CHAPTER 3

THE SÃO FRANCISCO PALAEOCONTINENT

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3.1. Introduction

The São Francisco Craton (Almeida, 1977) and its counterpart in Africa, the Congo Craton (Trompette, 1994), represent sectors of a Neoproterozoic palaeocontinent preserved from the Brasiliano–Pan African orogeny as recorded in their marginal belts. The limits between the São Francisco Craton and the surrounding belts (Brasília, Araçuai, Rio Pardo, Sergipano, Riacho do Pontal and Rio Preto belts; Figure 3.1) are marked by intense folding and overthrusting. In the interior of the craton, the Neoproterozoic sedimentary cover is only gently deformed or horizontal and preserved in isolated basins with distinct names: the São Francisco, Irecê and Una-Utinga basins. The São Francisco basin includes basal assemblages of Palaeo- and Mesoproterozoic age (e.g. Martins-Neto and Alkmim, 2001). However, in this chapter, we restrict to the discussion to the Neoproterozoic sedimentary assemblages comprising the São Francisco Supergroup (Alkmim and Martins-Neto, 2001; Martins-Neto and Alkmim, 2001), which is characterised by a glaciogenic unit at the base Jequitai Formation and an argillaceous–calcareous–arkosic unit (Bambuí Group) at the top.

The Neoproterozoic evolution of the São Francisco Craton, including the Brasiliano marginal belts and the sedimentary cratonic cover, is reviewed in this chapter. The lithostratigraphy, Neoproterozoic-aged mineralisation (lead, zinc, fluorine, barium and phosphates) and available geochronological and stable isotope data are also presented.

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It is well known that the late Neoproterozoic was marked by significant changes in Earth’s climate. Clues to these environmental perturbations are recorded in glacial deposits, sometimes formed in equatorial latitudes (Sohl et al., 1999) and often found overlain by dolostones or limestones (so-called cap carbonates; e.g. Kennedy et al., 1998). These strong climatic variations seem to have occurred between 750 and 580 Ma, during at least three distinct glacial epochs, commonly termed Sturtian, Marinoan and Gaskiers glacial events (Halverson et al., 2005). In this chapter, the geological record of Neoproterozoic glacial events and corresponding stable isotopic data for related cap carbonates in the São Francisco Craton are also discussed.

### 3.2. Synthesis of the Lithostratigraphic Units

#### 3.2.1. The Neoproterozoic cover of the São Francisco Craton

**3.2.1.1. The Bambuí Group**

Over the São Francisco Craton, the Bambuí Group (São Francisco basin; Figure 3.1) is the most important Neoproterozoic sedimentary unit, covering large cratonic areas and occupying, in central Brazil, segments of the eastern side of the Brasília belt. It overlies the Paranaí Group from which it is separated by an unconformity filled
by glacial diamictites of the Jequitaí Formation (Figure 3.2). The Paranoá sedimentary rocks consist mostly of mature siliciclastic sediments such as quartzites, with intercalations of metasiltstones and minor lenses of limestones and dolostones. This group has been divided into nine lithostratigraphic units (Faria, 1985), beginning with a paraconglomerate, followed by transgressive and regressive siliciclastic-dominated cycles, and ending with pelites and dolostones containing *Conophyton metula* Kirichenko stromatolites (Cloud and Dardenne, 1973; Cloud and Moeri, 1973; Dardenne et al., 1976; Dardenne, 1979). Available geochronological and microfossil data for the Paranoá Group point to an age of 1,170–950 Ma and a source region in the Palaeoproterozoic sialic basement of the craton, suggesting sedimentation on a passive margin (Guimarães, 1997; Pimentel et al., 1999).

The Jequitaí Formation (Figure 3.3a and b), at the base of the Bambuí Group, consists of clast-supported, greenish or grey diamictites (quartzite, granite, gneiss, limestone, dolostone and siltstone) with clay-rich matrix, sometimes marly, and minor siltstone and sandstone lenses. This unit crops out discontinuously in central Brazil and overlies both the Paranoá rocks and the basement (Dardenne, 1978a, 1979, 2000; Karfunkel and Hoppe, 1988; Uhlein, 1991, 2004; Uhlein et al., 1999) and represents syn-glacial deposition over a wide area of the São Francisco Craton. Locally, a thin horizon of a light-grey to pink cap dolostone overlies the glaciogenic diamictites.

Deposits of the Jequitaí Formation, probably Sturtian in age (~750 Ma), are well exposed at the margins of Serra do Cabral and Agua Fria areas in the southeastern portion of the São Francisco Craton. This formation, up to 150 m thick, overlies quartzites of the Palaeo-Mesoproterozoic Espinhaço Supergroup with a slight unconformity in the Serra do Cabral area. The tops of these quartzites show subglacial erosion structures such as grooved and striated pavements with striae-oriented ENE–WSW. The Jequitaí Formation consists of massive and stratified diamictites with granules, pebbles and boulders of gneisses, granites, quartzites and carbonates (Uhlein et al., 2004). The diamictites are massive in their lower part and in their upper part exhibit alternating clast-rich and

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<tr>
<th>Column</th>
<th>Lithology</th>
<th>Formation</th>
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<tr>
<td></td>
<td>greenish and reddish arkosises</td>
<td>Três Marias Formation</td>
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<tr>
<td></td>
<td>greenish siltstones</td>
<td>Serra da Saudade Formation</td>
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<td>greenish shales</td>
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<tr>
<td></td>
<td>black oolitic limestones and marls</td>
<td>Lagoa do Jacaré Formation</td>
<td>Bambuí Group</td>
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<tr>
<td></td>
<td>dark grey shales and siltstones with lense of</td>
<td>Serra da Santa Helena Formation</td>
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<td></td>
<td>sandstones and limestones</td>
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<td></td>
<td>light grey to pink dolomites</td>
<td>Sete Lagoas Formation</td>
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<td>grey limestones</td>
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<td>pink cap dolomites</td>
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<td></td>
<td>siltstones</td>
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*Figure 3.2* Lithostratigraphic column of the Bambuí Group (Dardenne, 2007a).
clast-poor beds (Figure 3.3). This formation also contains lenticular sandstones, siltstones and a few laminated siltstone—mudstone intercalations. Clast-poor diamictites overlying striated pavement suggest two phases of the glacial event. The first phase, probably on the continent, produced ice erosion (striated pavement) and the second one a clast-poor diamictite deposited by gravity flows during ice retreat and sea-level rise.

The Jequitai Formation, also named the Carrancas Formation, is correlative to the Bebedouro Formation that underlies carbonates of the Salitre Formation (Una Group) in the northeastern part of the São Francisco cratonic area. In the Onça do Pitangui map sheet, in the state of Minas Gerais, layers of unmetamorphosed varvites (Figure 3.4) of the Carrancas Formation form a 30m-thick deposit with dropstones (informally named the Moema sequence by Rocha-Campos et al., 2007) towards the top of the Jequitai Formation.

Figure 3.3  (a) Lithofacies variations of the Jequitai Formation near Serra do Cabral, Minas Gerais, and Cristalina, Goias (Uhlein et al., 1999); (b) diamictite of the Jequitai Formation at Serra da Água Fria, Minas Gerais.
The post-glacial transgression flooded the craton, during which time pelitic and carbonate sediments of the lower Bambuí Group were deposited (Figure 3.2) over at least 300,000 km² (Dardenne, 1978a,b, 1979, 2000; Misi and Kyle, 1994). This sedimentary association, which follows the Jequitai glaciation, is repeated in three regressive megacycles. Each of these megacycles begins with a regional marine transgression associated with basin subsidence, the evidence for which is seen in deep pelitic marine facies, passing upwards into shallow platform, subtidal and supratidal facies.

From base to top, these megacycles are arranged as follows. Megacycle I is pelitic–calcereous, corresponding to the Sete Lagoas Formation, forming a coarsening-upward sequence with calcilutites in the basal portion and passing into limestones and dolostones at the top. At the base, characteristic pink dolomitic calcilutites with green argillaceous films are generally observed. They pass progressively to dark-grey and black-laminated calcilutites. The limestones, generally dark-grey to black and microcrystalline, are well stratified and homogeneous, but lenticular in large scale. To the top, carbonate facies are predominantly dolomitic, distinguished by intraclasts and oolites. The first cycle ends with an extensive subaerial exposure marked by tepee structures, moulds of sulphate nodules, dissolution breccias, and laminated and columnar stromatolites (Dardenne, 1978a,b, 1979; Dardenne and Freitas-Silva, 1999).

Megacycle II begins with the Serra de Santa Helena Formation, which is dominantly pelitic at the base and signals sudden but broad subsidence of the basin. The pelites are followed by dark-grey platformal limestones of the Lagoa do Jacaré Formation, which were deposited in an environment dominated by storms and tidal currents.

Megacycle III is pelitic to sandy and represented by the pelitic Serra da Saudade Formation at the base, deposited in deep-platform environment with episodic influence of storms, and by the Três Marias Formation at the top, dominated by arkose and deposited in a shallow-platform environment influenced by storm currents with episodic tidal to supratidal facies. The pelites are distinctly greenish in colour, containing abundant detrital mica grains in bedding planes, while the greenish or reddish arkoses are fine-grained and rich in plagioclase and mica. Some contain volcanic lithic fragments derived from the erosion of a Brasiliano magmatic arc to the east and are regarded as a foreland basin molasse.

Along the São Francisco River, between the cities of Januária, Itacarambi, Montalvânia and Manga, well-developed and continuous carbonate horizons (Figure 3.5) have been named the Januária and Nhandutiba formations (Dardenne, 1978a,b). The former represents a typical coarsening-upward regressive megacycle, consisting, from bottom to top, of dark-grey calcilutites, dark-grey to grey limestones with lamellar breccias or oolites, pink saccharoidal dolarenites and beige lithographic dolostones. Evidence of emergence is sometimes observed between saccharoidal dolarenites and lithographic dolostones, indicating discontinuity or unconformity in the upper portion of the Januária Formation. This unconformity underlines the impermeable level of lithographic dolostone that has controlled the percolation of basinal fluids responsible for development of secondary dolomitisation, formation of saccharoidal dolarenite and development of Mississippi valley-type Pb-Zn-F mineralisation.

The carbonate and pelite facies are considerably thicker in the region between the Januária—Itacarambi cratonic area and Vazante (Figure 3.6), reflecting regional subsidence induced by differentiate movement along N-S faults during deposition, generating significant westward deepening of the basin (Alvarenga and Dardenne, 1978; Dardenne, 1978a,b, 1979, 2000).

The cap dolostone at the base of the Sete Lagos Formation overlying Carrancas diamictites has been dated at ca. 740 ± 22 Ma (whole-rock Pb-Pb isochron on carbonates; Babinski et al., 2007) and this age, along with C and Sr isotope data for carbonates of the Bambuí Group (Santos et al., 2000, 2004; Vieira et al., 2007), supports a Sturtian age for the Carrancas glaciation and a Cryogenian age for at least the base of the Bambuí Group.
3.2.1.2. The Una Group

The carbonate and siliciclastic units of the Una Group are present in four now disconnected “sub-basins” in the northeastern sector of the São Francisco Craton. The Irecê, Campinas, Una-Utinga and Ituçu “sub-basins” were probably connected before the end of the Pan African/Brasiliano orogeny, forming a large sedimentary basin.
covering much of the craton. Possible correlative successions (although with separate nomenclature) are present in the passive margin basins of the Sergipano (Vasa Barris Group) and Rio Pardo belts, bordering the cratonic area (Figure 3.7). The Una Group is composed of two megasequences (Figure 3.8):

(a) Glaciogenic megasequence: This interval is represented by the Bebedouro Formation that unconformably and variably overlies Palaeoproterozoic gneisses and migmatites and Mesoproterozoic metasedimentary rocks of the Chapada Diamantina Group. The Bebedouro Formation consists of diamictites, pelites and sandstones with a variety of lithofacies that are grouped into four associations (Guimarães, 1996): (i) aeolian (extraglacial), (ii) ice-contact, (iii) pro-glacial and (iv) iceberg melting.

(b) Carbonate megasequence: This package is dominantly carbonate, with subordinate siliciclastic units, and unconformably overlies the Bebedouro Formation. Referred as the Salitre Formation, the megasequence is subdivided into five informal units (Misi and Souto, 1974; Misi, 1979), each of which can be correlated with the Bambuí Group. From top to bottom:

Unit A1 is a black organic-rich, cross-stratified and rippled oolitic and pisolithitic limestone that correlates with the Lagoa do Jacaré Formation.

Unit A comprises interbedded marls, shales and siltstones with local lime grainstone beds. It correlates with the Serra de Santa Helena Formation.

Unit B1 consists mainly of grey to reddish dolostone with tepee structures, replaced, nodular evaporites and intraformational breccia. It correlates with the dolomitic facies at the top of the Sete Lagoas Formation.

Unit B is a grey-laminated limestone and dolomitic limestone that grades upward into the more dominantly dolomitic Unit B1. It correlates with the Sete Lagoas Formation beneath the upper dolomitic facies.

Unit C is a pink argillaceous dolostone that unconformably overlies the Bebedouro Formation and correlates with the base of the Sete Lagoas Formation.

