **81**, 1884-1898.

Rigobello, A.E., Branquinho, J.A., Dantas, M.G.S., Oliveira, T.F. and Neves Filho, W. (1988). Mina de zinco de Vazante. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 101-110. DNPM, Brasília.

Rios, F.J., Villas, R.N., Fuzikawa, K., Sial, A.N. and Mariano, G. (1998). Isótopos de oxigênio e temperatura de formação dos veios mineralizados com wolframita da jazida Pedra Preta, sul do Pará. *Revista Brasileira de Geociências*, **28**(3), 253-256.

Robert, F. (1996). *Tapajós Gold Project, Pará State, Brazil. Canada-Brazil Cooperation Project for Sustainable Development in the Mineral Sector* (CIDA Project 204/13886), Ottawa, 35 p. (unpublished report).

Roberto, F.A.C. and Souza, V.C. (1991). Depósitos de opala de Pedro II, Piauí. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVA, 337-346. DNPM/CPRM, Brasília.

Rocha, G.M.F. (1985). *Características das fácies ferríferas de Boquira, encaixante da mineralização de chumbo-zinco*. M.Sc. Thesis, UFBA, Salvador, 96 p.

Rodrigues, O.B., Kosuki, R. and Coelho Filho, A. (1986). Distrito manganesífero de Serra do Navio, Amapá. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S., vol. II, p. 167-175. DNPM, Brasília.

Roeser, H., Roeser, U.G., Grossi, A.W.R. and Flores, J.C.C. (1984). Contribuição à Origem das Jazidas de Bauxita de Cataguases, MG. *An. 33° Cong. Bras. Geol., Rio de Janeiro, SBG*, **8**, 3853-3865.

Romagna, G. and Costa, R.R. (1988). Jazida de zinco e chumbo de Morro Agudo, Paracatu, Minas Gerais. In: *Principais Depósitos Minerais* *do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. 3, p. 83-90. DNPM, Brasília.

Romano, A.W. (1989). *Evolution tectonique de la région nord-ouest du Quadrilatère Ferrifère, Minas Gerais, Brésil.* Ph.D. Thesis, University of Nancy, 259 p.

Romano, A.W., Bertrand, J.M., Michard, A. and Zimmermann, J.L. (1991). Tectonique tangentielle et décrochements d'âge Protérozoïque Inférieur (orogenèse transamazonienne, environ 2000Ma) au nord du Quadrilatère Ferrifère, (Minas Gerais, Brésil). *Comptes Rendus de l'Académie des Sciences, Paris*, **313**, 1195-1200.

Ronchi, L.H., Lindenmayer, Z.G., Bastos Neto, A. and Murta, C.R. (2000). O "stockwork" e a zonação do minério sulfetado no arenito inferior de Mina Uruguai-RS. In: *As Minas de Camaquã: um estudo multidisciplinar*, coords. Ronchi, L.H. and Lobato, A.O.C., pp. 165-190. Ed. Unisinos.

Ronchi, L.H., Touray, J.C., Michard, A. and Dardenne, M.A. (1993). The Ribeira fluorite district southern Brazil: Geological and geochemical (REE, Sm-Nd isotopes) characteristic. *Mineralium Deposita*, **28**, 240-252.

Rosa, D.B. (1988). *Les gisements d'opales nobles de la région de Pedro II dans l'état de Piauí (région nord-est du Brésil).* Doctorate Thesis, INPL, Nancy, França, 209 p. (Thése de Doctorat, Institut Nacional Polytechnique de Lorraine).

Sá, J.H.S, Inda, H.A.V., Mascarenhas, J.F. and Brito Neves, B.B. (1982). *Mapa Metalogenético da Bahia, Escala 1:1.000.000.* Secretaria das Minas e Energia, Coordenação da Produção Mineral, Salvador.

Sad, A.R., Campolino, A., Costa, A. Maia de, Lima, F.R.T. and Carvalho, R.S. (1997). Depósito de Potássio de Fazendinha, Nova Olinda do Norte, Amazonas- In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVC, parte C, 257-276. DNPM-CPRM.

Sad, A.R., Lima, F.R.T., Wolf, F., Soares, J.M. and Carvalho, R.S. (1982). Depósito Potassífero da Fazendinha-Bacia do Médio Amazonas. *An. 32° Cong. Bras. Geol., Salvador, Bahia, SBG*, **3**, 1086-1099.

