São Luís Craton and Gurupi Belt (Brazil): possible links with the West African Craton and surrounding Pan-African belts

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Abstract: The São Luís Craton and the Palaeoproterozoic basement rocks of the Neoproterozoic Gurupi Belt in northern Brazil are part of an orogen having an early accretionary phase at 2240–2150 Ma and a late collisional phase at 2080 ± 20 Ma. Geological, geochronological and isotopic evidence, along with palaeogeographic reconstructions, strongly suggest that these Brazilian terrains were contiguous with the West African Craton in Palaeoproterozoic times, and that this landmass apparently survived subsequent continental break-up until its incorporation in Rodinia.

The Gurupi Belt is an orogen developed in the southern margin of the West African–São Luís Craton at *c*. 750–550 Ma, after the break up of Rodinia. Factors such as present-day and possible past geographical positions, the timing of a few well-characterized events, the structural polarity and internal structure of the belt, in addition to other indirect evidence, all favour correlation between the Gurupi Belt and other Brasiliano/Pan-African belts, especially the Médio Coreaú domain of the Borborema Province and the Trans-Saharan Belt of Africa, despite the lack of proven physical links between them. These Neoproterozoic belts are part of the branched system of orogens associated with amalgamation of the Amazonian, West Africa–São Luís, São Francisco and other cratons and minor continental blocks into the West Gondwana supercontinent.

The São Luís Craton and the Neoproterozoic Gurupi Belt in northern Brazil crop out through Phanerozoic sedimentary cover in response to Cretaceous tectonic uplift and doming that preceded the rifting stage and opening of the Atlantic Ocean and subsequent erosive removal of more than 6 km of Mesozoic and Palaeozoic sediments (Rezende & Pamplona 1970). The widespread remains of the sedimentary cover and the absence of palaeomagnetic information hinder a better understanding of the relationships, if any, of these two Precambrian terrains to the surrounding Precambrian units of the present-day South American continent, such as the Amazonian and São Francisco cratons, and the Neoproterozoic Borborema and Araguaia belts (Fig. 1).

Nevertheless, some attempts have been made to correlate the Precambrian units of northern and northeastern Brazil with those of northwestern Africa in pre-Pangaea palaeogeographic reconstructions. Hurley *et al.* (1967, 1968) and Torquato & Cordani (1981) presented Rb–Sr and K–Ar geochronological data and, following the pre-drift continental fit for the Atlantic Ocean (Bullard *et al.* 1965), observed geochronological similarities between Brazil and Africa in that litho-structural units having a Neoproterozoic imprint surrounded units having a Palaeoproterozoic signature. Lesquer *et al.* (1984) used regional scale structural and geophysical information to discuss the correlation.

These studies were important in establishing the broad limits between major geotectonic units, such as cratons, mobile belts and sedimentary basins. However, the internal geological framework of each of these units remained uncertain. Despite unavoidable and continuous debate, knowledge of the geology and tectonic evolution of the northwestern African terrains has experienced significant advances in the last 15 years (Abouchami *et al.* 1990; Feybesse & Milési 1994; Trompette 1997; Egal *et al.* 2002; Caby 2003, and many others). In Brazil, only recently the geological evolution of the São Luís Craton and Gurupi Belt became better understood, on the basis of regional mapping programmes (Pastana 1995; Costa 2000), more robust geochronological information and reinterpretation of tectonic settings (Klein & Moura 2001, 2003; Moura *et al.* 2003; Klein *et al.* 2005*a*, *b*).

Taking into account these new advances and problems, this paper intends to reassess possible

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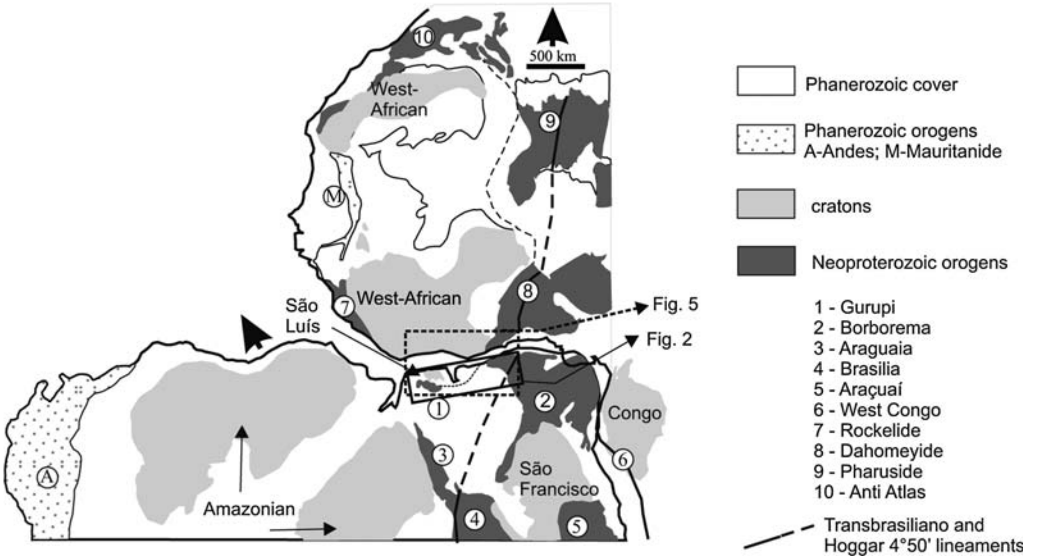

Fig. 1. Location of the São Luís craton and Gurupi Belt in relation to the main tectonic units of northern South America and northwestern Africa.

correlations between the São Luís Craton and the West African Craton (Fig. 1), and between the Gurupi Belt and the Brasiliano/Pan-African belts of northern Brazil and northwestern Africa (Figs 1 and 2). This will be done by comparing similarities (and differences) between rock associations and the tectonic settings in which these associations formed, and by the investigation of the timing of relatively well characterised geological events, such as magmatism, metamorphism, sedimentation and tectonism. Correlation problems will also be discussed, as well as their implications for the assembly and dispersal of supercontinents.