A composite stratigraphic section of the Una Group is represented in Figure 3.8, while Table 3.1 shows the possible correlations between the Una (Vieira et al., 2007) and Bambuí groups.
3.2.2. The surrounding belts

3.2.2.1. The Brasília belt

In the western margin of the São Francisco Craton, a thick Meso–Neoproterozoic sedimentary/metasedimentary pile forms the Brasília belt (Araí, Paranoá, Serra da Mesa, Araxá, Ibiá, Vazante, Canastra and Bambui) that extends for more than 1,000 km (Figure 3.9) and is part of a passive margin association. These rocks are mostly undeformed and unmetamorphosed over the craton and increasingly deformed and metamorphosed westward, reaching amphibolite and granulite facies conditions in the central part of the belt (Dardenne, 1978a,b, 2000; Fuck et al., 1994; Pimentel et al., 2000; Piuzana et al., 2003). The evolution of the deformation and metamorphism reflects the vergence of this belt towards the São Francisco Craton. The tectonic zonation proposed by Costa and Angeiras (1971) was reformulated by Fuck et al. (1994) who identified an internal zone in the west, a cratonic zone in the east and an external zone in between.

A WNW-ESE lineament at about the latitude of Brasília allows a subdivision of the Brasília belt into two segments with distinct geotectonic histories. In the northern segment, sedimentary units were metamorphosed to greenschist facies or are unmetamorphosed and the well-preserved stratigraphy allows the reconstitution of the palaeogeography and depositional systems. In the area to the north of Brasília, where sequences of the Paranoá and Araí groups occur, the compressional tectonics is expressed in right lateral transcurrent faults and thrusts that locally have affected the sedimentary cover.

The southern segment exhibits distinct tectonic features when compared to the northern one. Deformation and metamorphism are very intense and have obliterated stratigraphic relationships between the various units. The Araxá, Canastra, Ibiá and Vazante groups are involved in a complex, imbricated system of nappes and thrusts.
indicating a large amount of tectonic transport, on the order of tens to hundreds of kilometres. The contacts between the various involved assemblages correspond to low-angle shear zones, with sheath folds and lateral ramps. The external zone is characterised by a well-developed thin-skinned system of folding and overthrusting from west to east, with increasing deformation and metamorphism towards the internal zone to the west. From east to west, it is possible to recognise successive rock assemblages belonging to the Bambui, Vazante and Ibia–Araxá groups.

Lithostratigraphic units of the Bambui Group are easily identified on the flanks of anticlines and synclines, frequently faulted, with N-S to N20E trending axes, as in the Formosa and Bezerra areas to the east of Brasília. In the core of anticlines, one observes sandstones, arkoses, rhythmites and stromatolitic dolomites of the Paranoá Group, overlain in erosional unconformity by Bambui sedimentary rocks of the Jequitaí, Sete Lagoas, Serra de Santa Helena, Lagoa do Jacaré, Serra da Saudade and Três Marias formations (Figure 3.2).

In the Cristalina area (Figure 3.1), diamictites of the Jequitaí Formation occur in the flanks of a large brachiatricline, disconformably overlying quartzites of the Paranoá Group (Faria, 1985; Cukrov et al., 2004; Uhlein et al., 2004). In this area, phyllites of the Canastra Group overthrust Jequitaí diamictites.

![Composite stratigraphic cartoon showing the two megasequences of the Una Group](image)

**Figure 3.8** Composite stratigraphic cartoon showing the two megasequences of the Una Group (modified from Misi et al., 2007).

**Table 3.1** Correlations between the Una and Bambui groups at the formalional level.

<table>
<thead>
<tr>
<th>Group</th>
<th>Bambui (Formations)</th>
<th>Una (Formations)</th>
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</thead>
<tbody>
<tr>
<td>Lithotypes</td>
<td>Arkose, siltstone</td>
<td>Três Marias</td>
</tr>
<tr>
<td></td>
<td>Siltstone, pelites</td>
<td>Serra da Saudade</td>
</tr>
<tr>
<td></td>
<td>Black oolitic limestone</td>
<td>Lagoa do Jacaré</td>
</tr>
<tr>
<td></td>
<td>Marl, shales</td>
<td>Serra Santa Helena</td>
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<tr>
<td></td>
<td>Dolostone with tepee</td>
<td>Sete Lagoas</td>
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<tr>
<td></td>
<td>Laminated limestone</td>
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<tr>
<td></td>
<td>Pink dolostone</td>
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<tr>
<td></td>
<td>Diamictite, arkose, pelitic rocks</td>
<td>Jequitai, Carrancas</td>
</tr>
</tbody>
</table>

indicating a large amount of tectonic transport, on the order of tens to hundreds of kilometres. The contacts between the various involved assemblages correspond to low-angle shear zones, with sheath folds and lateral ramps. The external zone is characterised by a well-developed thin-skinned system of folding and overthrusting from west to east, with increasing deformation and metamorphism towards the internal zone to the west. From east to west, it is possible to recognise successive rock assemblages belonging to the Bambui, Vazante and Ibia–Araxá groups.

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Figure 3.9  Simplified geological map of the Brasilia belt showing the Palaeo-Meso-Neoproterozoic units, according to Dardenne (2000), Pimentel et al. (2001) and Valeriano et al. (2004). Pb-Zn deposits: (1) Vazante; (2) Morro Agudo; (3) Fagundes; (4) Ambrosia (Cunha et al., 2007).
In the southwestern portion of the São Francisco Craton, unusual conglomeratic facies occur in the Lagoa Formosa, Samburá and Carmo do Rio Claro formations of the upper Bambuí Group. The Lagoa Formosa Formation (Seer, 2001) crops out for over 300 km between the towns of Tiros and Presidente Olegário (Serra do Abaeté Mountains), where it shows little deformation with only a weak schistosity indicating northeastern vergence. This assemblage (Figure 3.10) probably correlates with the Serra da Saudade Formation and comprises two interbedded sequences: (i) stratified and massive diamictites (500–1,000 m thick) and (ii) a well-stratified 1,000–2,000 m-thick sequence of fine-grained diamictites, siltstones, sandstones, greywackes, jaspillites and limestones (Uhlein et al., 2004). Pebbles and boulders are predominantly of intrabasinal arkoses and siltstones floating in a greenish grey pelitic to sandy-pelitic matrix. Some clasts are composed of limestone, chert, quartzite and granite-gneiss. These paraconglomerates are interpreted as gravitational debris flows originated by regional uplift of the internal zone of the Brasília belt. This interpretation is supported by a northeasterly decrease in the size of clasts in the diamictites from boulders to granules. At the base of the Lagoa Formosa Formation, thin horizons of phosphorite are intercalated with green pelites, with abundant potassium-rich illite, and known as the Verdete facies. Carbonate beds (calcirudites, calcarenites and columnar stromatolites up to 10 m thick) are interbedded with siltstones in the lower portion of the unit and may correlate with either the top of the Lagoa do Jacará Formation (Dardenne, 1979) or the base of the Lagoa Formosa Formation (Lima and Uhlein, 2005).

The Samburá Formation (Branco, 1957) is the name given to various minor occurrences of conglomerates, generally isolated within pelites of the upper portion of the Bambuí Group, showing similar characteristics. Most of these conglomerates are clast-supported and composed mainly of granitic, rhyolitic, arkosic and quartzitic pebbles. They are interpreted to have been deposited in eastward-tapering alluvial fans (Figure 3.11) (Castro and Dardenne, 1996).

The Carmo do Rio Claro Formation near the Furnas dam, described by Heilbron et al. (1987), is a combination of conglomerates, diamictites, arkoses and siltstones that correlate with the Samburá Formation, as well as slates, phyllites and marbles. The diamictites, which contain rounded gneiss, metapelite and metasandstone...
pebbles, are considered to be derived from the west. The sequence is overthrust by rocks of Canastra and Araxá groups, which form the Passos Nappe. The Carmo do Rio Claro Formation is stratigraphically and genetically equivalent to the Lagoa Formosa Formation.

The Vazante Group metasediments are a thick sequence of marine pelites and dolostones that crop out for over 300 km along a north-south trend in northwestern Minas Gerais state, spanning the cities of Coromandel, Lagamar, Vazante, Paracatu and Unaí. Misi et al. (2007) have argued that the Vazante and Bambui groups are correlative, based on equivalent Sr isotope signatures in both successions ($\sim$ 0.7074) and the interpretation of seismic profiles by Romeiro-Silva and Zalan (2005). This controversial argument conflicts with the contrasting C isotope records between the two units and new Re-Os ages (unpublished) of organic shales of the Serra do Garrote, Serra do Poço Verde and Serra da Lapa formations that imply a late Mesoproterozoic age (A.J. Kaufman and K. Azmy, oral communications).

The Vazante Group is subdivided into seven formations (Figure 3.12), from base to top: Santo Antônio do Bonito, Rocinha, Lagamar, Serra do Garrote, Serra do Poço Verde, Morro do Calcario and Lapa formations (Dardenne, 2001).

The Santo Antônio do Bonito Formation (Souza, 1997), considered to be the basal formation, consists of beds of white quartzites, sometimes conglomeratic, intercalated with slates. In the Santo Antônio do Bonito and Santo Inacio rivers, this formation is characterised by the presence of diamictite horizons within the quartzites that contain limestone, dolostone, metasiltstone and granitic pebbles floating in a pelitic matrix, which is locally phosphatic. Larger concentrations of phosphate are found in the slaty facies and in phospharenite layers, rich in

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**Figure 3.11** Schematic geological column of the Samburá Formation near Samburá hydroelectric dam (after Castro and Dardenne, 1996).
intraclasts and pellets (Phosphorite 1). The diamictites represent debris flows deposited in relatively deep waters in a glacial environment.

The Rocinha Formation (Dardenne et al., 1997) is composed of a basal rhythmic sandy and pelitic sequence that is gradational with the underlying Santo Antônio do Bonito Formation. The upper Rocinha Formation is a thick package of regularly intercalated slates and metasiltstones. It passes upward into dark-grey marly and pyritic slates, with thin phosphatic laminations that transition into the intraclast- and pellet-rich phospharenites (Phosphorite 2), forming the Rocinha phosphate deposit. In the upper part of the formation, rhythmic sediments (quartzites and siltstones) contain the Lagamar phosphate deposit composed almost entirely of phospharenites (Phosphorite 3).

![Lithostratigraphic column of the Vazante Group (Dardenne, 2007b).](image)

**Figure 3.12** Lithostratigraphic column of the Vazante Group (Dardenne, 2007b).
The Lagamar Formation is a psammitic—pelite—carbonate unit represented in its basal portion by alternations of conglomerate, quartzite, metasiltstone and slate. The conglomerates, ascribed to the Arrependido Member, consist of a framework of quartzite, metasiltstone and dark-grey limestones clasts. The psammites are over lain by dolomitic intraformational breccias passing upward into dark-grey, well-stratified limestones with intercalations of lamellar breccias followed by stromatolitic dolostones, which comprise beige to pale pink bioherms composed of microbialaminated dolostones, oncolitic dolarenites and dolorudites, and columnar stromatolites with convex and conical laminations classified as *C. metula* Kirichenko and *Jasutophyton* type (Dardenne et al., 1976). Laterally and vertically, these bioherms interfinger with marly metasiltstones and slates.

The Serra do Garrote Formation (Madalosso and Valle, 1978; Madalosso, 1980), which is about 400 m thick, comprises thick, dark-grey to greenish grey slates, sometimes rhythmic, carbonaceous and pyritic, with fine-grained quartzite intercalations. The Serra do Poço Verde Formation (Dardenne, 1979) is dominantly dolomitic and is divided into four members (Rigobello et al., 1988), described successively from the base to the top. The 500 m-thick Lower Morro do Pinheiro Member is composed of light-grey- and/or pink-laminated dolostones with cyanobacteria mats, intercalated with oncolitic dolarenites and intraformational breccias and lenses of dolostones with columnar stromatolites. The Upper Morro do Pinheiro Member is composed of medium- to dark-grey microbialaminated dolostones with common birdseye fenestrae, intercalated with layers of dolarenites, lamellar breccias and carbonaceous shales (thickness from 300 to 500 m). The occurrence of *Conophyton cylindricus* Maslov or *C. metula* Kirichenko (Moeri, 1972; Cloud and Dardenne, 1973) near the Cabeludo village is probably related to this unit. The Lower Pamplona Member is 100–200 m thick and composed of grey, green and purple siltstones intercalated with micritic and microbialaminated dolostones with small lenses of fine-grained to conglomeratic sandstones. The ~400 m-thick Middle Pamplona Member is composed of light-grey to pink microbialaminated dolostones with locally developed mud cracks and nodular barite, intercalated with beds of dolarenite, lamellar breccia and columnar stromatolites, and with lenses of shales. The 200–300 m-thick Upper Pamplona Member is characterised by the presence of light-grey to pink stromatolitic dolostones constituting biostrones and bioherms with convex-laminated columns and associated with oolitic and oncolitic dolarenites and dolorudites. In the region of Morro Agudo, Paracatu and Unai, this unit corresponds to the Morro do Calcario Formation, which is here >900 m thick and composed mainly of dolorudites derived from reworked but still partially preserved stromatolitic bioherms and associated oolitic and oncolitic dolarenites. In this region, the Morro do Calcario and Serra do Poço Verde formations cannot be easily subdivided as they are in the Vazante region.