Sães, G.S. (1999). *Evolução tectônica e paleogeográfica do aulacógeno Aguapeí (1,2-1,0Ga) e dos terrenos do seu embasamento na porção sul do Cráton Amazônico.* Doctorate Thesis, USP, São Paulo, 135 p.

Sães, G.S. and Fragoso Cesar, A.R.S. (1994). The Aguapeí basin: a Grenville age aulacogen of the Sunsas orogen. *Cong. Bras. Geol., 38, Camboriú, SBG, Anais*, 207-209.

Sães, G.S., Pinho, F.E.C. and Leite, J.A.D. (1991). Coberturas metassedimentares do Proterozóico Médio no sul do Cráton Amazônico e suas mineralizações auríferas. *Simp. Geol. Centro-Oeste, 3, Cuiabá, SBG, Anais*, 37-47.

Sales, M.A.S. (1998). *The geological setting of the Lamego banded-iron-formatio-hosted gold deposit, Quadrilátero Ferrífero District, Minas Gerais, Brazil.* Ph.D. Thesis, University of Queen, Canada, 183 p.

Salim, J. (1993). *Géologie, petrologie et geochimie des skarns à scheelite de la mine de Brejuí, Currais Novos, région du Seridó, NE du Brésil.* Doctorate Thesis, Univ. Louvain, 272p.

Samama, J., Meyer, R., Bartoli, F. and Moura, F. (1983). Caractérisation chimico-minéralogique de l'opale noble des gisements du nordest du Brésil. *Comptes Rendus de l'Académie des Sciences, Paris*, **296**, 625-630.

Sampaio, D.R., Costa, E.D.A. and Araújo Neto, M.C. (1994). Diamantes e carbonados do alto Paraguaçu: geologia e potencialidade econômica. Companhia Bahiana de Pesquisa Mineral, *Série Arquivos Abertos 8*, Salvador, 23 p.

Santos, M.D. (1995). *O papel dos granitóides na gênese dos depósitos de ouro tipo lode arqueano: caso da jazida de Cumaru – PA.* Doctorate Thesis, IG/UnB, Brasília, 157 p.

Santos, M.D. and Leonardos, O.H. (1995). Sistema de fluidos e modelo genético do depósito aurífero do Cumaru, SE do Estado do Pará. *Boletim do Museu Paraense Emílio Goeldi*, Série Ciências da Terra, 7, 237-262.

Santos, M.D., Leonardos, O.H., Foster, R.P. and Fallick, A.E. (1998). The lode-porphyry model as deduced from the Cumaru mesothermal granitoid-hosted gold deposit, southern Pará, Brazil. *Revista Brasileira de Geociências*, **28**(3), 327-338.

Santos, O.M., Vitorasso, E.C.L., Silva, R.M., Guerra, H.R.M., Chaves, J.J., Mantovani, T.J., Silva, R.A., Kalil Jr., A.R., Santos, V.A.M., Navarro, L.A.G. and Pena, L.S.T. (1988). Mina de ouro de Fazenda Brasileiro, Bahia. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 431-444. DNPM, Brasília.

Santos, B.A. (1983). *Amazônia - Potencial Mineral e Perspectivas de Desenvolvimento*, ed. Queiroz, T.A., São Paulo, 256 p.

Santos, E. L., Maciel, L.A.C. and Zir Filho, J.A. (1998). *Distritos mineiros do Rio Grande do Sul.* Programa Nacional de Distritos Mineiros, Departamento Nacional de Produção Mineral, Porto Alegre, 35 p.

Santos, J.O.S., Reis, N.J., Hartmann, L.A., McNaughton, N. and Fletcher, I. (1999). Associação anortosito-charnockito-rapakivi no Calimiliano do norte do Cráton Amazônico, Estado de Roraima, Brasil: evidências da geocronologia U-Pb (shrimp) em zircão e baddeleyita. In: *Simpósio de Geologia da Amazônia, 6. Manaus, Boletim de Resumos Expandidos, Sociedade Brasileira de Geologia*, 502-505.

Savi, C.N. and Dardenne, M.A. (1980). Zonação, paragênese e controles da mineralização da fluorita do filão 2ª Linha Torrens, município do Morro da Fumaça Santa Catarina. *Cong. Bras. Geol., 31, Camboriú, SBG, Anais*, vol. 3, 1743-1757.