Geological overview

The São Luís Craton

The São Luís Craton is composed of a metavolcano-sedimentary sequence and a few generations of granitoids (Fig. 3). The former consists of schists of variable composition, metavolcanic and metapyroclastic rocks, quartzite, metachert, and meta-mafic–ultramafic rocks. The metamorphic conditions are predominantly greenschist facies, but locally reached lower amphibolite facies. This supracrustal sequence is considered to have formed in island-are setting (Klein *et al.* 2005*a*). Zircon geochronology of this sequence is still limited. A single sample of metapyroclastic rock yielded an age of 2240 ± 5 Ma (single zircon Pb evaporation), and Sm–Nd T_{DM} model ages vary from 2.21 to 2.48 Ga, with εNd(t) values of +0.8 to +3.5 (Klein & Moura 2001; Klein *et al.* 2005*a*). A similar metavolcano-sedimentary sequence that occurs in the basement sequence of the Gurupi Belt (Fig. 3) has zircon Pb–Pb ages of 2148–2160 Ma (Klein & Moura 2001) and it is possible that the supracrustal sequence of the São Luís Craton also continued to develop until this time.

Granitoids make up the major part of the cratonic area, forming batholiths and stocks. The Tromaí Suite is composed of equigranular and massive to weakly foliated quartz-diorite, tonalite, diorite, granodiorite and minor trondhjemite. These rocks are calc-alkaline, metaluminous and sodic, with low to moderate K_2O contents. Zircon crystallisation ages vary between 2168 Ma and 2147 Ma, and Sm–Nd T_{DM} model ages are 2.22 to 2.26 Ga, with εNd(t) values of +1.9 to +2.6 (Klein & Moura 2001; Klein *et al.* 2005*a*). The Areal granite (2149 ± 4 Ma) is weakly peraluminous, K_2O-enriched and has similar Sm–Nd patterns to the Tromaí Suite (Klein & Moura 2003; Klein *et al.* 2005*a*); it is interpreted as having formed from parental magmas similar to those that produced the Tromaí calc-alkaline granitoids, along with reworked products of the island arcs in which they formed (Klein *et al.* 2005*a*).

A third suite of granitoids (Tracuateua, Fig. 3) consists of strongly peraluminous, S-type two-mica granites, derived from the partial melting of crustal rocks (Lowell & Villas 1983). These granites have zircon Pb–Pb crystallization ages of 2086–2091 Ma and Sm–Nd model ages

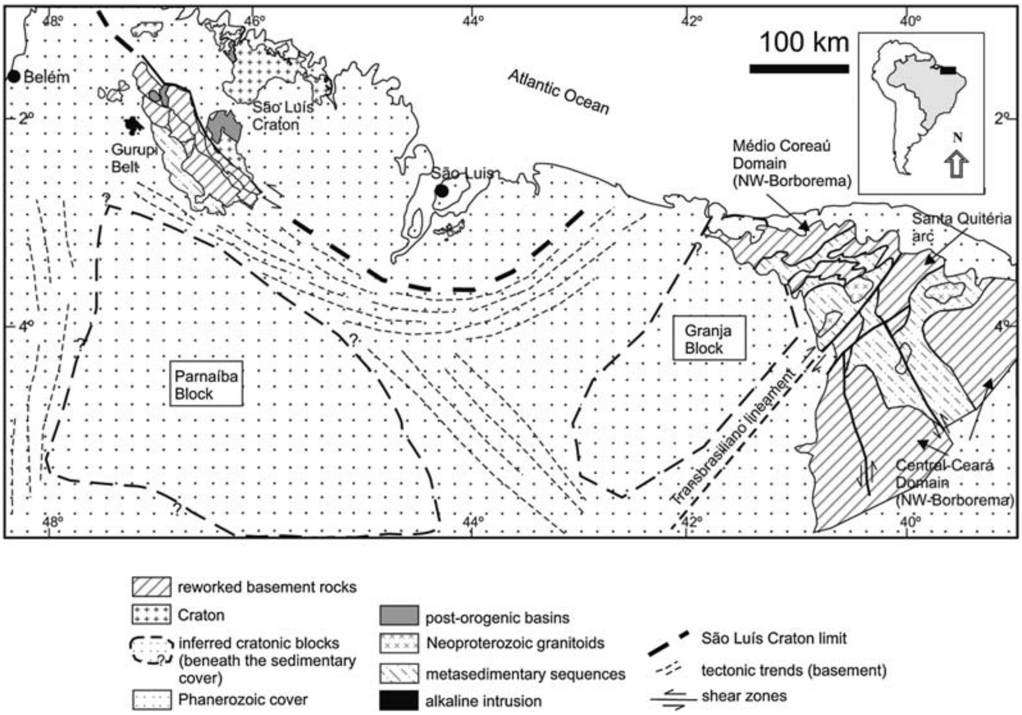

Fig. 2. Simplified map showing the location of the São Luís craton, Gurupi Belt and Médio Coreaú and Central-Ceará domains (NW Borborema Province) in relation to the interpreted tectonic trends of the basement rocks of the Phanerozoic sedimentary cover. These trends indicate the possible continuation, beneath the sedimentary cover, of the São Luís craton, Gurupi and Araguaia belts, and of the Transbrasiliano lineament. Also shown are the Parnaíba and Granja blocks inferred to underlie the Phanerozoic basins (adapted from Brito Neves *et al.* 1984; Nunes 1993; Fetter *et al.* 2003).

(T_{DM}) varying between 2.31 and 2.50 Ga, with εNd(t) values ranging from −1.3 to +1.1 (Moura *et al.* 2003).