The Lapa Formation constitutes the upper portion of the Vazante Group described in the region of Paracatu as Serra do Velosinho and Serra da Lapa members. These are represented by carbonaceous phyllites, marly metasiltstones, dolomitic lenses and quartzite layers. The dolomitic lenses show various facies, including microbialaminated dolostones, columnar stromatolites and intraformational breccias, interfinering with the pelitic sequence that regionally overlies the dominantly dolomitic formations of Morro do Calcario and Serra do Poço Verde. In the Unai region, the Lapa Formation comprises a rhythmic turbidite sequence of siltstones and pelite, locally with lithic sandstones and conglomerates intercalated with dark-grey slates. The clasts are fragments of phyllites, fine-grained quartzites and cherts.

The presence of *C. cylindricus* Maslov and *C. metula* Kirichenko in the dolomitic bioherms near Cabeludo and Lagamar implies a correlation with the Paranoá Group (Dardenne, 2000). Carbon and Sr isotope signatures for dolomitic rocks of these groups are fairly distinct from signatures for the Bambuí Group carbonates and are more consistent with a Mesoproterozoic age (Santos et al., 2000, 2004). Consequently, the glaciogenic diamictite of the Santo Antônio do Bonito and the overlying sedimentary sequence of Vazante Group have an unknown age, but are inferred to be Kaigas-equivalent (c. 750 Ma) or older.

Recently, various workers (Olcott et al., 2005; Azmy et al., 2005) have proposed the occurrence of a regional glacial event spanning most of the Upper Pamplona Member and Lapa Formation in the Vazante area, and Morro do Calcario, Mocambo and Lapa Formations near The Morro Agudo mine. This contention is based on the presence of a layer of brecciated rocks with dropstones, interpreted as possible glacial diamictites with a negative δ13C excursion in overlying dolostones. These breccias may be alternatively interpreted as intraformational dolomitic breccias, possibly of debris flow origin (Dardenne, 2007b). If confirmed, the glacial episode recorded in the Lapa Formation would be related to the Sturtian glaciation, represented regionally by the Jequitai Formation. This suggests that the Vazante Group constitutes a stratigraphic equivalent of the upper portion of the Paranoá Group, deposited in a passive margin setting, as previously proposed by Campos Neto (1984) and Dardenne (2007a,b).

### 3.2.2.2. The Aracuá belt

The name Aracuá was formerly applied only to the arcuate marginal belt located along the east and southeast edges of the São Francisco Craton (Almeida, 1977; Figure 3.1). The Aracuá orogen was defined after the identification of several geotectonic components (e.g. ophiolites, pre-collisional magmatic arc and syn-collisional
granites) related to that marginal belt (Pedrosa-Soares et al., 2001, 2008). This orogen is 700 km long and up to 500 km wide and encompasses the entire region between the São Francisco Craton and the Atlantic continental margin, between 15°S and 21°S in eastern Brazil (Figure 3.13), including the southeastern limit of the São Francisco Craton. The Araçuai orogen and its counterpart located in southwestern Africa (the West-Congo belt; Trompette, 1994; Tack et al., 2001) form a confined orogen evolved into an embayment carved into the São Francisco–Congo palaeocontinent (Pedrosa-Soares et al., 2001, 2008; Alkmim et al., 2006b).

The basement of the Araçuai orogen, not described here, consists of Archaean to Mesoproterozoic units, including the rift-sag Espinhaço Supergroup and the related anorogenic Borraçudos suite (Uhlein et al., 1998b; Martins-Neto, 2000; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Noce et al., 2007).

The Araçuai orogen comprises several Neoproterozoic stratigraphic units, such as the Macaúbas Group, which includes glacially influenced deposits and deep marine sedimentation associated with meta-mafic–ultramafic ophiolite slivers, tonalitic batholiths making up a pre-collisional magmatic arc (G1 suite), orogenic metasedimentary successions (e.g. Salinas Formation and Nova Venécia Complex) and a large sector dominated by syn-collisional granites (G2 suite). The belt is intruded by Cambrian-age post-collisional plutons (Figure 3.13).
Two magmatic episodes yield age constraints for the rift stage of the precursor basin of the Araçuaí orogen. The older age from a meta-dolerite dyke of Pedro Lessa, located ca. 50 km to the southeast of Diamantina, was dated at 906 ± 2 Ma (U-Pb, zircon and baddeleyite; Machado et al., 1989). The A-type granites of Salto da Divisa yielded a zircon U-Pb age of 875 ± 9 Ma (Silva et al., 2007).

The precursor rift basin was filled by units of the proximal Macaúbas Group (Figure 3.13). The oldest rift deposits consist of metamorphosed sandstones, arkoses, conglomerates and pelites that lack evidence of glaciation and represent fluvial sedimentation (Karfunkel and Hoppe, 1988; Noce et al., 1997; Martins, 2006). U-Pb ages of detrital zircon grains constrain the maximum sedimentation age for this early rift stage at around 900 Ma (Pedrosa-Soares et al., 2008). These pre-glacial deposits of the Macaúbas basin are covered by metamorphosed, generally massive diamictites with minor sandstones, pelites and rhythmites, representing glacio-terrestrial to gravitational glacio-marine sedimentation (Pedrosa-Soares et al., 1992; Uhlein et al., 1998b, 1999; Martins-Neto et al., 2001; Martins, 2006). This lower diamictite-rich unit was thrust over the Neoproterozoic carbonate-pelite cover of the São Francisco Craton and can be correlated with the glaciogenic Jequitai Formation of the cratonic region (Karfunkel and Hoppe, 1988; Uhlein et al., 1999; Martins-Neto and Hercos, 2002). The eastern part of the proximal Macaúbas Group includes a thick pile of sediments with layered diamictites, pebble-bearing iron formations, graded sandstones and pelites with dropstones, representing glaciogenic debris flows and turbidites deposited in submarine fan environment (Pedrosa-Soares et al., 1992; Uhlein et al., 1999; Martins-Neto et al., 2001). This succession includes mafic volcanic rocks, metamorphosed to greenschist facies, with pillow structures and other features of subaqueous flows and is derived from tholeiitic basalt protoliths with a dominant within-plate signature (Gradim et al., 2005). Inherited zircon grains from older felsic rocks yielded U-Pb ages from Archaean to Late Mesoproterozoic (the youngest at 1,156 ± 7 Ma), with a whole-rock Sm-Nd TDM model age of ca. 1.5 Ga (Babinski et al., 2005). However, some samples with oceanic signatures and slightly positive εNd(900 Ma) of +0.23, together with the inherited zircon grains, suggest a transitional mafic magma that migrated through a thinned continental crust. Accordingly, these greenschists provide strong evidence of volcanism in an extensional marine basin floored by thin continental crust, during the very late rift stage of the Macaúbas basin (Gradim et al., 2005; Pedrosa-Soares et al., 2008).

The upper succession of the proximal Macaúbas Group consists of interlayered sandstones and pelites, metamorphosed from the greenschist to lower amphibolite facies, representing the proximal passive margin sedimentation of the Macaúbas basin. The youngest detrital zircon grain from a sandstone layer constrains the maximum sedimentation age of this shelf succession to around 864 Ma (Pedrosa-Soares et al., 2000a, 2008).

The distal part of the Macaúbas Group is a passive margin sequence free of diamictite, known as the Ribeirão da Folha Formation (Figure 3.13). This unit was formerly considered to be part of the Salinas Formation (Pedrosa-Soares et al., 1992, 1998, 2001), but since the redefinition presented by Lima et al. (2002), it has been separated (Pedrosa-Soares et al., 2008). The Ribeirão da Folha Formation was metamorphosed progressively from west to east to the garnet, staurolite, kyanite and sillimanite zones (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Queiroga et al., 2006). The western part of the Ribeirão da Folha Formation comprises banded quartz-mica schists (fine-grained turbidites) with thin calc-silicate (marl) and marble lenses. The eastern Ribeirão da Folha Formation is coarse-grained massive ortho-amphibolites, representing gabbroic protoliths, and Wiedemann-Leonardos, 2000b). These ortho-amphibolites have a geochemical signature akin to modern ocean-floor mafic rocks and are comparable to similar rocks of other Neoproterozoic ophiolites (Pedrosa-Soares et al., 1998, 2001; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Suiá et al., 2004; Queiroga et al., 2006). The meta-mafic rocks associated with the Ribeirão da Folha Formation are coarse-grained massive ortho-amphibolites, representing gabbroic protoliths, and medium- to fine-grained banded ortho-amphibolites that seem to be derived from dolerites and basalts (Pedrosa-Soares et al., 1998, 2008; Suiá et al., 2004). These ortho-amphibolites have a geochemical signature akin to modern ocean-floor mafic rocks and are comparable to similar rocks of other Neoproterozoic ophiolites (Pedrosa-Soares et al., 1998, 2001; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Suiá et al., 2004). The U-Pb (LA-ICPMS) age of 660 ± 29 Ma obtained from zircon crystals of a plagiogranite vein suggests the epoch of magmatic crystallisation of at least part of the oceanic magmatic rocks (Queiroga et al., 2007).

The Jequitinhonha Complex comprises a thick paragneiss (kinzigitic) association, consisting of migmatised biotite gneisses, with variable contents of garnet, cordierite, sillimanite and graphite, thick intercalations of graphite-rich gneiss, and minor quartzite and calc-silicate rocks. The paragneiss protoliths comprise marine arkosic greywackes to pelaruminous pelites, deposited under oxidising conditions, with horizons of carbonateous black mud that were deposited in a restricted marine environment. The sedimentary protoliths are interpreted to have been deposited during the passive margin stage of the Macaúbas basin (Pedrosa-Soares and Wiedemann-Leonardos, 2000b).

The western external domain of the Araçuaí orogen, adjacent to the southeastern edge of the São Francisco Craton, is a west-verging fold-thrust belt, with increasing intensity of metamorphism from west to east.
The northern segment of the external tectonic domain bends to the east, outlining the northern curvature of the orogen, where it shows north-verging structures and metamorphic grade increasing from north to south (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Pedrosa-Soares et al., 2001, 2008; Uhlein, 2004; Alkmim et al., 2006b).

Along the western external domain, the major normal faults of the precursor rift basin, active during the Tonian, were reactivated as thrust faults during the contractional orogenic event (Uhlein et al., 1999; Martins-Neto and Hercos, 2002; Uhlein, 2004). The main deformational event, related to the syn-collisional stage, is characterised by zones with asymmetrical folds showing westward vergence separated by mylonitic belts of brittle—ductile shear zones. An east—southwest dipping S1 cleavage or mylonitic schistosity with down-dip stretching lineation and kinematic indicators such as S-C surfaces, patterns of deformed and boudinaged veins, asymmetric deformation around porphyroclasts and clasts, and sheath folds indicate a westward tectonic transport towards the São Francisco Craton (Uhlein, 1991; Pedrosa-Soares et al., 1992; Uhlein et al., 1995, 1998b; Cunningham et al., 1998; Alkmim et al., 2006b). Thrusts and reverse faults are defined by mylonites in which S-C fabrics or porphyroclast tails indicate top-up-to-the-west movement. Pebbles within Macaúbas diamictite layers were stretched with their long axis parallel to the E-W mineral lineation (Uhlein, 1991; Pedrosa-Soares et al., 1992). Rocks of the eastern proximal Macaúbas Group show asymmetric extensional crenulation cleavage (S2) as the axial surface of F2 folds. These structures show top-down-to-the-east tectonic transport and indicate a regional-scale normal-sense shear zone probably related to the extensional (gravitational) collapse of the Araçuai orogen (Marshak et al., 2006).

The internal tectonic domain is the high-grade metamorphic core of the orogen, which includes the pre-collisional magmatic arc, paragneiss complexes and a huge anatectic zone dominated by syn-collisional S-type granites and post-collisional intrusions (Figure 3.13). The internal tectonic domain is generally west-verging, but the back–arc zone between 17°30′S and 20°S shows syn-collisional tectonic transport to the east (Pedrosa-Soares et al., 2001, 2006, 2008; Alkmim et al., 2006b; Vauchez et al., 2007). The southern sector of the Araçuai orogen and its connection with the Ribeira orogen are marked by syn-collisional west-vergent thrusts cut by late, dextral strike-slip shear zones (Cunningham et al., 1998; Heilbron et al., 2004; Peres et al., 2004; Alkmim et al., 2006b). High-grade metamorphism and partial melting are syn-kinematic with the regional foliation and took place at ca. 585—560 Ma. The internal domain shows widespread granitoids of the G1 and G2 suites (Figure 3.13). In general, G1 and G2 batholiths display high-temperature (600—700°C) solid-state foliation parallel to the magmatic orientation, both concordant with the regional foliation of the country rocks, suggesting synchronous deformation of the magmatic rocks and their host rocks (Uhlein et al., 1999a; Pedrosa-Soares et al., 2001, 2006; Vauchez et al., 2007).