Scarpelli, W. (1991). Aspects of gold mineralization in the Iron Quadrangle, Brazil. In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 151-157. Rotterdam, A.A. Balkema.

Scarpelli, W. (1991). Precambrian auriferous quartz-pebble conglomerates in Brazil. In: *Gisements alluviaux d'or*, pp. 261-273. La Paz.

Scarpelli, W. (1966). Aspectos genéticos e metamórficos das rochas do Distrito de Serra do Navio. *Avulso, Divisão de Geologia e Mineralogia*, **41**, 37-55.

Schmitt, J.C.C., Camatti, C. and Barcellos, R.C. (1991). Depósitos de amestista e ágata no estado do Rio Grande do Sul. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVA, p. 271-285. DNPM/CPRM.

Schneider, R.L., Mullmann, H., Tommasi, E., Medeiros, R.A., Daemon, R.S.F. and Nogueira, A.A. (1974). Revisão Estratigráfica da Bacia do Paraná. *An 28° Cong. Bras. Geol., SBG*, Porto Alegre, vol. 1, 41-66.

Schobbenhaus, C. and Brito Neves, B.B. (1996). Geological map of South America (CGMW): The South America Platform. *Intern. Geol. Cong., 30, Beijing, Abstracts*, p. 509.

Schobbenhaus, C. and Campos, D.A. (1984). A evolução da Plataforma Sul-Americana no Brasil e suas principais concentrações minerais. In: *Geologia do Brasil*, coords. Schobbenhaus, C., Campos, D.A., Derze, G.R. and Asmus, H.E., pp. 9-55. MME/DNPM, Brasília.

Schobbenhaus, C. (1996). As trafogêneses superpostas Espinhaço e Santo Onofre, Estado da Bahia: revisão e novas propostas. *Revista Brasileira de Geociências*, **26**(4), 265-276.

Schobbenhaus, C. (1984). Distribution of Mineral Deposits through Geologic Time in Brazil. In: *International Geologic Congress, 27, Moscow, Abstracts*, **6**(12), 289.

Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., eds. (1997). *Principais Depósitos Minerais do Brasil, Rochas e Minerais Industriais.* Departamento Nacional de Produção Mineral/ Companhia de Pesquisa de Recursos Minerais, Brasília, vol. IV-C, 634 p.

Schulmann, A., Carvalho, A. and Valeton, I. (1997). Bauxite of Poços de Caldas. In: *Brazilian Bauxites*, eds. Carvalho, A., Boulangé, B., Melfi, A.J. and Lucas, Y., pp. 229-254. USP/FAPESP/ORSTOM, São Paulo.

Seer, H.J. (1985). *Geologia, deformação e mineralização de cobre no complexo vulcano-sedimentar de Bom Jardim de Goiás.* M.Sc. Thesis, IG/UnB, Brasília, 181 p. (unpublished).

Sidder, G.B. and Mendoza, V. (1995). Geology of the Venezuelan Guayana Shield and its relation to the Geology of the entire Guayana Shield. In: Geology and Mineral Deposits of the Venezuelan Guayana Shield, United States Geological Survey. *United States Geological Survey Bulletin*, **2124**, p. B1-B41.

Sidder, G.B. (1995). Mineral Deposits of the Venezuelan Guayana Shield. In: Geology and Mineral Deposits of the Venezuelan Guayana Shield, Washington. *United States Geological Survey Bulletin*, 2124, 01-020.

Sillitoe, R.H. (1991). Intrusion-related gold deposits. In: *Gold Metallogeny and Exploration*, ed. Foster, R.P., pp. 165-209. Glasgow, Blackie and Son Ltd.

Silva, A.R.B. and Cordeiro, A.A.C. (1988). Depósitos de ouro da Serra das Andorinhas, Rio Maria, Pará. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and Coelho, C.E.S., vol. III, p. 559-567. DNPM, Brasília.

Silva, C.M.G. and Villas, R.N. (1998). The Águas Claras Cu-sulfide ± Au deposit, Carajás region, Pará, Brazil: geological setting, wall-rock alteration and mineralizing fluids. *Revista Brasileira de Geociências*, **28**(3), 315-326.