The geodynamic evolution of the São Luís Craton has been discussed on the basis of rock assemblages and affinities, structural features, and limited geochemical information, combined with zircon geochronology and Nd isotope data (Klein *et al.* 2005*a*). At least three periods of rock generation have been recognized, occurring at about 2240 ± 5 Ma (supracrustal rocks), 2168–2147 Ma (calc-alkaline granitoids + supracrustal rocks), and 2086–2090 Ma (S-type granites), with almost all sequences showing positive εNd(t) values and therefore being derived from juvenile protoliths. The large association of juvenile calc-alkaline granitoids and volcano-sedimentary rocks and the lack of voluminous mafic rocks have been interpreted by Klein *et al.* (2005*a*) as indicating an intra-oceanic, are-related subduction setting for these sequences.

Only scarce relicts of a reworked Archaean crust have been indicated by slightly negative εNd(t) values and Sm–Nd model ages of the younger S-type granites (Moura *et al.* 2003). However, Sm–Nd T_{DM} model ages of 2.48 Ga and 2.42 Ga found in meta-dacite and dacite suggest that protoliths older than 2.24 Ga might have been involved at least in part of the cratonic evolution. As such, the model ages of about 2.4 Ga may record the age of the mafic protoliths (ocean crust?) that could have formed at that time and that were subsequently melted.

The time interval of 2240–2150 Ma records an accretionary phase of the Palaeoproterozoic orogen, whereas the collisional phase is represented by the S-type granitoids of 2086–2090 Ma, produced by melting of pre-existing crustal material. This phase is better represented in the basement of the Gurupi Belt (see next section).

The Gurupi Belt

The Gurupi Belt is a Neoproterozoic mobile belt located along the southern margin of the São Luís Craton (Fig. 3); it shows metamorphosed

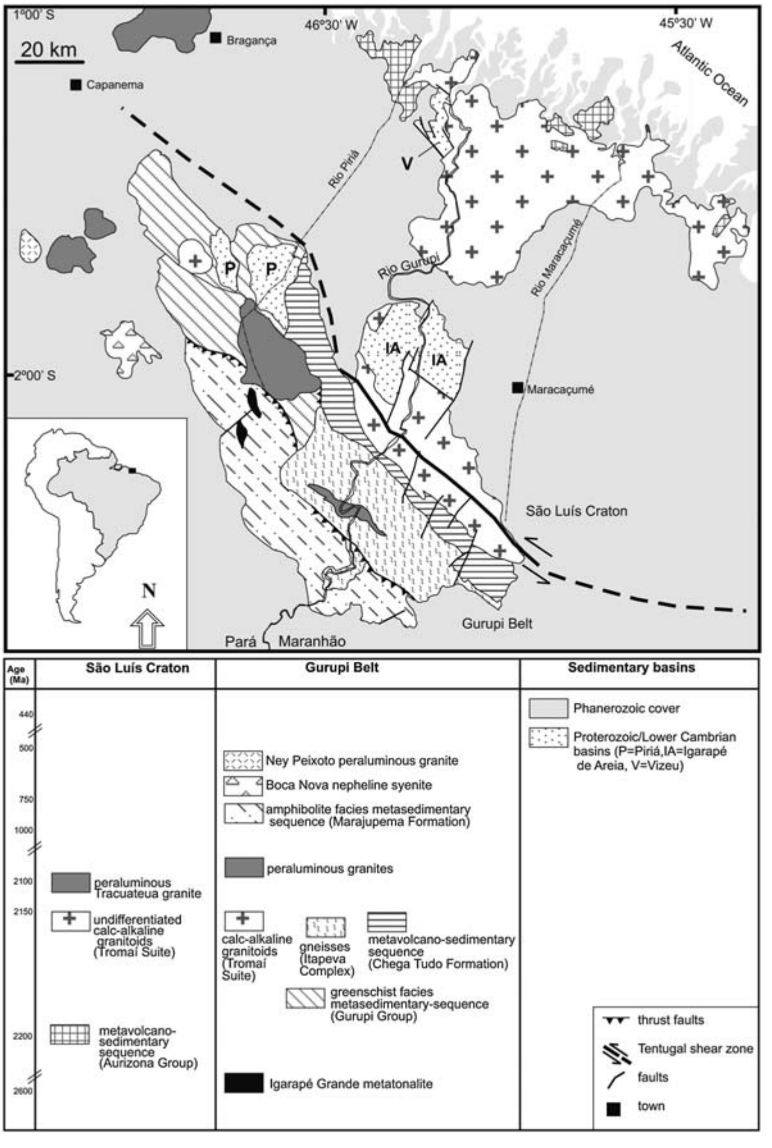

Fig. 3. Geological map of the São Luís craton and Gurupi Belt (adapted from Pastana 1995 and Klein *et al.* 2005*b*).

supracrustal and plutonic units originally formed in Archaean, Palaeoproterozoic and Neoproterozoic times (Klein & Moura 2001; Moura *et al.* 2003; Klein *et al.* 2005*b*). With the exception of some rounded granite stocks, most of the sequences of the Gurupi Belt form elongated bodies parallel to NW–SE cross-cutting structures. Structures dipping at low to moderate angles to SSW, with down-dip or oblique lineations, are mostly confined to the northwestern portion of the belt. These structures record tectonic transport from SW to NE, toward the São Luís Craton, and it is possible that they resulted from the convergence between this craton and an inferred landmass existing to the south. This block may correspond to the concealed Parnaíba block (Fig. 2), a cratonic nucleus having distinct age and structural trends in relation to the surrounding terrains, which has been proposed on the basis of geophysical evidence in addition to petrography and Rb–Sr and K–Ar geochronology of the basement rocks of the Phanerozoic basins (Brito Neves *et al.* 1984; Nunes 1993).