The Araçuai orogen records four evolutionary stages whose timing is constrained by U-Pb ages of the pre-collisional (ca. 630—585 Ma), syn-collisional (ca. 585—560 Ma), late collisional (ca. 560—530 Ma) and post-collisional (ca. 530—490 Ma) magmatic episodes (Pedrosa-Soares et al., 2001, 2008; Silva et al., 2005).

The pre-collisional stage (ca. 630—585 Ma) is mainly represented by the I-type G1 plutonic suite. This suite represents the roots of a magmatic arc formed in a continental active margin setting (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Nalini-Junior et al., 2000b; Pedrosa-Soares et al., 2001, 2008). It consists of deformed batholiths and stocks composed of tonalite, granodiorite and minor diorite, with dioritic to mafic enclaves. Chemical compositions from many samples of several G1 plutons outline a consistent metaluminous, calc-alkaline, volcanic arc signature (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Nalini-Junior et al., 2000b, 2005; Pedrosa-Soares et al., 2001; Campos et al., 2004; Martins et al., 2004b). U-Pb zircon and monazite ages constrain the evolution of the G1 suite to between ca. 630 and ca. 585 Ma (Noce et al., 2000, 2001; Nalini-Junior et al., 2000b; Whittington et al., 2001, Silva et al., 2005). The location of this magmatic arc relative to the zone with oceanic slivers suggests subduction to the east and a suture zone roughly located along the 42°W meridian, south of 17°S (Pedrosa-Soares et al., 1998, 2001, 2008).

The forearc-and-arc rock assemblage includes orogenic units that represent the volcano-sedimentary pile of the magmatic arc and/or contain sediments supplied by it (e.g. Rio Doce Group; Vieira, 2007), together with strongly deformed tonalitic bodies, thrust slices of basement rocks and probable passive margin deposits (Pedrosa-Soares et al., 2008).

The Nova Venécia Complex comprises pelite-rich sediments deposited in the back-arc basin of the Araçuai orogen (Noce et al., 2004). This complex contains migmatised, high-grade paragneisses rich in biotite, garnet, cordierite and/or sillimanite, with lenses of calc-silicate rocks (Pedrosa-Soares et al., 2006, and references therein). U-Pb ages for detrital zircons constrain the maximum age of sedimentation to around 608 ± 18 Ma (revised age by Pedrosa-Soares et al., 2008, after data from Noce et al., 2004). High-grade metamorphism and partial melting are syn-kinematic with the regional foliation and took place at around 580—570 Ma, but the complex was also involved in late orogenic episodes of granite genesis (Noce et al., 2004; Munhá et al., 2005; Pedrosa-Soares et al., 2006).
To the northwest of the magmatic arc is found the Salinas Formation (formerly included in the Macaúbas Group; e.g. Pedrosa-Soares et al., 2001; Lima et al., 2002), a metamorphosed greywacke—pelite succession with intercalations of clast-supported conglomerates locally rich in clasts of volcanic rocks, and calc-silicate (meta-marl) lenses, named. U-Pb data for detrital zircons from greywacke samples suggest a maximum sedimentation age of 588 ± 24 Ma for the Salinas Formation (revised age by Pedrosa-Soares et al., 2008, after data from Lima et al., 2002). Furthermore, zircons from cobbles of felsic volcanic rocks of the Salinas conglomerates yielded U-Pb magmatic ages around 630—600 Ma (Pedrosa-Soares et al., unpublished data). As the magmatic arc of the Araçuaí orogen is the most likely source for these volcanic clasts, as well as for the ca. 588 Ma detrital zircons, the Salinas Formation is considered to be an orogenic sedimentary pile (Lima et al., 2002; Santos, 2007; Pedrosa-Soares et al., 2008). The Salinas Formation records the regional deformation of the Araçuaí orogen, but in low-strain zones, well-preserved sedimentary features, such as convoluted bedding and intraformational breccias suggesting sedimentation in a tectonically active marine environment (Lima et al., 2002; Santos, 2007).

The Capelinha Formation is a metamorphosed sandstone succession, with intercalations of pelite and detrital iron formation and is also interpreted to be sourced from the adjacent orogen (Queiroga et al., 2006; Pedrosa-Soares et al., 2008).

The G2 suite represents the syn-collisional stage (ca. 585—560 Ma). It mainly consists of foliated, S-type, biotite ± garnet ± cordierite ± sillimanite granites and two-mica granites, with common mylonitic features. Xenoliths and roof pendants of metasedimentary rocks in several stages of assimilation and of variable size are abundant (Celino et al., 2000; Nalini-Junior et al., 2000a,b; Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Pedrosa-Soares et al., 2001, 2006; Campos et al., 2004). U-Pb ages suggest that the main episode of G2 granite generation took place at around 575—560 Ma (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Campos et al., 2004; Silva et al., 2005; Pedrosa-Soares et al., 2006), although some G2 granites crystallised at around 582 Ma (Nalini-Junior et al., 2000a). Large batholiths dominated by mylonitic garnet—biotite granites reveal well-preserved magmatic features with similar magmatic crystallisation ages (Pedrosa-Soares et al., 2006; Vauchez et al., 2007).

The late collisional stage apparently lasted from ca. 560 to 530 Ma, but these age limits are poorly constrained geochronologically. The S-type G3 suite is late to post-collisional. It consists of garnet- and/or cordierite-rich leucogranite, which generally occur as veins and small intrusions cutting G2 granites. Zircon and monazite U-Pb data from G3 leucogranite samples have yielded ages from ca. 550 to 520 Ma (Whittington et al., 2001; Campos et al., 2004; Silva et al., 2005; Pedrosa-Soares et al., 2006).

The post-collisional stage (530—490 Ma) is related to the gravitational collapse of the Araçuaí orogen (Pedrosa-Soares and Wiedemann-Leonardos, 2000b; Pedrosa-Soares et al., 2001). G4 and G5 are plutonic suites associated with this stage and both lack the regional foliation (Figure 3.13). The G4 suite includes relatively shallow (5—15 km), S-type granitic intrusions with variable contents of biotite, muscovite and/or garnet. They have circular surface outcrops and sometimes form clusters of amalgamated plutons with locally preserved pegmatoidal roofs (Pedrosa-Soares and Wiedemann-Leonardos, 2000b, and references therein). The G5 suite predominantly consists of I-type granitic intrusions, which may include charnockitic, enderbitic and/or mafic portions, and minor mafic bodies with subordinate granitic and/or charnockitic facies. Magma mingling and mixing features are very common. The granitic rocks are generally porphyritic to subporphyritic and have a high-K calc-alkaline signature (Wiedemann et al., 2002; Campos et al., 2004; Martins et al., 2004b).

The evolution of the precursor basin and orogenic stages of the Araçuaí orogen are presented in Table 3.2. This orogenic belt defines the southeastern limit for the São Francisco Craton, but it does not represent a common plate margin orogen. The Araçuaí orogen (together with its counterpart located in Africa, the West-Congo belt) is better characterised as a confined orogen because it developed inside an embayment (like a large gulf or a Red Sea-type basin) carved into the São Francisco—Congo palaeocontinent. Its precursor basin was not completely ensialic but was filled by continental rift (including glaciogenic) and passive margin sediments. The subsequent orogeny emplaced slivers of ophiolite and a pre-collisional magmatic arc, evidence that rifting proceeded to ocean-floor production (Pedrosa-Soares et al., 2001, 2008).

Regional structural studies provide explanations for the tectonic mechanisms underlying the different stages of the Araçuaí orogen (Uhlein, 2004; Alkmim et al., 2006b). The so-called “nutcracker tectonics” model explains this orogen as a series of kinematically distinct deformation stages: (1) early north-verging deformation; (2) development of the west-verging fold-thrust belt against the São Francisco Craton and an east-vergent zone in the back-arc region; (3) lateral escape to the south along dextral strike—slip faults; (4) late-stage gravitational collapse with development of extensional sites and intrusion of post-collisional plutons (Alkmim et al., 2006b).

3.2.2.3. The Rio Pardo Basin
The Rio Pardo Group is a Neoproterozoic-Early Palaeozoic sedimentary fill of the Rio Pardo basin in the southeastern portion of the São Francisco Craton (Figure 3.14). The first reference to rocks of this group was
made in 1870 (Hartt, 1941) when conglomeratic schist and fine-grained sandstones were described and the impression of a plant resembling Asterophyllites scutigera Dawson (previously known from the Upper Devonian of St. John, New Brunswick, Canada) was found. The Baixo Rio Pardo rocks were thus assigned a Lower Palaeozoic age. The interval for deposition of rocks of this group is still poorly constrained; Karmann et al. (1989) suggested a depositional range from 1,100 to 600 Ma.

Until the middle 1960s, studies of this basin were at the reconnaissance level, and the stratigraphy was subdivided into the Rio Pardo and Salobro formations (Oliveira and Leonardos, 1940; Almeida, 1968). Later, the low-grade metasedimentary rocks of the basin were divided into six formations, formalised by Pedreira et al. (1969): the Panelinha, Camacan, Salobro, Água Preta, Serra do Paraíso and Santa Maria formations, with the Rio Pardo Formation, previously used by Almeida (1968), elevated to group status. The ‘layer cake’ stratigraphy was somewhat modified in subsequent studies (Andrade and Nunes, 1974; Siqueira et al., 1978) that placed the Salobro Formation at the top of the group and introduced other minor changes. Karmann et al. (1989) once again revised the stratigraphy and the interpretation of the basin’s tectonic evolution, from an aborted-rift, giving way to a thermally subsiding basin, followed by a foreland basin.

A stratigraphic chart for the Rio Pardo Basin and its basement is shown in Figure 3.15. The first column shows the ages (palaeontological or geochronological) of the distinct units; the second, third and fourth columns show the relationships among the lithostratigraphic units (group, subgroup and formations); in the central panel the same relationships are in a two-dimensional display, in a north—south section and, in the last two columns, the possible tectonic evolution and depositional environments for the units.

Rocks of the Rio Pardo Basin were deposited on Archaean basement (Moraes Filho and Lima, 2007) that consists of charnockites, enderbites and granulitic trondhjemites, as well as enderbitic-trondhjemitic orthogneisses and meta-gabbro-norite (Ibicaraí Complex). These rocks were intruded by rocks of the Pau Brasil suite, which include charnockites, monzo-diorites, granites, quartz-monzonites and tonalities, the Anuri Syenite, and diabase dykes (Souto et al., 1972). The Itapetinga Complex is composed of syeno-granitic to tonalitic orthogneisses, intruded by anorogenic granitoids that crop out in the southern region of the basin.

The basal unit of the Rio Pardo Group, the Panelinha Formation, is observed as discontinuous outcrops in the northern, northwestern and western borders of the basin and its lower contact is non-conformable on the

<table>
<thead>
<tr>
<th>Age</th>
<th>Stage; environment and/or process</th>
<th>Stratigraphic unit and/or structural feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt;900 Ma</td>
<td>Early continental rift; pre-glacial, fluvial</td>
<td>Diamictite-free, lowermost proximal Macaúbas Group</td>
</tr>
<tr>
<td>ca. 875 Ma</td>
<td>Continental rift; pre-glacial, fluvial to marine</td>
<td>Salto da Divisa anorogenic suite; Diamictite-free proximal Macaúbas Group</td>
</tr>
<tr>
<td>&lt;875 Ma</td>
<td>Late continental rift; mainly gravitational glacio-marine sedimentation and transitional mafic volcanism</td>
<td>Diamictite-rich and diamictite-turbidite sequences with transitional basalts of the proximal Macaúbas Group</td>
</tr>
<tr>
<td>&lt;864 Ma</td>
<td>Proximal passive margin; continental shelf, marine transgression</td>
<td>Diamictite-free, sand-pellet unit of the uppermost proximal Macaúbas Group</td>
</tr>
<tr>
<td>ca. 816 Ma to ?</td>
<td>Distal passive margin and ocean spreading; continental slope to ocean floor</td>
<td>Ribeirão da Folha Formation, Dom Silvério Group and mafic-ultramafic oceanic rocks (ophiolite slivers)</td>
</tr>
<tr>
<td>&gt; 630 Ma</td>
<td>Start of oceanic lithosphere subduction from west to east</td>
<td>Nucleation of north-verging structures (cf. nutcracker tectonics model; see text)</td>
</tr>
<tr>
<td>630—585 Ma</td>
<td>Pre-collisional; continental magmatic arc and related basins</td>
<td>G1 suite, Rio Doce Group, Nova Venêcia Complex and Salinas Formation</td>
</tr>
<tr>
<td>585—560 Ma</td>
<td>Syn-collisional; fold-and-thrust tectonics, syn-kinematic metamorphism and partial melting</td>
<td>G2 suite and migmatization of other units; west- and east-verging thrust-and-folding with low-angle dip shear zones</td>
</tr>
<tr>
<td>560—500 Ma</td>
<td>Late- to post-collisional; adiabatic post-kinematic S-type anatexis, partially coeval to lateral escape to the south</td>
<td>G3 and G4 suites; NNE to NE-trending, high-angle dip, dextral, strike-slip shear zones controlled the lateral escape</td>
</tr>
<tr>
<td>520—490 Ma</td>
<td>Post-collisional; gravitational collapse, rising of mantelic magma and partial melting of deep crust</td>
<td>G5 suite; extensional sites along NE and NW trends of lineaments</td>
</tr>
</tbody>
</table>

Table 3.2 Evolution of the precursor basin and orogenic stages of the Araçuaí orogen, southeastern Brazil (age references in text).
Figure 3.14  Simplified geological map of the Rio Pardo basin, eastern São Francisco Craton (modified from Pedreira, 1999).