Silva, C.R. and Rizzoto, G.J. (1994). Província aurífera Guaporé. *Cong. Bras. Geol., 38, Camboriú, SBG, Anais*, **1**, 323-325.

Silva, M.A.M. (1988). Evaporitos do Cretáceo da Bacia do Araripe: Ambiente de Deposição e História Diagenética. *Boletim de Geociências da Petrobras*, **2**(1), 53-63.

Silva, M.G. and Cunha, J.C. (1999). Greenstone belts and equivalent volcano-sedimentary sequences of the São Francisco Craton, Bahia, Brazil: Geology and mineral potential. In: *Base Metal Deposits of Brazil*, eds. Silva, M.G. and Misi, A., pp. 92-99. MME/CPRM/DNPM, Belo Horizonte.

Silva, M.R.R. and Dantas, J.R.A. (1997). Província pegmatítica da Borborema-Seridó, Paraíba e Rio Grande do Norte. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C., Queiroz, E.T. and Coelho, C.E.S., vol. IVB, 441-467. DNPM, Brasília.

Silva, W.L. and Legrand, J.M. (1996). Termobarometria no depósitos aurífero São Francisco (RN): Evolução metamórfica e caracterização P-T da mineralização. In: *Congresso Brasileiro de Geologia, 39, Salvador, Anais, Sociedade Brasileira de Geologia*, **3**, 346-349.

Siqueira, J.B. (1996). *Aspectos lito-estruturais e controle das mineralizações do depósito do Salobo 3A (Serra dos Carajás/PA)*. Ph.D. Thesis, UFPA, Belém.

Siqueira, A.J.B. (1997). *Geologia da mina de ouro do Filão do Paraíba, região de Peixoto de Azevedo, norte de Mato Grosso.* M.Sc. Thesis, Universidade Federal do Rio de Janeiro, 98 p.

Soares, A.D.V., Ronzê, P.C., Santos, M.G.S., Leal, E.D.L. and Barreira, C.F. (1999). Geologia e mineralizações do depósito Cu-Au Alemão, Província Mineral de Carajás (PA). *Simp. Geol. Amazônia, 6, Manaus, SBG, Anais*, 144-147.

Soares, A.D.V., Santos, A.B., Vieira, E.A., Bella, V.M. and Martins, L.P.B. (1994). Área Águas Claras: contexto geológico e mineralizações. *Simp. Geol. Amazônia, 4, Belém, SBG, Anais*, 379-382.

Souza, J.O., Moreton, L.C. and Camargo, M.A. (1995). Geologia das seqüências metavulcanossedimentares da Serra do Tapa (SE do Pará) e Xambioá (NW do Tocantins): ocorrências de metabasaltos com pillow lavas. *Boletim de Geociências Centro-Oeste*, **18**(1/2), 20-31.

Souza Neto, J.A., Legrand, J.M., Sommet, Ph. and Melo Jr., G. (1996). Metassomatic alteration styles applied to exploration for gold-bearing calc-silicate rocks in the Borborema Province, Northeast Brazil. *Cong. Bras. Geol., 39, Salvador, SBG, Anais*, p. 244-247.

Souza, Z.S., Medeiros, H., Althoff, F.J. and Dall'Agnol, R. (1990). Geologia do terreno granito-"greenstone" arqueano da região de Rio Maria, sudeste do Pará. In: *Cong. Bras. Geol., 36, Natal, 1990. Anais... Natal, SBG*, **4**, 2913-2928.

Souza, Z.S., Potrel, A., Medeiros, H., Lafon, J.M., Dall'Agnol, R., Althoff, F.J., Oliveira, C.G. and Pimentel, M.M. (2000). Nd, Pb and Sr isotopes of Archean greenstone belt of the Rio Maria region and their implications for the geodynamic evolution of the Southern Amazonian Craton. *Revista Brasileira de Geociências* (In press).

Souza, N.B. (1991). Depósitos diamantíferos de Poxoréo, Mato Grosso. In: *Principais Depósitos Minerais do Brasil, Gemas e Rochas Ornamentais*, eds. Schobbenhaus, C. E.T. de Queiroz and C.E.S. Coelho, vol. IVA, p. 149-154. DNPM/CPRM, Brasília.

Souza, N.B. (1988). Principais depósitos de ouro do Estado de Mato Grosso. In: *Congresso Brasileiro de Geologia, 35, Belém, Anais, Sociedade Brasileira de Geologia*, **1**, 116-129.