Steeply dipping strike-slip shear zones are concentrated in the central and southeastern portions of the belt, being associated with the Tentugal Shear Zone, which represents the litho-structural and geochronological (Rb–Sr, K–Ar) boundary between the São Luís Craton and the Gurupi Belt (Klein *et al.* 2005*b*). The Tentugal Shear Zone was active during the Neoproterozoic orogeny, but probably resulted from the reactivation of older structures related to the Palaeoproterozoic evolution of the São Luís Craton.

Most of the exposed rocks of the Gurupi Belt are Palaeoproterozoic units that show physical continuity and broadly exhibit the same age and Nd isotope patterns displayed by the Palaeoproterozoic rocks of the São Luís Craton. For instance, bodies correlated with the cratonic calc-alkaline granitoids (2168–2147 Ma) occur within the belt, where they show variable effects of deformation. In addition, a tonalite gneiss of 2167 $\pm$ 2.5 Ma (protolith crystallization age), has T_{DM} model ages between 2.22 and 2.31 Ga, and εNd(t) values ranging from +1.4 to +1.6 (Klein *et al.* 2005*b*). These gneisses could represent the same calc-alkaline granitoids as those of the São Luís Craton that underwent more severe metamorphic and deformational conditions, but additional studies are needed to solve this issue.

A metavolcano-sedimentary succession (Chega Tudo Formation) contains felsic volcanic rocks of 2148–2160 Ma with a juvenile Nd isotope signature, and was probably formed within are systems (Klein & Moura 2001; Klein *et al.* 2005*b*). Several plutons of peraluminous, biotite- and muscovite-bearing granite intruded the Palaeoproterozoic supracrustal and gneissic units between 2100 and 2060 Ma, and at least one of these plutons is clearly syntectonic. Inherited zircon crystals and Nd isotopes indicate that the peraluminous rocks were formed by variable degrees of reworking of Palaeoproterozoic and minor Archaean crust (Moura *et al.* 2003; Klein *et al.* 2005*b*). A sub-greenschist to greenschist facies metasedimentary sequence of unknown age (the Gurupi Group) is tentatively considered to be older than 2159 $\pm$ 13 Ma, based on supposed intrusion relationships (Costa 2000).

All these Palaeoproterozoic units appear to have been originally related to the evolution of the present-day São Luís Craton sequences, broadly representing an accretionary (São Luís) and a collisional phase (Gurupi) of a Palaeoproterozoic orogen. However, the units located in the Gurupi Belt show variable but widespread evidence of resetting of Rb–Sr and K–Ar isotopic systems by Neoproterozoic events (Klein *et al.* 2005*b* and references therein). Furthermore, they show a distinct structural pattern in relation to the present-day São Luís Craton. These Palaeoproterozoic rocks, along with subordinate lenses of an Archaean metatonalite (Igarapé Grande, Fig. 3) represent the continental basement on which the Neoproterozoic Gurupi orogen developed, and now survive as the external portion of the orogen.

Apart from the generalized resetting of isotopic systems, only two magmatic events after the Palaeoproterozoic orogeny have so far been characterized in the Gurupi Belt, both occurring in Neoproterozoic times. The first event was the intrusion of the Boca Nova nepheline syenite pluton 732 $\pm$ 7 Ma ago (Klein *et al.* 2005*b*), probably recording the rifting of the pre-existing crust amalgamated in the Palaeoproterozoic era. This intrusion (Figs 2 and 3) was subsequently deformed and metamorphosed under amphibolite-facies conditions (Lowell & Villas 1983); gneissic banding strikes NW–SE and dips at low angles to the WSW. The second event was the intrusion of a peraluminous, muscovite-bearing granite at 549 $\pm$ 4 Ma (Ney Peixoto Granite, Moura *et al.* 2003). This granite was only moderately affected by the widespread NW–SE strike-slip shearing, and is interpreted as a late- to post-tectonic intrusion.

Further evidence of younger (Neoproterozoic?) activity is the presence of detrital zircon crystals as young as 1100 Ma in the amphibolite-facies metasedimentary Marajupema Formation (Fig. 3), which has Sm–Nd T_{DM} model age of 1.41 Ga (Klein *et al.* 2005*b*). These authors suggested that the sedimentation of detritus from Archaean, Palaeoproterozoic, and Mesoproterozoic/Neoproterozoic sources could have occurred in the rift in which the nepheline syenite intruded, or on a continental margin. A possibility (not unique) is that this basin was subsequently closed at the end of the Ediacaran period (580–550 Ma), after a period of inferred subduction, arc construction, and collision.

Small sedimentary basins formed over sequences of the São Luís Craton (Vizeu and Igarapé de Areia basins) and Gurupi Belt (Piriá Basin) and show similar lithological, metamorphic and structural aspects (Fig. 3). The basins comprise variable proportions of arkose, sandstone, pelite and conglomerate that record continental semi-arid conditions to shallow lake or marine waters (Pastana 1995). A large proportion of detrital zircons (>80%) found in an arkose from one of these basins shows Pb–Pb ages between 700 and 500 Ma (Pinheiro *et al.* 2003), indicating that this basin at least formed late in the Neoproterozoic era or even in the Early Cambrian epoch and leading most authors to relate the basins to the post-orogenic development of the Gurupi Belt (Pinheiro *et al.* 2003; Klein *et al.* 2005*b*).

The Médio Coreaú Domain of the Borborema Province

The Borborema Province in northeastern Brazil (location in Fig. 1) is a large branching system of orogenic belts that shows a long-lived and polycyclic geological history. Its basement rocks record orogenic and anorogenic processes from 3500 Ma to 940 Ma. The Neoproterozoic (Brasiliano) orogeny took place diachronously throughout the province, but involved basically the same sequence of events, which can be summarized as follows (Brito Neves *et al.* 2000; Fetter *et al.* 2000, 2003): (1) 850–700 Ma, continental rift phase, with small volcano-sedimentary and plutonic complexes; (2) 650–620 Ma, extensive subduction-related calc-alkaline magmatism, and pre- to syn-collisional metavolcano-sedimentary sequences; (3) 620–570 Ma, collisional plutonism; (4) 580–510 Ma, post-tectonic to anorogenic plutonism, uplift and extrusion tectonics.