Figure 3.15  Stratigraphic chart for the Rio Pardo Group (modified from Pedreira, 1999).
crystalline basement. Its upper contact is with sedimentary rocks of the Camacan, Água Preta, Serra do Paraíso and Salobro formations (Figure 3.14). The Panelinha Formation is about 200 m thick (Karmann et al., 1989) and its lower portion is composed of immature breccias with coarse-grained matrix, superposed by polymeric metaconglomerates with pebbles of basement rocks, including syenite gneisses (Pedreira et al., 1969). These coarse clastics grade upwards into massive and laminated metagreywackes with conglomerate intercalations and disseminated pyrite, followed by metarkoses with conglomeratic intercalations; the metarkoses show parallel lamination, cross-bedding and graded bedding.

The Itaimbé Subgroup (Figure 3.14) was proposed by Karmann et al. (1989) to encompass the Camacan, Água Preta, Santa Maria Eterna and Serra do Paraíso Formations, whose contacts are lateral and separated by possible unconformity from the Panelinha Formation and unconformably overlain by the Salobro Formation.

The Camacan Formation (Pedreira et al., 1969) crops out in the northern border of the basin, along a belt between the towns of Camacan and Santa Luzia, and has a thickness of 200 m. The northern limit of the belt is the São Pedro River and the southern one is found in the Chororão, Rochedo and Lapão ranges, in the domain of the Salobro Formation (Figure 3.14). The lower and upper contacts of the Camacan Formation are unconformable, respectively, with the Panelinha and Salobro formations. The Camacan Formation is composed of metasiltstones and slates, argillaceous metalimestones and dark-grey dolomites, pyrite-bearing metagreywackes and fine-grained sandstones. The base of the formation consists of limestones and dolostones with intraclasts, followed by a thick interval where fine-grained sandstones alternate with argillite, shale and slate. Approximately in the middle part of this formation, a bed of greywacke separates this lower succession from an upper one of similar composition. Costa Pinto (1977) suggested a transition from continental environment, as evidenced by evaporation of the base of this formation, to open and then restricted marine, during which time carbonates were deposited.

The Água Preta Formation (Pedreira et al., 1969) crops out in the central part of the basin, in an area limited by the town of Pau Brasil and the villages of Gurupá-Mirim, Itaimbé and Teixeira do Progresso, as well as in erosional windows, east and northeast of Teixeira do Progresso village. The northeastern contact of the Água Preta Formation with the Salobro Formation is represented by a contractional fault (Rio Pardo–Água Preta Fault). The minimum thickness of the Água Preta Formation is of about 2,300 m.

In the Água Preta Formation, laminated phyllites and metasiltstones predominate, followed by micaceous and calcareous sandstones; subordinate argillaceous and dolomitic metalimestones and metadolostones also occur. This formation contains a few lenses of carbonate rocks southeast of the town of Pau Brasil. Costa Pinto (1977) analysed C and O isotopes in one of these lenses and found isotopic signatures similar to those in the Serra do Paraíso Formation.

The Santa Maria Eterna Formation (Lima et al., 1981) consists of massive dolomitic marbles with intercalations of grey metadolostone and micaceous quartzite. It crops out in the southeastern portion of the basin and is mostly overlain by the Neogene Barreiras Formation. It is in fault contact with the Água Preta Formation and interfingers with the Serra do Paraíso Formation (Karmann et al., 1989). Owing to the lack of continuous outcrops, its thickness has never been estimated.

The Serra do Paraíso Formation (Pedreira et al., 1969) makes up the southern and western borders of the basin. It is bound below by the basement and the Água Preta Formation. The former contact is generally in the form of reverse faults while the latter is gradational and interfingered. In the type-section, south of Itaimbé village, the formation is 1,575 m thick. North of the town of Pau Brasil, this formation consists of metadolomites and calcareous metadolomites. Tepee structures occur close to its base, indicating subaerial exposure. On the road between Pau Brasil and the Pardo River, it consists of micaceous metalimestones, metasiltstones and phyllites. In the region of the Serra do Paraíso–Gurupá Mirim, it is predominantly sand-carbonate metarhythmites, metalimestones and metarenites, while along the BR-101 road, it begins with calc-schists at the base, followed by alternating metalimestones, metadolomites and quartzites.

The Salobro Formation crops out in the northeastern sector of the Rio Pardo Basin, in an area of high relief formed by ranges such as the Chororão, Rochedo, Pacuipe, and Lapão. Its contact to the southwest with Água Preta Formation is a reverse fault, with tectonic transport to northeast (Rio Pardo–Água Preta Fault). The northern contact is a significant unconformity with the Camacan Formation. To northeast, the Salobro Formation is covered by the Barreiras Formation. Karmann et al. (1989) determined a thickness of about 5,000 m.

The Salobro Formation consists of immature metarenites, fine- and coarse-grained greywackes, metarkoses, metasiltstones, slates and metaconglomerates (both clast- and matrix-supported) rhythmically interstratified. Conglomerate lenses occur throughout this formation, the thickest being the Lapão conglomerate, which crops out in the northeastern sector of the formation, east of the town of Santa Luzia, and was formerly washed for diamonds. It is poorly sorted, with clasts whose diameter ranges from 1 to 50 cm; the clasts are from the basement (granulites, syenite gneisses), chert, metadolomite and metalimestone. The other conglomerate lenses have similar
composition, with clasts of phyllites and fine-grained quartzites. The Serra do Lapão conglomerate, in the lower Salobro Formation, occurs south of the town of Santa Luzia and comprises metadolostone pebbles with a calcarenite matrix.

The source for the Salobro Formation sediments was represented by two areas of high topographic relief (granulites and syenites from the north and sedimentary rocks from the south). The burial must have been fast, as indicated by the immaturity of the sediments. The interfingering facies of these conglomerates occurred on tectonic ramps, as demonstrated elsewhere by Steidtmann and Schmitt (1988).

The sedimentation in the Rio Pardo Basin started with the deposition of alluvial fans in graben-type basins, limited by extensional faults. The alluvial fans were formed by the conglomerates and breccias of the Panelinha Formation (Figure 3.14). Distally, these fans grade to fluvial systems (Figure 3.15) with abandoned channels and swampy areas evidenced by horizons of conglomerate intercalated with pyritic greywackes in the upper part of the formation.

Rifting ceased before development of oceanic crust, and the Camacan, Água Preta, Serra do Paraíso and Santa Maria Eterna formations (Itaímbé Subgroup) were deposited on the Panelinha Formation under marine conditions during the subsequent thermal subsidence phase. In the deeper parts of the basin, turbidites of the Água Preta Formation were deposited. Carbonate lenses (Mutuns creek, Nancy, Gurupá Mirim and others) in gradational contact with these turbidites were probably deposited in shallower environments.

The Água Preta Formation grades into Camacan, Serra do Paraíso and Santa Maria Eterna formations towards the shallower parts of the basin. Evidence for shallow deposition is found in all of these units: ripple marks and features that indicate subaerial exposure, such as intraclasts and mud cracks, occur in the Camacan Formation; intraclasts and tepees are found in the Santa Maria Eterna Formation; and stromatolites at the base of the Serra do Paraíso Formation indicate deposition under inter- to subtidal conditions. These formations can be interpreted, respectively, to have been deposited in muddy tidal flat, littoral and carbonate platform environments (Figure 3.15). The balance between the siliciclastics influx and the shape of the coastline can account for the distinct facies within these formations. Facies variability is most pronounced in the Serra do Paraíso Formation, whose depositional environment grades from outer platform to supratidal. Delgado et al. (2003) interpreted these formations as deposited on a passive margin during a retrograding depositional cycle.

After lithification of the limestones, sands and clays, tectonic reactivation uplifted both the basement north of the basin and the sedimentary rocks south of the Rio Pardo—Água Preta Fault. Both are source areas for the sediments of the Salobro Formation, as indicated by the pebble composition of conglomerates in the unit.

This brief description of the Rio Pardo Group demonstrates that the sedimentary environment and tectonic setting during deposition of the Rio Pardo Group sediments evolved from an intracontinental rift through a phase of thermal subsidence and into a foreland basin.

### 3.2.2.4. The Sergipano belt

The Sergipano is a Brasiliano/Pan-African fold-and-thrust belt (Figures 3.11–3.13) located on the northeastern edge of the São Francisco Craton. The structure and lithology of this belt were compared to the Ndjolé Series of northern Gabon by Allard and Hurst (1969). Later, Cordani (1973) proposed a comparison with the sequences of Mbalmayo-Bengbis, Djä and Sembe Ouesso of southern Cameroon and the extreme north of the Congo. According to Trompette (1994), the Sergipano belt represents the Brazilian continuation of the Ouandigues and, together, they form a roughly E-W elongated mega-orogen more than 5,000 km long. More recently, the Sergipano basin has been correlated with the Yaoundé belt (Cameroon, Africa) by Oliveira et al. (2006) based on palaeocontinental reconstructions, lateral variation of rock units and tectonic histories.

This belt was formed by continental collision between the Congo—São Francisco Craton and the Pernambuco-Alagoas massif during the Brasiliano/Pan-African orogeny (Brito Neves and others, 1977). Initially, it was interpreted as a geosyncline (Humphrey and Allard, 1968; Silva Filho and Brito Neves, 1979) and later as a collage of lithostratigraphic domains (Davison and Santos, 1989; Silva Filho, 1998) or as a Neoproterozoic fold-and-thrust belt produced by inversion of a passive margin basin located at the northeastern edge of the ancient São Francisco plate (D’el Rey Silva, 1999).

The São Francisco Craton, to the south of the Sergipano belt, is a granite-greenstone belt sequence with high-grade terranes with ages ranging from 3,400 to 2,080 Ma (Oliveira et al., 2006). The Pernambuco-Alagoas massif, to the north of Sergipano belt, comprises Palaeoproterozoic to Mesoproterozoic high-grade gneisses and migmatites intruded by Neoproterozoic granitoids (Brito Neves et al., 1982; Silva Filho et al., 2002). The Sergipano belt consists, from north to south, of six lithostratigraphic domains (Figures 3.16–3.18) separated by major shear zones: Canindé, Poço Redondo, Marancó, Macururé, Vaza Barris and Estância (Santos and Souza, 1988; Davison and Santos, 1989; Silva Filho, 1998). The Macururé, Vaza Barris and Estância are composed mostly of metasedimentary rocks with metamorphic grade, varying from unmetamorphosed in the Estância domain, through greenschist grade in the Vaza Barris, to amphibolite facies in the Macururé domain.
Figure 3.16 Domains of the Sergipano belt (modified from D’el Rey Silva, 1995). MSZ, BMJSZ, SMASZ and ISZ represent, respectively, the Macururé, Belo Monte-Jeremoabo, São Miguel do Aleixo and Itaporanga shear zones.

Figure 3.17 Simplified geological map of the eastern portion of the Sergipano belt (modified from D’el Rey Silva, 1999).
Silva Filho and Torres (2002) and Silva Filho et al. (2003) suggested three additional domains: Rio Coruripe, Viçosa and Pernambuco–Alagoas. It is possible that the Rio Coruripe domain, metamorphosed to granulite facies and later retrograded to amphibolite and greenschist grade, is a high-grade counterpart of the Macurure´ domain, judging from the lateral continuity of similar rock types (Oliveira et al., 2006). All these domains, except for the Estaˆncia and Vaza Barris, were intruded by Brasiliano/Pan-African granitoids. Additional descriptions of all of the domains in the Sergipano can be found in Davison and Santos (1989) and Santos et al. (1998), and a summary on the geochronological data, including recent geochronology, is found in Oliveira et al. (2006).

The Caninde´ domain (Figure 3.16) comprises (a) an elongated pink granite sheet (Garrote unit); (b) a metavolcano-sedimentary sequence (Novo Gosto unit) represented by fine-grained amphibolite, marble, graphite schist, micaschist and metagreywacke; (c) subvolcanic microgabbro–quartz diorite complex (Gentileza unit); (d) the Caninde´ gabbro–leucogabbro complex (gabbro, gabbronorite, peridotite and pegmatitic gabbro) and (e) granitic plutons (tonalites, granodiorites and quartz-syenites). The ages of the rocks in this domain range from about 715 to 609 Ma (Silva Filho et al., 1997; Nascimento et al., 2005; Oliveira et al., 2006).