Spier, C.A. and Ferreira Filho, C.F. (1999). Geologia, estratigrafia e depósitos minerais do Projeto Vila Nova, Escudo das Guianas, Amapá, Brazil. *Revista Brasileira de Geociências*, **29**(2), 173-178.

Suguio, K., Svisero, D.P. and Felitti, W. (1979). Conglomerados polimícticos diamantiníferos de idade cretácea de Romaria (MG): um exemplo de sedimentação de leques aluviais. In: *Simpósio Regional De Geologia De Minas Gerais, 2, Belo Horizonte, 1979. Anais...* Belo Horizonte, SBG, 217-229.

Suita, M.T.F. (1996). *Geoquímica e metalogenia de elementos do grupo da platina EPG + Au) em complexos máfico-ultramáficos do Brasil: critérios e guias com ênfase no complexo máfico-ultramáfico acamadado de alto grau de Barro Alto (CBA, Goiás).* Doctorate Thesis, UFRGS, Porto Alegre, 525 p. (unpublished).

Suita, M.T.F. (1998). Late Paleo-Neoproterozoic PGE + Au metallogeny of giant layered high-grade mafic ultramafic intrusions (Barro Alto and Niquelândia), Tocantins Province, Goiás, Central Brazil. *Abstr. 14th Int. Conf. on Basement Tectonics, Ouro Preto, MG-Brazil*, 179-180.

Suita, M.T.F. and Nilson, A.A. (1991). O depósito de cromita estratiforme do Complexo de Luanga, Província Carajás, Pará. Aspectos Geoquímicos. *Cong. Bras. Geoq., 3, São Paulo, SBGq, vol. res.*, 203-206.

Szatmari, P., Carvalho, R.S. and Simões, I.A. (1979). A Comparision of Evaporite Facies in the Late Paleozoic Amazon and the Middle Cretaceous South Atlantic Salt Basins. *Economic Geology*, **74**(2), 432-447.

Szubert, E.C., Orlandi Filho, V. and Shintaku, I. (1978). Geologia dos jazimentos de ametista do Alto Uruguai, RS. In: *Congresso Brasileiro de Geologia, 30, Recife, SBG, Anais*, 1833-1892.

Tallarico, F.H.B., Coimbra, C.R. and Costa, C.H.C. (2000a). The Serra Leste sediment-hosted Au-(Pd-Pt) mineralization, Carajás Province, Brazil. *Revista Brasileira de Geociências* (in press).

Tallarico, F.H.B., Oliveira, C.G. and Figueiredo, B.R. (2000b). The Igarapé Bahia primary Cu-Au mineralization, Carajás Province, Brazil: A descriptive model and genetic considerations. *Revista Brasileira de Geociências* (in press).

Tallarico, F.H.B., Rego, J.L. and Oliveira, C.G. (1998). A mineralização de Au-Cu de Igarapé Bahia, Carajás: um depósito da classe óxido de Fe (Cu-U-Au-ETR). *Cong. Bras. Geol., 40, Belo Horizonte, SBG, Anais*, p. 116.

Tassinari, C.C.G. (1996). *O mapa geocronológico do Cráton do Amazônico no Brasil. Revisão dos dados isotópicos.* IG/USP, São Paulo, Livre Docência Thesis, 139 p.

Tassinari, C.C.G. and Mellito, K.M. (1994). The time-bound characteristics of gold deposits in Brazil and their tectonic implications. *Comunicaciones*, **45**, 45-54.

Tassinari, C.C.G., Teixeira, W. and Siga Jr., O. (1978). Considerações cronoestratigráficas da região da Chapada do Cachimbo e Dardanelos. In: *Congresso Brasileiro de Geologia, 30, Recife,* *Anais, Sociedade Brasileira de Geologia*, 1, 477-490.

Tassinari, C.G. and Macambira, M.J.B. (1999). Geochronological provinces of the Amazonian Craton. *Episodes*, **22**(3), 174-182.

Teixeira, J.B. (1994). *Geochemistry, petrology and tectonic setting of Archean basaltic and dioritic rocks from the N4 iron deposit, Serra dos Carajás, Pará, Brazil.* Ph.D. Thesis, University of Pennsylvania, 175 p.