The northwestern portion of the Borborema Province is represented by the Médio Coreaú domain (Figs 1 and 2). This domain, which occurs to the west of the Transbrasiliano lineament, comprises sedimentary and volcano-sedimentary sequences deformed and metamorphosed in the Neoproterozoic era, during the Brasiliano cycle of orogenies. These sequences were deposited over a middle- to high-grade basement composed of gneiss and migmatite, along with subordinate enderbite, charnockite, and kinzigite, with magmatic ages varying from 2356 Ma to 2176 Ma and Sm–Nd model ages clustering between 2.42 and 2.48 Ga, indicating a juvenile character for most of these rocks (Brito Neves *et al.* 2000; Fetter *et al.* 2000). Neoproterozoic orogenic magmatism (777 ± 11 to 591 ± 8 Ma) falls in approximately the age intervals reported in the previous paragraph for the Borborema province as a whole and the tectonic setting of the orogenic granites is interpreted as related to back-arc and fore-arc basins (Fetter *et al.* 2003).

The West African Craton

The West African Craton is represented by two Precambrian shields, Reguibat to the north and Man to the south, separated by Neoproterozoic to Palaeozoic cover (Fig. 4a). The Man shield, which is of more interest for the present work, is subdivided into a western domain, predominantly of Archaean age (Kénema–Man domain), and a central–eastern domain, composed of Palaeoproterozoic rocks (Baoulé–Mossi domain). These domains are separated by the strike-slip Sassandra Shear Zone and by an Archaean-Palaeoproterozoic transitional domain (Feybesse & Milési 1994; Caby *et al.* 2000; Egal *et al.* 2002).

The Kénema–Man domain is composed of granulite gneiss, migmatite and charnockite of Archaean age, along with subordinate Palaeoproterozoic granitoid, volcanic and sedimentary rocks. Crust formation events, at least in part juvenile, have been identified at 3542 ± 13 Ma, 3300–3200 Ma (pre-Leonian and Leonian orogeny), whereas an extensive period of granite magmatism and granulite-facies metamorphism, reworking of older crust and apparent absence of deposition of supracrustal rocks occurred between 2910 Ma and 2800 Ma (Liberian orogeny). Perturbation of isotopic systems occurred between 2250 Ma and 2020 Ma (Thiéblemont *et al.* 2004 and references therein), as result of the Palaeoproterozoic Eburnian orogeny that is widespread in the Baoulé–Mossi domain.

The Baoulé–Mossi domain (Fig. 4a) consists of several NNE–SSW-orientated belts of metavolcanic, metasedimentary and metavolcano-sedimentary rocks, with voluminous batholiths of granitoid rocks having variable ages and sources, as well as variable petrographic, geochemical and structural characteristics, and subordinate mafic–ultramafic rocks. These rocks formed mostly during, and were affected by, the widespread Palaeoproterozoic Eburnian orogeny. The metavolcanic rocks comprise tholeiitic to subordinate komatiitic basalts, along with calc-alkaline rhyolite and rhyodacite, whereas the sedimentary and metavolcano-sedimentary belts are composed of clastic sedimentary rocks intercalated with metavolcaniclastic and felsic to intermediate metavolcanic and pyroclastic rocks (Sylvester & Attoh 1992; Feybesse & Milési 1994; Caby *et al.* 2000; Hein *et al.* 2004). Greenschist-facies metamorphism is largely predominant, but lower amphibolite conditions occur locally (Vidal *et al.* 1996). Gabbro–diorite–pyroxenite bodies intrude the supracrustal sequences and appear to predate the granitic magmatism (Hein *et al.* 2004).

Two main phases of granitic magmatism have been characterised in the Baoulé–Mossi domain (Doumbia *et al.* 1998). The dominant type of granitoids comprises sodic–calcic and metaluminous calc-alkaline tonalites and granodiorites. These rocks derived from juvenile sources and intruded supracrustal rocks chiefly at about 2155 ± 15 Ma. The other type is composed of more potassic and peraluminous granitoids, including crust-derived two-mica granites that intruded the supracrustal sequences at 2100 ± 10 Ma.

There is a general consensus about the above description regarding the Baoulé–Mossi domain. However, at least three main subjects of debate still remain. Firstly, the stratigraphic position of