The Pocô Redondo domain (Figure 3.16) is composed of biotite gneisses, migmatises and granitic intrusions, such as the Serra Negra augen-gneisses, the Sítios Novos granodiorite and leucogranite sheets. Grey gneisses from the palaeosome of migmatites yielded ages slightly older (960–980 Ma U-Pb SHRIMP ages; Carvalho et al., 2005) than the Serra Negra augen-gneisses (952 Ma U-Pb SHRIMP age; Carvalho et al., 2005). The Sítios Novos granodiorite has been dated at 651 ± 11 Ma (U-Pb TIMS; Carvalho et al., 2005). The Marancó domain is mainly formed of a metavolcano-sedimentary sequence (quartzite, conglomerate, micaschist, phyllites and lenses of andesite, dacite and quartz porphyry) with peridotite and amphibolite lenses (Figure 3.16). The presence of andesites and dacites, dated at about 603 Ma (Carvalho et al., 2005) and conformably interleaved with phyllites, and of granodiorite stocks dated at 595 ± 11 Ma (Silva Filho et al., 1997), points to an Ediacaran
depositional age. The Macururê domain consists mainly of garnet micaschists and phyllites (Figure 3.16) with some quartzite and marble. These metasedimentary rocks were intruded by granitic plutons dated at about 625–630 Ma (Bueno et al., 2005; Long et al., 2005).

The Vaza Barris domain (Figure 3.16) shows two cycles of sedimentation, both with a continental to shallow-marine, basal siliciclastic megasequence, overlain by a carbonate sequence (D’el Rey Silva, 1995, 1999). The lower siliciclastic megasequence is represented by the Juetê (sandstone, diamictite; Figure 3.19), Ribeirópolis (silty phyllites, metagreywackes, pebbly phyllites, diamictites; Figure 3.20) and Itabaiana (conglomerate, quartzite, metasiltite) formations. The lower carbonate megasequence is known as the Jacoca Formation, which is stratigraphically equivalent to the Acauã Formation. These two megasequences form the Estância-Miaba Group. The upper siliciclastic megasequence (Simão Dias Group) is represented by the Lagarto-Palmares, the Jacaré and the Frei Paulo formations and is overlain by the Vaza Barris Group that comprises diamictites of the Palestina Formation and the upper carbonate megasequence, the Olhos D’Água Formation. All of these rocks spread continuously across the craton margin into the Sergipano belt where they occur around the Itabaiana and Simão Dias basement domes and are overlain by a metadiamictite and metacarbonate belonging to the upper megasequence. According to D’el Rey Silva (1999), both the basement and cover experience the same Neoproterozoic compressive deformation under subgreenschist facies.

The thick Olhos D’Água Formation is composed of a sequence of marbles and interbedded green, calcareous chlorite schists and silty phyllites; it overlies diamictites and pebbly metagreywackes of the Palestina Formation.

Figure 3.19 Diamictite of the Juetê Formation, overlain by carbonates of the Acauã Formation, at Serra da Borracha, about 20 km west from Patamuté village, state of Bahia.
Figure 3.20  Diamictite of the Ribeiropolis Formation in sharp contact with dolostones of the Jacoca Formation, Capitão Farm, state of Sergipe. Dolostone shows wavy contact and some bedding is observed in the diamictite. Some clasts seem to be concentrated and salient at the top of diamictite, which suggests lithification and erosion prior to deposition of the dolostone.

Figure 3.21  Pebbly metagreywacke of the Palestina Formation, sometimes with dropstones (about 12 km from Rosario village, state of Bahia).

(Figure 3.21). Marble beds are interbedded with blue to black, fine-grained metalimestone and grey metadolostone. The thick carbonates around the Simão Dias dome (Figure 3.17) pass upward into supertidal—intertidal facies with oolites, and wave-reworked structures indicating a near-shore environment (D’el Rey Silva, 1995). The Jacoca Formation (older) and Olhos D’Água Formation (younger) carbonates have been interpreted as cap carbonates by Sial et al. (2000a, 2006).

An acidic tuff within the Ribeiropolis diamictites in an outcrop on the Ribeiropolis-Gloria road, 7 km from Ribeiropolis town, was dated at around 730 Ma (zircon, U-Pb; B.B. Brito Neves, written communication), suggesting that the deposition of the overlying Jacoca carbonates happened right after the Sturtian glaciation event. The youngest U-Pb age of zircon grains (615 Ma) from the Frei Paulo metagreywackes above the Olhos D’Água carbonates constrains the deposition of Jacoca and Olhos D’Água to the interval 730—615 Ma.

The Estância domain (Figure 3.16) comprises, from base to top, sandstones and argillites of the Juêté Formation, dolostones and limestones of the Acauã Formation, sandstones and conglomerate lenses of the Largato Formation, and sandstone and minor conglomerate lenses of the Palmares Formation (Silva Filho and Brito Neves,
1979). Although D’el Rey-Silva (1999) assumed that deposition in the Estância and Vaza Barris domains occurred during two cycles of sedimentation on a passive continental margin of the ancient São Francisco plate (detritus sources from the São Francisco Craton), Oliveira et al. (2006) proposed instead deposition in foreland basins, based on detrital zircon populations with ages between 570 and 657 Ma and an inferred provenance from the Sergipano belt and Borborema province to the north.

3.2.2.5. The Riacho do Pontal and Rio Preto belts

The Riacho do Pontal and Rio Preto belts border the northern portion of the São Francisco Craton (Figure 3.1) and are part of the southern portion of the Borborema Province (Brito Neves, 1983; Brito Neves et al., 2000). The Riacho do Pontal belt consists of the southern frontal zone of the Borborema Province that overthrusts directly onto the basin of the São Francisco Craton and represents a subautochthonous metasedimentary sequence (Jardim de Sá et al., 1992; Oliveira, 1998a). The Rio Preto belt is located at the northwestern border of the São Francisco Craton and is interpreted as an intracontinental belt (Egydio-Silva, 1987; Egydio-Silva et al., 1990; Andrade Filho et al., 1994; Trompette, 1994).

The Riacho do Pontal belt stretches from the southern region of the states of Piauí and Pernambuco to the northern region of the state of Bahia. This belt can be divided into four zones or structural domains, from south to north: (a) cratonic domain, (b) southern zone, (c) central zone and (d) internal or northern zone (Figure 3.22). The cratonic domain consists of biotite-hornblende gneisses, migmatites, gneissic batholiths of granodioritic-tonalitic composition and of patches of metasedimentary rocks. The rocks in this domain are Archaean to Palaeoproterozoic in age, but affected by the Pan-African/Brasiliano cycle. The boundary between the Riacho do Pontal belt and the São Francisco Craton is somewhat unclear in this domain. A southward thrust overriding the craton was part of a Pan-African/Brasiliano tectonic event that also deformed carbonates of the Una Group (Alkmim et al., 1993; Brito Neves, 1983).

The southern zone is represented by a thrust-fold belt with southern vergence onto the São Francisco Craton (Angelim, 1988; Gomes, 1990; Jardim de Sá et al., 1992; Oliveira, 1998a,b). The rocks that form the Riacho do Pontal belt are variably referred as the Salgueiro or Casa Nova groups, depending on the location. These rocks are terrigenous-shelf metasedimentary rocks represented by garnet micaschists, quartzites, marbles and gneisses. Deep-sea, flysch-like metagreywackes with carbon and/or feldspathic clasts are regarded as turbidites. Syn-tectonic granitoids are interwoven into the metasedimentary rocks, as are foliated sheets of two-mica monzogranitic tonalites, known as Rajada orthogneisses. The syn-tectonic Serra da Boa Esperança granite and quartz-syenite yielded an age of 555 ± 10 Ma (Rb-Sr, Pb-Pb; Jardim de Sá et al., 1996). The WSW-ENE foliation shows NNW-SSE stretching lineation and sheath folds with axes subparallel to that lineation and resulted from the same event that has generated the nappes. The thrust deformation induced a decollement of the metasedimentary sequence over the cratonic massif (Jardim de Sá et al., 1992).

The central zone consists of metabasalts, amphibolites, chlorite-, talc-, tremolite- and actinolite-bearing rocks regarded as ultramafic intercalations in schists, and metachert. This zone contains the Monte Orebe and Brejo Seco mafic–ultramafic massifs with metasedimentary intercalations (Marimon, 1990; Moraes, 1992). The mafic rocks are Neoproterozoic tholeiitic ocean-floor basalts (Moraes, 1992), believed to be an ophiolitic complex.

![Figure 3.22](image-url) Cross-section for the Riacho do Pontal belt that can be divided into four zones or structural domains, from south to north: (a) cratonic domain, (b) southern zone, (c) central zone and (d) internal or northern zone (Uhlein et al., 2008).
The internal zone is bounded by ductile right-lateral shear zone and corresponds to a tectonically complex region where subvertical mylonitic foliation shows recumbent S structures with right-lateral movements. This zone consists of remnants of Brasiliano metasedimentary sequences, intrusive granitoids and slices from the basement that form stretched amygdaloidal bodies. The mylonitic zone is up to 1 km thick with proto-mylonites through ultramylonites (Vauchez and Egydio-Silva, 1992; Vauchez et al., 1995).

The Rio Preto belt is considered a marginal fold belt forming a westward extension of the Riacho do Pontal and Sergipano belts. This belt is of great structural and lithostratigraphic interest because it displays a transition between cratonic facies, in the south, to metasedimentary rocks at the central part, and gneisses of the basement towards the northern border. The metasedimentary sequences are equivalent to the São Francisco Supergroup in the state of Bahia.

From south to north, the Rio Preto belt can be subdivided into three structural units, each with distinct lithologic, structural and metamorphic characteristics, separated by disrupted structures (Figure 3.23). In the southern unit, in the portion of the São Francisco Craton and adjacent to this belt (the São Desidério region), the sequence begins with microcrystalline limestone with argillaceous intercalations, characterised by 50 m-thick horizontal layers, and shows no signs of metamorphism or deformation (cratonic domain). This unit was named the São Desidério Formation (Egydio-Silva, 1987; Egydio-Silva et al., 1989) and probably correlates with the Sete Lagoas Formation of the Bambuí Group in Minas Gerais (Figure 3.24).

To the north, cratonward from the fold belt, these limestones are overlain by clastic rocks with intercalations of marls and limestones. This sequence (Serra da Mamona Formation) has been correlated with the Santa Helena Formation (Egydio-Silva, 1987; Egydio-Silva et al., 1989). North of Barreiras, they are overlain by meta-arkoses, fine- to medium-grained quartzites, locally feldspathic, metagreywackes and rare carbonate intercalations. This sequence (Riachão das Neves Formation) is considered partially equivalent to the Três Marias Formation of the Bambuí Group (Egydio-Silva, 1987; Trompette, 1994).

In the central portion of the belt, the lower part of the succession is more complete with the occurrence of boulder diamictites, with clasts up to 50 cm in diameter of gneiss, quartzite, metasiltstone, marble and schist. Referred here as the Canabravinha Formation (Figure 3.25), this unit has been correlated to the Bebedouro Formation, although a glaciogenic origin remains unsubstantiated. Further north, in the region of Formoso do Rio Preto (Figure 3.23), the Rio Preto Group contains predominantly micaceous quartzites and schists with a few intercalations of amphibolites and itabirites. It has been correlated with the Chapada Diamantina Group (Inda and Barbosa, 1978).

The general structure of the Rio Preto belt is of an asymmetric fan, probably associated with late ENE–WSW shearing that also occurred in the Borborema Province. Structural analysis of this belt defined three structural units (Egydio-Silva, 1987; Egydio-Silva et al., 1990; Andrade Filho et al., 1994). The central and southern ones, well developed between the craton and the Cariparê thrust, show a single deformation phase with southern vergence.
towards the São Francisco Craton. The northern unit is metamorphosed to the greenschist facies, is polyphase and is thrust northwards over the Piauí basement. The absence of granitic intrusions and rare volcanism favour an ensialic origin for this belt (Egydio-Silva, 1987; Trompette, 1994).

The Riacho do Pontal belt appears to have participated in the collision between the São Francisco Craton and the Pernambuco-Alagoas massif. The basin can be schematically reconstructed with terrigenous-shelf sediments and deep-sea sediments (turbidites) in a passive margin context. Mafic–ultramafic massifs (Monte Orebe and Brejo Seco) may be remnants of a narrow oceanic crust-floored basin (Oliveira, 1998a,b). The Brasiliano collisional deformation is complex, with thrusts towards the south over the São Francisco Craton associated with greenschist to amphibolite metamorphism (Jardim de Sá et al., 1992). In the internal zone, a continental-scale subvertical dextral ductile shear zone is the main structural feature.

The Rio Preto belt is a good example of an intracontinental fold belt whose central structure seems to have been controlled by the geometry of a Mesoproterozoic trough and by the intervention of late shears. The sedimentary basin, as reconstructed, is composite, with a northern trough of Mesoproterozoic age filled by the Rio Preto Group and juxtaposed against a younger subsiding southern basin. In this younger basin, a mostly clastic and thickened equivalent of the late Proterozoic Una Supergroup was deposited. The migration of the depocenter of the basins towards the south implies the intervention, before deposition of the Bebedouro Group, of
epeirogeny uplifting northern highlands against which the Neoproterozoic basin was juxtaposed (Egydio-Silva, 1987; Trompette, 1994; Andrade Filho et al., 1994).