Teixeira, J.B.G. and Eggler, D.H. (1994). Petrology, geochemistry and tectonic setting of Archean basaltic and dioritic rocks from the N4 iron deposit, Serra dos Carajás, Pará, Brazil. *Acta Geológica Leopoldênsia*, **40**, 71-114.

Teixeira, J.B.G., Kishida, A., Marimon, M.P.C., Xavier, R.P. and McReath, I. (1990). The Fazenda Brasileiro gold deposit, Bahia: Geology, hydrothermal alteration and fluid inclusion studies. *Economic Geology*, **85**, 990-1009.

Teixeira, N. (1996). Assoalho oceânico no complexo ultramáfico do Quatipuru (PA): implicações geotectônicas para a faixa móvel Araguaia. *Cong. Bras. Geol., 39, Salvador, SBG, Anais*, **6**, 117-120.

Teixeira, N. (1998). Geologia, petrologia e implicações prospectivas da província kimberlítica de Juína (MT). Exame de Qualificação, UnB, 84p.

Teixeira, N.A., Gaspar, J.C., Brenner, T.L., Cheney, J.T. and Marchetto, C.M.L. (1987). Geologia e implicações geotectônicas do *greenstone belt* do Morro do Ferro (Fortaleza de Minas-MG). *Revista Brasileira de Geociências*, **17**, 209-220.

Teixeira, W. (1985). *A evolução geotectônica da porção meridional do Cráton do São Francisco, com base em interpretações geocronológicas.* Doctorate Thesis, USP, São Paulo, 207 p.

Teixeira, W. and Gonzalez, M. (1988). Minas de Camaquã, município de Caçapava do sul, RS. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 33-41. DNPM, Brasília.

Teixeira, W., Tassinari, C.C.G., Cordani, U.G. and Kawashita, K. (1989). A review of the geochronology of the Amazonian Craton: Tectonic implications. *Precambrian Research*, **42**(1989), 213-227.

Thomson, M.L. and Fyfe, W.S. (1990). The Crixás gold deposit, Brazil: Thrust-related postpeak metamorphic gold mineralization of possible Brasíliano Cycle age. *Economic Geology*, **85**, 928-942.

Thorpe, R.I., Cummings, G.L. and Krstic, D. (1984). Lead isotope evidence regarding the age of gold deposits in the Nova Lima District, Minas Gerais. *Revista Brasileira de Geociências*, **14**(3), 147-152.

Tinoco, I.M. (1971). Contribuição ao conhecimento da gênese do fosfato de Olinda (Estado de Pernambuco). *Arquivo do Museu Nacional*, 54, 177-182.

Tolbert, G.E., Tremaine, J.W., Melcher, G.C. and Gomes, C.B. (1971). The recently discovered Serra dos Carajás iron deposit, northern Brazil. *Economic Geology*, **66**, 985-994.

Tompkins, L.A. and Gonzaga, G.M. (1989). Diamonds in Brazil and proposed model for the origin and distribution of diamonds in the Coromandel region, Minas Gerais, Brazil. *Economic Geology*, **84**, 591-602.

Trompette, R. (1994). *Geology of western Gondwana (2000-500Ma).* Amsterdam, Balkema, 350 p.

Trompette, R., Alvarenga, C.J.S. and Walde, D.H.G. (1998). Geological evolution of the Neoproterozoic Corumbá graben system (Brazil): Depositional context of the stratified Fe and Mn ores of the Jacadigo Group. *Journal of South American Earth Sciences*, 11(6), 587-597.

Truckenbrodt, W. and Kotschoubey, B. (1981). Argila de Belterra, Cobertura Terciária das Bauxitas Amazônicas. *Revista Brasileira de Geociências*, 11(3), 203-208.

Turpin, L., Maruejol, P. and Cuney, M. (1988). U-Pb, Rb-Sr and Sm-Nd chronology of granitic basement, hydrothermal albitites and uranium mineralization (Lagoa Real, South Bahia, Brazil). *Contributions to Mineralogy and Petrology*, **98**, 139-147.

Uhlein, A. (1991). *Transição cráton-faixa dobrada: um exemplo do Cráton São Francisco e da Faixa Aracuaí (Ciclo Brasiliano) no estado de Minas Gerais.* Doctorate Thesis, USP, São Paulo, 295 p.