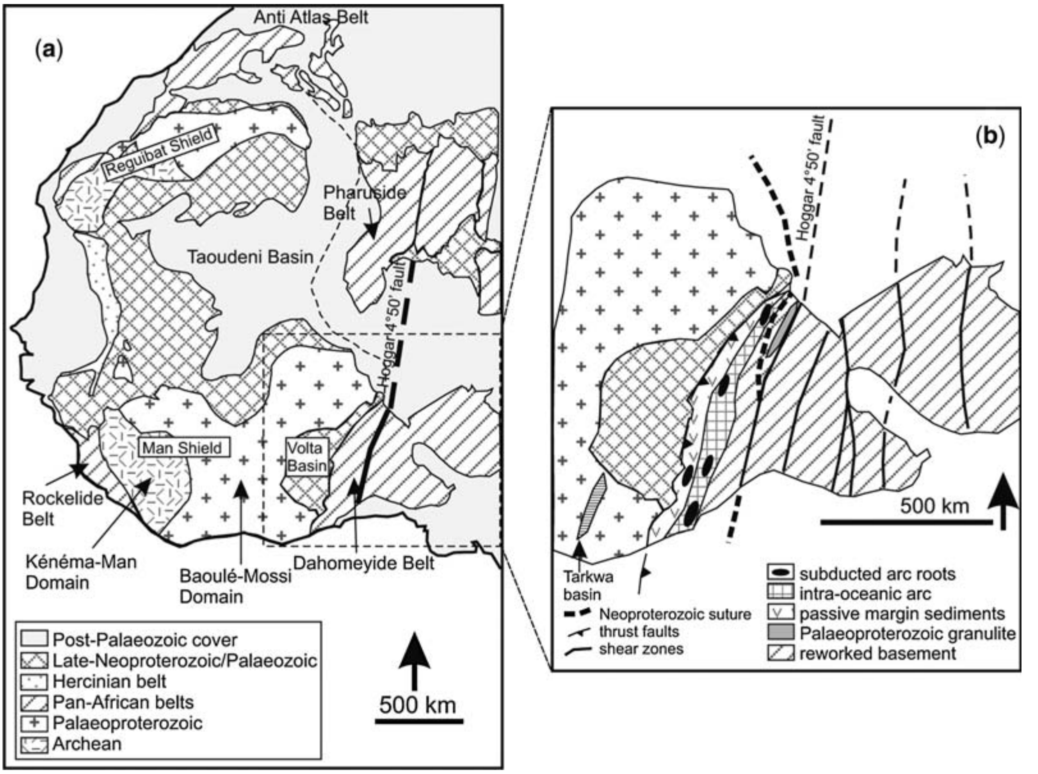

Fig. 4. (**a**) Simplified tectonic map of northwestern Africa (adapted from Feybesse & Milési 1994); (**b**) Detail of the Dahomeyide Belt in the Ghana–Nigeria region (adapted from Caby 1998), showing the tectonic elements of the Neoproterozoic orogen. The location of the Tarkwa basin is also shown.

the metasedimentary units, which are placed either below (Feybesse & Milési 1994) or above (Hirdes *et al.* 1996; Pouclet *et al.* 1996) the metavolcanic sequences.

Secondly, the tectonic setting in which the granitoid and (meta-)volcanic rocks formed: continental rift in a pre-Birimian basement older than 2200 Ma that would have evolved to an oceanic basin (Vidal & Alric 1994); oceanic within-plate setting (Abouchami *et al.* 1990); immature island arcs constructed over an oceanic crust (Sylvester & Attoh 1992); marginal marine setting adjacent to a volcanic centre or island arc (Hein *et al.* 2004). Because of the large extension of the Palaeoproterozoic domain, it is reasonable to expect some variability in terms of tectonic settings. Thus, since different authors have often worked in distinct portions of the domain, it is likely that most of the interpretations are to some extent feasible.

The third subject of discussion concerns the geotectonic evolution of the Palaeoproterozoic domain, with one group of researchers accepting modern plate tectonics (e.g., Abouchami *et al.* 1990; Ledru *et al.* 1994; Egal *et al.* 2002), and other researchers invoking vertical, plume-related, 'Archaean type' dynamics associated with strike-slip tectonics (e.g., Vidal *et al.* 1996; Caby *et al.* 2000). Irrespective of the models, the two schools describe basically the same sequence of events that can be grouped as follows, at least for the post-2200 Ma period: (1) early oceanic stage producing tholeiitic basic rocks; (2) juvenile, metaluminous, calc-alkaline, sodic plutonism and volcanism; (3) clastic and volcaniclastic sedimentation derived from the precedent stages; (4) another generation of calc-alkaline magmatism and extensive production of crust derived leucogranites; (5) metamorphism and deformation.

Compilation of more than one hundred zircon U–Pb and Pb–Pb age determinations and Nd information shows that this evolution is nearly continuous between at least 2200 Ma and 2060 Ma, with peaks of activity around 2155 Ma and 2095 Ma, and that the Palaeoproterozoic Baoulé-Mossi domain is essentially juvenile. Notwithstanding, the existence of a not yet understood pre-2.2 Ga (early-Birimian?) crustal growth episode is becoming evident, as suggested by geochronological (Lahondère *et al.* 2002; Gasquet *et al.* 2003) and stratigraphic–structural evidence (Hein *et al.* 2004). As such, recycling of pre-2.2 Ga crust might also have occurred in the genesis of the Birimian rocks of the Baoulé–Mossi domain.

The Tarkwa sedimentary sequence

The Tarkwa sedimentary mega-sequence shows very low metamorphism and well-preserved sedimentary structures. From the bottom to the top it consists of: (1) polymict conglomerate with pebbles of phyllite, granite, felsic and mafic lavas, pyroclastic rocks, quartz, and chert; (2) sandstone and auriferous monomict conglomerate with quartz pebbles; (3) phyllite and sandstone. These rocks were deposited in elongated graben formed over the Birimian belts and record continental to mature fluviatile and deltaic environmental conditions, and they are locally associated with pyroclastic flows and rhyolites, and with dolerite and gabbro sills (Davis *et al.* 1994; Feybesse & Milési 1994; Ledru *et al.* 1994). Most of the detrital zircons of this sequence have ages in the range of 2185–2155 Ma, with subordinate ages of 2245 Ma, 2132 Ma and 2124 Ma (Davis *et al.* 1994; Bossière *et al.* 1996), and the maximum age of sedimentation would be bracketed by intrusion relationships as between 2080 and 1960 Ma (Ledru *et al.* 1994; Bossière *et al.* 1996). There is a certain agreement in that both the Birimian and Tarkwa successions have been deformed together during the final stages of the Palaeoproterozoic Eburnian orogeny (Davis *et al.* 1994; Ledru *et al.* 1994).

The Pan-African belts

The West African Craton is nearly entirely surrounded (Fig. 4a) by orogenic belts of Neoproterozoic (Dahomeyide, Pharuside, Anti-Atlas, Bassaride, Rockelide) and Hercynian (Mauritanide) age. In general, the Neoproterozoic belts have been interpreted as the result of a protracted and diachronous (1000–500 Ma) succession of subduction and collisional events that involved disparate tectonic settings, including passive and active continental margins, platform covers, fragments of older continents, ocean basins and crust and magmatic arcs (Trompette 1997; Caby 2003).