### 3.3. Isotope Chemostratigraphy

C and O isotope studies were performed across different sections of some sequences in the southwestern portion of the São Francisco Craton (Santos et al., 2000, 2004; Alvarenga et al., 2007). The most complete profile is at the Serra de São Domingos region, where over 700 m of Bambuí Group sediments was deposited atop siliciclastic rocks of the Paraná Group. These two groups are separated by a karstic limestone breccia infilled with well-sorted sand grains that may be the lateral equivalent of Jequitão Formation diamictites. In the base of this section, dolostones of the upper Paraná Group have δ13C values close to 0‰ (VPDB), similar to values observed in other areas where this unit is exposed. The Paraná—Bambuí transition is characterised by a spike of low δ13C values (~−5‰), followed by limestones with slightly positive values. The uppermost dolostones of the Conselheiro Mata Group (Espinhaço Supergroup) and dolostones of the Rio Pardo Grande Formation (lowermost portion of the Macaúbas Group) display δ13C values around 0‰ (Figure 3.26; Santos et al., 2000), typical of Mesoproterozoic to early Neoproterozoic carbonates. The most remarkable aspect of these profiles is a positive δ13C excursion in the upper Sete Lagoas Formation that continues through the Serra de Santa Helena and Lagoa do Jacaré Formation limestones (Figure 3.26). This excursion, identified across different profiles of the northern and southern reaches of the Bambuí basin and used to reconstruct the palaeogeography of the Bambuí basin (Santos et al., 2004), is a robust stratigraphic marker related to a Neoproterozoic positive carbon isotope excursion recognised worldwide.

A sedimentological and C—O isotope study carried out by Vieira et al. (2007) in nine sections of the Sete Lagoas cap carbonate sequence at its classical outcropping area, in the southern tip of the São Francisco Craton, helped refine its stratigraphy. According to these authors, the Sete Lagoas Formation comprises two shallowward-upward megacycles, each cycle bound by a flooding surface and amalgamated with a third-order sequence boundary. The first megacycle is represented by deep-platform deposits with abundant crystal fans (aragonite pseudomorphs) that show negative δ13C values (−4.5‰) that grade upward to storm wave and tide-influenced layers with values around 0‰. A second megacycle is characterised by a thick, mixed substorm wave-base succession deposited during another transgression that drowned the platform, with lime mudstone—pelite rhythmites that grade to crystalline limestone rich in organic matter with δ13C values up to +14‰.

Vieira et al. (2007) proposed that the very high δ13C values observed in the second megacycle, coupled with geochronological data, support correlation with post-Sturtian sequences. Some differences in the depositional record for the Sete Lagoas and other post-Sturtian units described in North America, Australia and Namibia are perhaps due to the deposition of the Sete Lagoas carbonates in shallower settings, preserving a thick record of storm- and wave-influenced sedimentation, not found elsewhere. Alternatively, these differences may be ascribed to diachronous deposition of the post-Sturtian cap carbonate sequences.

Chemostratigraphic studies of the Una Group have been carried out by Torquato and Misi (1977), Misi and Kyle (1994) and Misi and Veizer (1998). Sulphur, carbon and oxygen isotope analysis, as well as Sr isotope determinations, confirmed the evolution of the sedimentary sequences as proposed by Misi and Souto (1974) and by Misi (1979). In the study by Misi and Veizer (1998), samples for isotope analysis were collected along the eastern border of the Irêcê “basin” covering the stratigraphic section of carbonate lithofacies from the laminated limestones of Unit B to the black oolitic limestones of Unit A1. Based on petrographic studies and trace element determinations, the best preserved samples (micrites, with Mn/Sr < 0.09 and Sr > 300 ppm) were selected for isotope analysis. For the best preserved samples, there is a progressive increase in δ13C (+0.3‰ to +8.7‰), decline in 87Sr/86Sr (0.70789−0.70745) and a relatively uniform δ18O signal (−5‰ to −7.5‰ VPDB) upsection (Figure 3.27).

Previous C and O isotope studies by Torquato and Misi (1977) of 63 whole-rock carbonate samples from the same stratigraphic section demonstrated the same increase in δ13C values in the upper section (Unit A1, ca. +8‰) and showed significant negative shifts in the pinkish dolostones of Unit C, in the lower section (around −6‰; n = 12). Considering only the best preserved samples and considering the data of Torquato and Misi (1977) from the same section, it is possible to show the variation of the isotopic composition through the Una Group (Figure 3.27).

δ18O/δ13C variation, when compared with the seawater evolution curves through the Neoproterozoic (Jacobsen and Kaufman, 1999), suggests an age of sedimentation between 750 and 600 Ma. δ13C variation, with a negative shift at the base and a high positive excursion in the upper section, is compatible with that observed in the Bambuí Group (Misi et al., 2007).

A preliminary C and O isotope study on carbonates of the Rio Pardo Group was carried out by Costa Pinto (1977), but as the 120 samples involved lack of stratigraphic control, they are not used in our compilations. Costa
Figure 3.26  C and O isotope profiles for carbonates of the Bambuí Group at Serra de São Domingos, southeastern portion of the São Francisco Craton and for carbonates of the Domingas Formation (Macaúbas Group) and Rio Pardo Grande Formation of the Espinhaço Supergroup (modified from Santos et al., 2004).
Pinto (1977) concluded that this sequence was deposited in a continental to restricted marine to open-marine sedimentary environment. Camacan Formation carbonates show $\delta^{13}C$ values from $-5.7\%$ to $+4.2\%$ and $\delta^{18}O$ values from $-10\%$ to $-5\%$, while carbonates from the Salobro Formation have $\delta^{13}C$ from $-2\%$ to $+2\%$ and $\delta^{18}O$ from $-19.7\%$ to $-8\%$. Agua Preta Formation carbonates display $\delta^{13}C$ values from $-1.3\%$ to $+6.2\%$ and $\delta^{18}O$ from $-11.7\%$ to $-5.8\%$, while carbonates of the Serra do Paraíso Formation yielded $\delta^{13}C$ values from $-1.7\%$ to $+8.9\%$ and $\delta^{18}O$ from $-8.4\%$ to $-1.8\%$. Although a true chemostratigraphic profile is not available for these carbonates, the data imply large $\delta^{13}C$ fluctuations for the Camacan ($-5.7\%$ to $+4\%$) and the Serra do Paraíso ($-1.7\%$ to $+8.9\%$) formations, of a similar magnitude to those observed in cap carbonates. The Serra do Paraíso organic-rich limestones exhibit $\delta^{13}C$ variation somehow similar to that of Olhos D’Água Formation carbonates in the Sergipano belt. C and O isotope values suggest a transition from continental to marine environment for the Camacan (Costa Pinto, 1977). These marine environments continued southwards as expressed in the Água Preta and Serra do Paraíso formations, suggesting deposition of the Rio Pardo Group on a passive margin.

Thirty samples from the Serra do Paraíso Formation at Toca da Onça locality ($Mg/Ca \sim 0.5-0.6$) show $\delta^{13}C$ values from $0.1\%$ to $+2.2\%$ and $\delta^{18}O$ from $-4.7\%$ to $-4.2\%$ (A.N. Sial, unpublished data) within the range reported by Costa Pinto (1977). Four dark, organic matter-rich, Mg-carbonates of the Agua Preta Formation (Nancy locality) display $\delta^{13}C$ values of $+4.1\%$ to $+8.2\%$ (A.N. Sial, unpublished data).

For carbonate formations of the Sergipano belt, Sial et al. (2000a, 2006) reported $\delta^{13}C$ values for the Jacoca Formation, from a section along the Vaza Barris River, of around $-4\%$ and $\delta^{18}O$ values around $-8\%$ (Figure 3.28). Strong oscillations are also observed in Mg/Ca ratios and Si, Fe, Mn and Rb concentrations in samples from the lowest portion of this profile. To the north of the town of Simão Dias, the Olhos D’Água Formation is represented by marbles, interbedded calcareous chlorite-schists and silty phyllites and overlies diamicites, pebbly metagreywackes of the Palestina Formation. Sial et al. (2006) produced a chemostratigraphic profile ($n = 120$)
starting from the contact between the Olhos D’Água Formation and metagreywackes of the Palestina Formation. Basal marly and dolomitic carbonates display $\delta^{13}C$ values of $-4\%$ to $-5\%$, increasing upsection ($0-1\%$). Towards the top, $\delta^{13}C$ changes dramatically to values of $\sim +8\%$ to $+10\%$ and the same behaviour is observed in a section at Serra do Capitão (Figure 3.29) and in another section near Rosario village in Bahia. This vigorous isotope fluctuation suggests contrasting, base to top, environmental conditions during deposition of this sequence, allowing for an enormous C-isotope oscillation, typical of Sturtian cap carbonates (Kennedy, 1996; Hoffman and Schrag, 2002). $\delta^{18}O$ values vary from $-7\%$ to $-10\%$, a typical for Cyrogenian carbonates.
Sial et al. (2006) reported a C isotope profile for dolostones of the Acauã˜ Formation (stratigraphically equivalent to the Jacoca Formation) overlying diamictites of the Jueteˆ Formation at the Sã³o Gonçalo Farm near the town of Euclides da Cunha in the state of Bahia. These dolostones have laminations whose thickness gradually decreases upsection, indicating deposition during a transgression. The diamictite contains boulders of granite, orthogneiss, phyllite, quartz, some of which are dropstones, and an iron-rich claystone layer within a matrix-supported granular wackestone. Values of $\delta^{13}C$ start around $-5\%$ and gradually increase to $-3\%$ at about 15 m in height, while $\delta^{18}O$ varies from around $-4.5\%$ to about $-6.5\%$ upsection. At Serra da Borracha locality, about 20 km west of Patamute village, a similar C-isotope profile is observed in a carbonate section that overlies 1-m-thick diamictite bed resting on the basement. $\delta^{13}C$ values are around $-5\%$ while $\delta^{18}O$ values are around $-10\%$ (Figure 3.30).

In summary, in proximal or distal sections of the Jacoca/Acauã˜ carbonates, $\delta^{13}C$ values are virtually always negative, mostly clustering around $-5\%$, whereas three sections of the Olhos D’Água carbonates show distinct C-isotope profiles, with strong fluctuations between negative ($-5\%$) and highly positive ($+8\%$ to $+10\%$) values.

The Jacoca and Olhos D’Água formations display similar $^{87}Sr/^{86}Sr$ ratios (on average, 0.707974 and 0.707922, respectively) in the eastern portion of this belt. The $^{87}Sr/^{86}Sr$ ratios for Acauã˜ carbonates at Borracha Hill vary from 0.70717 to 0.70732, at Patamute they range from 0.70751 to 0.70937, and at Sã³o Gonçalo Farm, from 0.70942 to 0.71099. For the Olhos D’Água carbonates from the western portion of the belt, ratios vary from 0.70753 to 0.70828. The average $^{87}Sr/^{86}Sr$ for Olhos D’Água Formation carbonates is in good agreement with values reported for seawater at ca. 720Ma (Jacobsen and Kaufman, 1999). The average $^{87}Sr/^{86}Sr$ for the Jacoca Formation, however, is slightly too high for the seawater at ca.740 Ma. The Sr temporal variation curve compiled by Jacobsen and Kaufman (1999), however, is not well constrained for the interval around 740 Ma.

### 3.4. GEochRONOLOGY AND PROVENANCE: SYNTHESIS OF U-Pb AND Sm-Nd DATA

The Neoproterozoic successions deposited on the Sã£o Francisco Craton contain basal glaciogenic units (Jequitait Formation and correlative units), which are capped by carbonates and then followed by siliciclastic sediments of the Bambuí Group. The available geochronological constraints were obtained on detrital zircons recovered from the matrix of the glacial damicites that provide maximum depositional ages, and on the cap carbonates and overlying pelitic rocks of the Bambuí Group.
Nd isotope data indicate a Palaeoproterozoic source region for sediments of the Paranoá, Canastra and Vazante Groups (Pimentel et al., 2001). The Paranoá and Canastra Groups show the oldest Nd model ages that range between 2.3 and 2.1 Ga and, consequently, a provenance of clastic material from Palaeoproterozoic sources in the São Francisco Craton and passive margin tectonic setting. The Vazante Group, in turn, has yielded Nd model ages between 1.9 and 1.7 Ga, which may indicate a transition between the Paranoá and Bambuí Groups. Araxá and Ibiá sedimentary rocks present a bimodal Nd model age pattern (1.8–2.1 and 1.0–1.3 Ga). The Bambuí Group exhibits more uniform Nd model ages (1.4–1.9 Ga). The younger Nd model ages for these last units are interpreted as a juvenile source related to the Goiás magmatic arc, in the west of the Brásilia belt.

U-Pb SHRIMP ages obtained on detrital zircon grains recovered from sandstones and conglomerates of the Vazante Group have given ages from 2.3 to 0.95 Ga, while detrital zircon grains from arkoses and conglomerates of the Bambuí Group yielded younger ages of ca. 650 Ma (Dardenne et al., 2003).