Uhlein, A., Chaves, M.S.C. and Dossin, I.A. (1986). Recursos minerais da Serra do espinhaço Meridional (MG): uma síntese baseada no contexto litoestratigráfico regional. *Cong. Bras. Geol., 34, Goiânia, SBG, Anais*, 5, 2453-2464.

Uhlein, A., Trompette, R.R. and Alvarenga, C.J.S. (1999). Neoproterozoic glacial and gravitational sedimentation on a continental rifted margin: the Jequitaí-Macaúbas sequence (Minas Gerais, Brazil). *Journal of South American Earth Sciences*, **12**, 435-451.

Ulbrich, H.H.G.J. and Gomes, C.B. (1981). Alkaline rocks from continental Brazil. A review. *Earth Sciences Review*, 17(1-2), 135-154.

Valarelli, J.V., Bernadelli, A.L. and Beisiegel, R.W. (1978). Aspectos genéticos do minério de manganês do Azul. *Congr. Bras. Geol., 30, Recife, SBG, Anais*, 4, 1670-1679.

Valeton, I., Beissner, H. and Carvalho, A. (1991). The Tertiary Bauxite Belt on Tectonic Uplift Areas in the Serra da Mantiqueira, South-East Brazil. *Contributions to Sedimentology*, **17**, 1-101.

Valeton, I. and Melfi, A.J. (1988). Distribution Pattern of Bauxites in the Cataguases Area (SE Brazil), in Relation to Lower Tertiary Paleogeography and Younger Tectonics. *Science Géologique Bulletin*, **41**(1), 85-98.

Van Schmus, W.R., Brito Neves, B.B., Hackspacher, P.C., Babinsky, M., Fetter, A. and Dantas, E.L. (1995). Neoproterozoic and late mesoproterozoic sedimentary and volcanic sequences in the Borborema Province, NE Brazil. *Simp. Geol. Nordeste, 16, Recife, SBG, Atas*, **14**, 391-393.

Veiga, A.T.C. (1988). Mina de ouro de Novo Planeta, Alta Floresta, Mato Grosso. In: *Principais Depósitos Minerais do Brasil*, eds. Schobbenhaus, C. and C.E.S, Coelho, vol. III, 569-574. DNPM, Brasília.

Veiga, A.T.C. (1988). Geologia da província aurífera do Cassiporé, Amapá. *Simp. Geol. Amazônia, 2, Belém, SBG, Anais*, 3, 135-146.

Veiga, A.T.C., Dardenne, M.A. and Salomão, E.P. (1988). Geologia dos aluviões auríferos e estaníferos da Amazônia. *Cong. Bras. Geol., 35, Belém, SBG, Anais*, 266-270.

Veigel, R. and Dardenne, M.A. (1990). Paragênese e sucessão mineral nas diferentes etapas de evolução da mineralização Cu-Pb-Zn do Distrito Camaquã. RS. *Revista Brasileira de Geociências*, **20**, 55-67.

Vial, D.S. (1988). Mina de ouro de Passagem, Mariana, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 421-430. DNPM, Brasília.

Vieira, E.A.P., Saueressig, R., Siqueira, J.B., Silva, E.R.P., Rêgo, J.L. and Castro, F.D.C. (1988). Caracterização geológica da jazida polimetálica do Salobo 3A - Reavaliação. In: *Cong. Bras. Geol., 35, Belém, 1988. Anexo aos Anais... Belém, SBG*, 97-111.

Vieira, F.W.R. (1987a). Novo contexto geológico para a mina de ouro de Raposos. *Simp. Geol. Minas Gerais, 4, Belo Horizonte, SBG, Anais*, 7, 347-357.

Vieira, F.W.R. (1987b). Gênese das mineralizações auríferas da mina de Raposos. *Simp. Geol. Minas Gerais, 4, Belo Horizonte, SBG, Anais*, 7, 358-368.

Vieira, F.W.R. (1988). Processos epigenéticos da formação dos depósitos auríferos e zonas de alteração hidrotermal do Grupo Nova Lima, Quadrilátero Ferrífero, Minas Gerais. *Cong. Bras. Geol., 35, Belém, SBG, Anais*, 1, 76-86.