The Pan-African Rockelide–Bassaride and Dahomeyide belts are located in the southern margins of the West African Craton (Fig. 4a). These belts were initiated as the margins of the West African Craton rifted, forming passive margins with clastic and carbonate sedimentation, followed by opening of an oceanic basin. Both belts are characterized by extensive reworking of older sequences and high-grade metamorphism (Villeneuve & Cornée 1994; Trompette 1997; Caby 1998).

The Rockelide Belt seems to represent an intracontinental orogeny, or passive margin, with diachronous evolution from north to south. The reworked basement is composed of high-grade gneisses, deep crustal granitoids, and Mesoproterozoic to Neoproterozoic (>700 Ma) cratonic covers, whereas the rift sequences comprise volcanic and volcaniclastic rocks. Intracontinental rifting occurred before about 550 Ma, and collision with the oriental portion of the Guyana Shield occurred at 550 Ma, provoking the thrusting of the Rockelide successions onto the West African Craton (Villeneuve & Cornée 1994).

The Dahomeyide Belt is the southern portion of the larger Trans-Saharan Belt that includes the Pharuside Belt to the north (Fig. 4a). It comprises a pre-orogenic rift phase that evolved to an active margin, with subduction and calc-alkaline magmatism occurring between 700 Ma and 600 Ma, and final collision against the eastern margin of the West African Craton at 610–600 Ma, producing granulite-facies metamorphism. Post-collisional plutonism occurred until 500 Ma (Villeneuve & Cornée 1994; Trompette 1997). In more detail, Caby (1998) recognized several tectonic elements that make up the Dahomeyide Belt in the southeastern portion of the West African Craton. These include (Fig 4b): (1) basement rocks composed of Eburnian (2.2 to 2.1 Ga) and allochthonous polycyclic gneisses that represent subducted fragments of Palaeoproterozoic continental crust; (2) passive margin sediments (Volta Basin); (3) subducted passive palaeo-margin sediments (mainly quartzites) of the West African Craton; (4) an intra-oceanic island arc (volcano-volcaniclastic sequence) and subducted arc roots (mafic–ultramafic to tonalitic rocks); (5) Pan-African suture zone; (6) thrust faults; (7) major ductile shear zones (e.g., the Hoggar 4°50' lineament). Caby (1998) describes voluminous calc-alkaline magmatism only in the northern continuation of the Dahomeyide Belt, the Pharusian Belt.

Sedimentary covers of the West African Craton

The West African Craton is widely covered by sedimentary sequences whose infilling occurred mostly in Neoproterozoic and Cambrian times and extended, in some places, until the Carboniferous period, being more or less affected by tectonism during the Pan-African cycle of orogenies. These sequences include those of the Tindouf Basin to the north, the Volta Basin to the SE, and several sub-basins in the central portion of the craton separating the Reguibat and Man shields (Fig. 4a) that may be broadly grouped together in the Taoudeni Basin. The thickness of these basins attained 8 km in some places, with glacial, marine and continental sediments recording a complex tectono-sedimentary evolution (Villeneuve & Cornée 1994).

Attempting to correlate northern Brazil and Western Africa

In this section we reassess possible links between the Precambrian terrains of northern Brazil and Western Africa, based on similarities (and differences) of rock association and geochemistry, zircon geochronology and isotope data. These elements can help in the discussion of the assembly and break up events through the identification of possible tectonic environments in which the associated rocks formed and, in consequence, of processes such as rifting, formation of oceanic basins and island arcs, and collision.

The Palaeoproterozoic domains

Palaeomagnetic data are lacking for the São Luís Craton. There is recent information for the Amazonian and West African cratons (Nomade *et al.* 2003; Tohver *et al.* 2006), but the two studies show different reconstructions, and one (Nomade *et al.* 2003) does not take the São Luís Craton into account. An alternative starting point comes from the study of the evolution of the central Atlantic transform faults based on sea floor topographic and gravity evidence (Sandwell & Smith 1995), and of the sedimentary and tectonic evolution of Phanerozoic sedimentary basins on both sides of the equatorial Atlantic (Matos 2000; Pletsch *et al.* 2001). Palaeogeographic reconstructions based on these studies consistently put the São Luís Craton and the Gurupi Belt opposite the present-day coastline of Ivory Coast (Fig. 5).

The three periods of magmatic activity defined in the São Luís Craton and in the Palaeoproterozoic portion of the Gurupi Belt (2240 ± 5 Ma, 2160 ± 10 Ma, 2080 ± 20 Ma) are also found in the West African Craton, having the same lithological and tectonic characteristics (Fig. 6). Furthermore, the geotectonic evolution of São Luís and