### 3.4.1. Geochronology of the glacial deposits

U-Pb SHRIMP ages were obtained on detrital zircon recovered from a glacial succession from the Bebedouro Formation in the Irecê basin, Bahia (Figure 3.1). This unit starts with shale with dropstones, interbedded and overlain by diamictite. The succession is capped by carbonates. Detrital zircon from the laminated facies and the diamictite range in age from 3,050 to 875 Ma (n = 19), indicating multiple source terranes (Babinski et al., 2004). The older ages are from the Archaean to Palaeoproterozoic basement; Mesoproterozoic zircon grains may be either from the Espinhaça Supergroup volcano-sedimentary sequence or from the ca. 1,000 Ma dykes that cut both the basement and the Espinhaça rocks. The source of the youngest zircon grains has been recently identified and precisely dated at 875 ± 9 Ma (Silva et al., 2007). It represents the anorogenic Salto da Divisa granites, which record continental rifting in the Aráuá Orogen (Pedrosa-Soares et al., 2008).

Detrital zircons from diamictites of the Jequitai Formation, sampled at the Água Fria Range (Figure 3.1), were dated by Pb-Pb evaporation and yielded a maximum age of ca. 900 Ma (Buschwaldt et al., 1999). Correlation of glacial deposits from surrounding belts on the craton provides additional, albeit interpretive, age constraints. U-Pb SHRIMP ages were reported for detrital zircon recovered from sandstones from the uppermost part of the Chapada Açu Formation (Macuábas Group), a post-glacial shelf succession of interlayered sandstones and siltites (Pedrosa-Soares et al., 1992; Noce et al., 1997). Most detrital zircon grains yielded ages between 1,000 and 950 Ma, but the almost concordant youngest zircon grain constrains the maximum sedimentation age for the upper Chapada Açu Formation at 864 ± 30 Ma (Pedrosa-Soares et al., 2000a). U-Pb SHRIMP age determinations of granitic pebbles from the Santo Antônio do Bonito (Vazante Group) and Cubatão (Ibiá Group) diamictites have given Palaeoproterozoic ages (2.3 Ga), providing no useful age constraints, but indicating a cratonic source for these formations (Dardenne et al., 2003).

A recent geochronological study on detrital zircon from different units of the Macuábas Group, including pre-glacial and glacial rocks, allowed identification of source areas with ages ranging from 900 to 2,740 Ma. The maximum depositional age for the glacial deposits was defined by the youngest detrital zircon dated at 900 ± 21 Ma (Babinski et al., 2007; Pedrosa-Soares et al., 2008). A U-Pb zircon age of 730 Ma (B.B. Brito Neves, written communication) on an acidic tuff within the Ribeiropolis diamictites underlying the Jacoca carbonates, coupled with U-Pb age of zircon grains (615 Ma) from Frei Paulo metagreywackes above the Olhos D’Água carbonates, constrains the deposition of Jacoca and Olhos D’Água Formations to the interval 730–615 Ma.

### 3.4.2. Geochronology of the Bambuí Group

Data for Bambuí Group sedimentary rocks were obtained by the Rb–Sr method on clays and whole-rock samples and by the K–Ar method on clays (Bonhomme, 1976; Thomaz Filho and Bonhomme, 1979; Parenti Couto et al., 1981; Bonhomme et al., 1982; Thomaz Filho et al., 1998). Rb–Sr ages range from 695 ± 12 Ma (R0 = 0.7077) to 560 ± 40 Ma (R0 = 0.7110) and K–Ar ages on fine fraction clays range from 662 ± 18 to ~478 Ma. The older ages were considered as minimum depositional ages, and the younger ones as a result of later thermal events related to the Brasiliano orogeny. A Pb–isotopic study on the carbonate rocks from the Sete Lagos and Lagoa do Jacaré formations was done by Babinski et al. (1999) and a minimum depositional age of 686 ± 69 Ma was suggested for these rocks. It was also suggested by those authors that a large-scale fluid percolation event affected most of the rocks from the southern part of the São Francisco basin at ca. 520 Ma. The same event could have been responsible for the remagnetisation of these carbonates (D’Agrella Filho et al., 2000; Trindade et al., 2004).

Recently, radiometric data were determined for deep-platform rocks organised in centimetre-scale cycles of lime mudstone and calcite crystal fans, interpreted as aragonite pseudomorph, that crop out at the Sambra Quarry,
near Sete Lagoas (Figure 3.1) at the southern part of the São Francisco Craton. These aragonite-pseudomorph crystals are black to dark-grey with acicular morphology, laterally connected to thin, millimetric cement-crusts. Crystal layers are covered by light-grey or red lime mudstones showing parallel to undulating lamination often truncated by stylolites. These carbonates yielded a Pb-Pb isochron age of $740 \pm 22$ Ma (Babinski et al., 2007). Because the rocks of this outcrop are extremely well preserved and undeformed, suggesting that the Pb-isotopic system was not disturbed by later events, this value was interpreted as the depositional age for this cap carbonate (Figure 3.31).

Vieira et al. (2007) have identified two depositional sequences in the Sete Lagoas Formation, the lower one essentially consisting of carbonates and the upper one of pelite-calcilutite in the base, overlain by black limestone. The age of the upper sequence is younger than 610 Ma as recently constrained by a U-Pb (LA-MC-ICP-MS) age for detrital zircon grains from siltstones (Rodrigues and Pimentel, 2008). Thus, geochronological constraints available for the Neoproterozoic rocks deposited on the craton suggest that while the lower Bambuí Group may record the Sturtian glacial event, the upper Bambuí Group is Ediacaran in age.

### 3.5. Synthesis of Neoproterozoic Mineralisations: Lead, Zinc, Fluorine, Barium and Phosphates

Neoproterozoic successions of the São Francisco Craton host a variety of mineral deposits, some of them of great economic importance. Among the known mineral deposits are the following:

(i) Rapitan-type Fe-Mn formation in the Porteirinha in the Rio Vacaria valley is associated with glacial sedimentation of the Macaúbas Group in the Araçuai belt, state of Minas Gerais. The source of iron and manganese is probably related to exhalative processes and mafic volcanism, as suggested by Dardenne and Schobbenhaus (2000).

(ii) Phosphorites deposits (Figures 3.32 and 3.33) are associated with the lower Vazante Group at Coromandel, Rocinha and Lagamar (Dardenne et al., 1986, 1997; Da Rocha Araújo et al., 1992; Misi et al., 2005a,b), to the base of the Sete Lagoas Formation of the Bambuí Group at Campos Belos (Dardenne et al., 1986), to the lower portion of the Serra da Saudade Formation of the Bambuí Group at Cedro do Abaeté (Dardenne et al., 1986) and to stromatolitic beds of the Salitre Formation in the Irecê basin (Misi and Kyle, 1994). These phosphate accumulations are stratigraphically controlled and present in both, cratonic non-deformed strata and marginal fold belts (see Dardenne et al., 1986, 1997; Da Rocha Araújo et al., 1992; Misi and Kyle, 1994; Misi et al., 2005b). According to Misi et al. (2005b), primary phosphate concentrations are related to global episodes of phosphatization associated with glacial events.

(iii) Mississippi valley-type CaF$_2$-Ba-Pb-Zn deposits are associated with the upper portion of the first carbonate megacycle of the Bambuí Group near Januária–Itacarambi, Montalvânia, Serra do Ramalho and Salitre basin.
All the data indicate migration of mineralising fluids from the Brasília belt towards the palaeogeographic high of the São Francisco Craton and the incorporation of radiogenic lead lixiviated from basement granite-gneisses.

(iv) Irish-type Pb-Zn deposits occur in the dolomitic facies of the Vazante Group at Morro Agudo mine (Misi et al., 1997, 1999a,b, 2005a,b; Dardenne and Freitas-Silva, 1999; Bettencourt et al., 2001; Monteiro, 2002). This mineralisation (Figure 3.9) was initiated by progressive connate fluid expulsion from the basinal sedimentary sequences under compression, related to uplift of Brasília belt, and subsequent channelisation along large normal fault systems.

(v) Hydrothermal Zn deposits, such as Vazante Mine, are associated with major faults (Figure 3.9). The mineralisation is related to the evolution of connate basinal and metamorphic fluids (Dardenne and Freitas-Silva, 1999; Misi et al., 2000, 2005a,b, 2007; Bettencourt et al., 2001; Monteiro, 2002).

3.6. Conclusions and Regional Analysis

A regional picture of the Neoproterozoic evolution of the São Francisco palaeocontinent depicted from a review of pertaining geological literature, coupled with relevant unpublished information and syntheses, leads to the following major conclusions:

(1) The correlation between the Vazante and Bambuí Groups is still a matter of debate. Sr isotopic data of carbonate fluorapatite and of associated micritic limestones (~0.7074 in both successions; see Misi et al., 2007) may support this correlation. Nevertheless, ages of ca. 650 and 950 Ma on detrital zircon grains from Bambuí and Vazante clastic units, respectively (Dardenne et al., 2003, 2007a,b; Rodrigues, 2007), and new unpublished Re-Os ages suggest that the Vazante Group may be older as previously thought.

One glacial event has been recorded at the base of the Vazante Group by Dardenne (2001, 2006), represented by diamicrites of the Santo Antônio do Bonito Formation. Azmy et al. (2006) have recognised the
A record of a second glaciation at the base of the Serra da Lapa Formation, at the top of the Vazante Group, based on the presence of diamictites and black shales with dropstones and on the presence of a negative $\delta^{13}C$ excursion in the overlying carbonates. The veracity of this second glaciation, however, is a matter of debate and requires further investigation.

(2) There is no unambiguous geochronological control on most carbonate formations. The Pb-Pb isochron age of $740^{+22}_{-22}$ Ma at the base of the Sete Lagoas Formation (Babinski et al., 2007) is the only successful age on a cap carbonate to date. It is, if correlations are correct, an indirect age on the Sturtian glaciation (Jequitai/Carrancas). However, the age of the carbonates and, consequently, of the Bambuí Group remains debatable considering that detrital zircon grains from the Serra de Santa Helena and Três Marias formations have yielded ages of ca. 650 Ma.

(3) In the light of similar $\delta^{13}C$ profiles, it is assumed that the Sete Lagoas Formation, the Salitre Formation (Una Group) and Olhos D’Água Formation (Sergipano belt) are probably correlative cap carbonates, deposited in the aftermath of the Sturtian snowball event. In saying so, a long-standing Sturtian interval is assumed representing either different glaciations or a protracted interval encompassing almost 80 Ma (Babinski et al., 2007).

All these carbonate sequences start with negative $\delta^{13}C$ (around −5‰) that increase gradually to a pronounced positive excursion (≥+8‰ to 10‰ or even higher in the case of the Sete Lagoas Formation) that has proven useful in regional palaeogeographic investigations (e.g. identification of palaeohighs in the Serra do Cabral and Jequitai regions; Santos et al., 2004). Carbonates of the Agua Preta Formation (Rio Pardo belt) show $\delta^{13}C$ values within the range of the Olhos D’Água Formation, despite limited C isotope data.

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**Figure 3.33** Macro and microscopic features in the Zn-Pb mineralisation of the Neoproterozoic basins of the São Francisco Craton. (a) Massive stratiform mineralisation at the N ore body, Morro Agudo deposit, Vazante Group, formed by millimetric layers of chert, dolomite (Dol) and fine-grained sphalerite (Sph), with patches of galena (Gn) and pyrite (Py) and with calcite veins (Cc). (b) Massive coarse-grained sphalerite replacing oolithic dolarenite, JKL orebody, Morro Agudo mine, Vazante Group. (c) Complex nodules within dolostone (Dol) with an inner core of pyrite (Py) surrounded by coarse-crystalline sphalerite (Sph) at Irecê basin, Una Group. Sphalerite also occurs as pore-filling crystals in dolostone. Microcline detrital grains (Mi) in blue colour. Cathodoluminescence photomicrograph by Misi and Kyle (1994). (d) Coarse crystalline sphalerite (Sph) within microquartz (Qz: length-slow quartz) in contact with carbonate fluorapatite (Ap) in dolostone in the Una Group, Irecê basin. Probably a sulphate nodule replaced by microquartz and by sphalerite post-dating the apatite formation. Transmitted light photomicrograph (Misi et al., 2005a).
The Jacoca Formation carbonates (older than the Olhos D’Água Formation), overlying 730 Ma-old Ribeiropolis diamictites in the Sergipano belt (B.B. Brito Neves, written communication), display distinct δ13C signatures, where virtually all values are negative in sections examined to date. The age of 615 Ma for metagreywackes of the Frei Paulo Formation on top of the Olhos D’Água Formation constrains the deposition of Jacoca (older) and Olhos D’Água (younger) Formations to the age range of 730—615 Ma.

4) The speculation that the Agua Preta (Rio Pardo belt; δ13C values as high as +8‰) and São Desidério Formation carbonates (Rio Preto belt) were deposited concomitantly with the Sete Lagoas awaits confirmation from further geological and isotopic investigations.

5) Primary phosphate concentrations (phosphorites) in the Neoproterozoic of the São Francisco Craton are related to global episodes of phosphatisation, probably associated with glaciogenic events. Most of these mineralisations are unequivocally above glacial horizons; some others are not, but occur immediately above negative δ13C excursions, suggesting they may nevertheless be related to glacial events.

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