Vieira, F.W.R. (1991). Textures and process of hydrothermal alteration and mineralization in the Nova Lima Group, Minas Gerais, Brazil. In: *Brazil Gold'91*, ed. Ladeira, E.A., pp. 319-325. Rotterdam, A.A. Balkema.

Vieira, F.W.R. and Oliveira, G.A.I. (1988). Geologia do distrito aurífero Nova Lima, Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 377-391. DNPM, Brasília.

Vilela, O.V. (1986). As jazidas de minério de ferro dos municípios de Porteirinha, Rio Pardo de Minas, Riacho dos Machados e Grão-Mogol, norte de Minas Gerais. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 111-120. DNPM, Brasília.

Walde, D.H.G., Gierth, E. and Leonardos, O.H. (1981). Stratigraphy and mineralogy of the manganese ores of Urucum, Mato Grosso, Brazil. *Geologische Rundschau*, **70**, 1077-1085.

Weska, R.K., Svisero, D.P. and Leonardos, O.H. (1997). Geologia da região diamantífera de Poxoréu e áreas adjacentes, Mato Grosso. In: *Simpósio Brasileiro de Geologia do Diamante, 2, Cuiabá, Programa, Resumos, Palestra e Roteiro de Excursão, Universidade Federal de Mato Grosso, Gráfica Universitária, Publicação Especial 03/97*, 35-36.

White, M.G. (1957). The investigation of Uranium possibilities in the Cretaceous sequence in Northeast Bahia. *IUBR-35* (unpublished report).

Wirth, K.R., Gibbs, A.K. and Olszewski Jr., W.J. (1986). U-Pb ages of zircons from the Grão Pará Group and Serra dos Carajás Granite-Pará (Brazil). *Revista Brasileira de Geociências*, **16**, 195-200.

Yamaoka, W.N. and Araújo, E.M. (1988). Depósito de ouro de Mina III, Crixás, Goiás. In: *Principais Depósitos Minerais do Brasil*, coords. Schobbenhaus, C. and Coelho, C.E.S., vol. III, 491-498. DNPM, Brasília.

Zang, W. and Fyfe, W.S. (1993). A three stage genetic model for the Igarapé Bahia lateritic gold deposit, Carajás, Brazil. *Economic Geology*, **88**, 1768-1779.

Almeida, M.E., Ferreira, A.L., Brito, M.F.L. and Monteiro, M.A.S. (1999). Proposta de evolução tectono-estrutural para a região do médio-alto curso do rio Tapajós (estados do Pará e Amazonas). In: *Simpósio de Geologia da Amazônia, 6, Boletim de Resumos Expandidos, Sociedade Brasileira de Geologia, Manaus*, 297-300.

Faraco, M.T.L., Carvalho, J.M.A. and Klein, E.L. (1996). Carta Metalogenética da Província Aurífera do Tapajós. In: *Contribuições à Geologia da Amazônia*, coords. Costa, M.L. and Angélica, R.S., pp. 423-437. FINEP-Sociedade Brasileira de Geologia, Belém,

Giuliani, G., Cheilletz, A., Zimmermann, J.L., Ribeiro-Althoff, A.M., France-lanord, C. and Ferrand, G. (1997). Les gisements d'émeraude du Brésil: genèse et typologie. *Chronique de la Recherche Minière*, 526, 17-61.

Schobbenhaus, C. and Bellizzia, A. (2000). *Geologic Map of South America, 1:5 000 000*. Commission of the Geological Map of the World-CGMW/National Department of Mineral Production-DNPM/Geological Survey of Brazil-CPRM/ The United Nations Educational, Scientific and Cultural Organization-UNESCO (in press).

Schobbenhaus, C., Campos, D.A., Derze, G.R. and Asmus, H.E. (1981). *Geologic Map of Brazil and Adjoinig Ocean Floor Including Mineral Deposits, 1:2 500 000*. Departamento Nacional de Produção Mineral - DNPM, Brasília.

Vasquez, M.L., Klein, E.L., Quadros, M.L.E., Bahia, R.B.C., Santos, A., Ricci, P.S.F., Sachett, C.R., Silva, C.M.G. and Macambira, M.J.B. (1999). Magmatismo Uatumã na Província Tapajós - Novos dados Geocronológicos. In: *Simpósio de Geologia da Amazônia, 6, Boletim de Resumos Expandidos, Sociedade Brasileira de Geologia*, Manaus, 471-474.

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