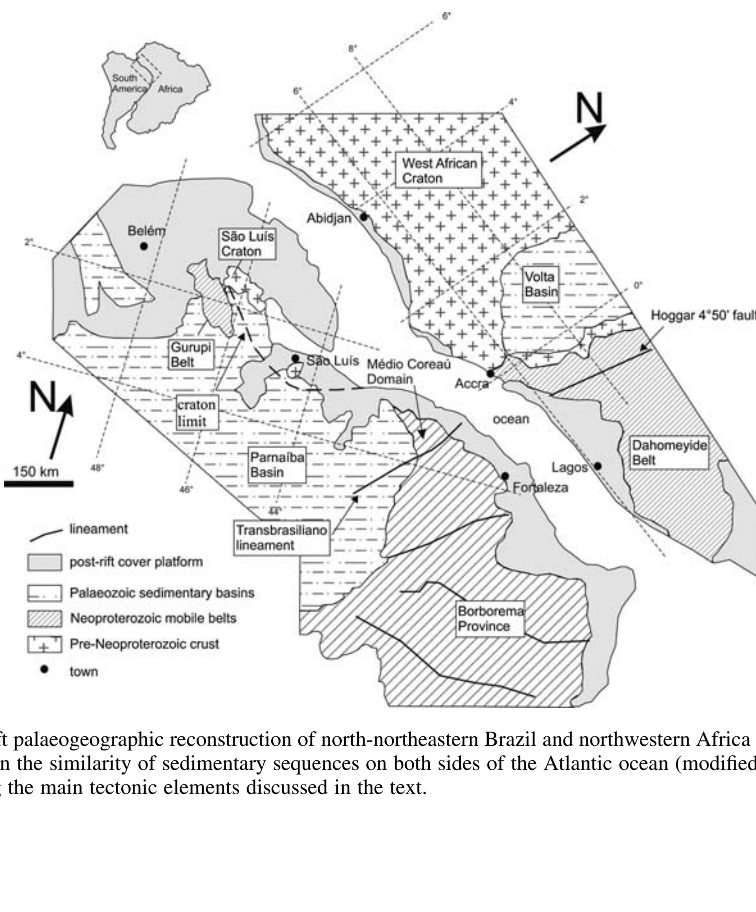

Fig. 5. Pre-drift palaeogeographic reconstruction of north-northeastern Brazil and northwestern Africa for the Aptian period, based on the similarity of sedimentary sequences on both sides of the Atlantic ocean (modified from Matos 2000), showing the main tectonic elements discussed in the text.

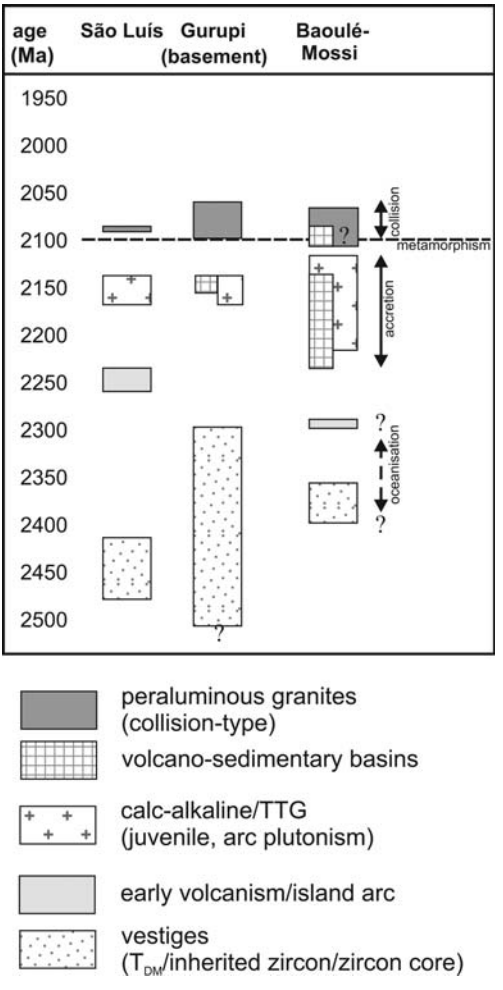

Fig. 6. Tectonic correlation chart for the Paleoproterozoic domains. See text for references and discussion.

West Africa in the Palaeoproterozoic era is interpreted in a similar way. This supports the correlation between the Brazilian and African terrains and the view that the São Luís Craton represents a fragment of the southern portion of the West African Craton (Baoulé–Mossi domain).

Several differences must be highlighted, because the West African Craton records a much more complex and nearly continuous evolution between *c.* 2220 Ma and 2060 Ma. For instance, in the São Luís Craton/Gurupi Belt there is no known geological record in the 2220–2170 Ma and 2140–2100 Ma time intervals, both of which are characterized by plutonic and volcanic activity in the West African Craton. Also, the granitoids of the West African Craton show a wider diversity in terms of petrology, geochemistry, age and association. The initial oceanic stage in the West African Craton, marked by extensive mafic–ultramafic volcanism, including abundant komatiite, has not yet been recognized in the São Luís Craton/Gurupi Belt, where basic magmatism is subordinate. The Tarkwa sedimentary sequence is a very important component of the West African Craton stratigraphy but no such sequence is known in the São Luís Craton so far. Despite the fact that the age and tectonic meaning of the Viseu Basin (Fig. 3) is far from being understood, this basin discordantly overlies the Palaeoproterozoic rocks of the São Luís Craton.

These differences may be real or due to distinct levels of knowledge in Brazilian and African counterparts, but could also be related to the fact that the Brazilian area is comparatively smaller than the African one.

The Neoproterozoic domains

The correlation of the Gurupi Belt with any other Brasiliano/Pan-African belt is not straightforward (see a comparison between the main events described for the Neoproterozoic marginal belts of the São Luís–West African craton in Fig. 7) because of fundamental problems that come from the Phanerozoic tectonic (uplift), sedimentary and erosive history, the widespread Phanerozoic cover and the relatively limited exposure of the Gurupi Belt with consequent difficulties in understanding the limits and the internal architecture of the belt.

The problem of the concealed limits of the belt was at least partially resolved by the geophysical evidence and the study of the basement rocks of the Phanerozoic basins (Lesquer *et al.* 1984; Nunes 1993). These data indicate that the Gurupi Belt extends 60–80 km to the south below the basins, where it is limited by the inferred Parnaíba block (Fig. 2).

To the east, the geophysical information (Lesquer *et al.* 1984; Nunes 1993) highlights the curvilinear shape of the Gurupi Belt, probably outlining the margin of the São Luís Craton (Fig. 2). Furthermore, the basement rocks a few tens of kilometres east of the town of São Luís have been affected by Neoproterozoic events. As such, the Gurupi Belt is probably connected with the Médio Coreaú domain (Fig. 2), despite differences in the age, metamorphism and lithological content of the basement rocks (older, higher-grade rocks occur in the Médio Coreaú domain).

To the west, the continuation of the Gurupi Belt is still obscure. An interesting solution has been proposed by Villeneuve & Cornée (1994), in which the Rockelide–Araguaia–Gurupi belts formed a triple junction by the convergence of the Amazonian and São Luís–West African cratons,

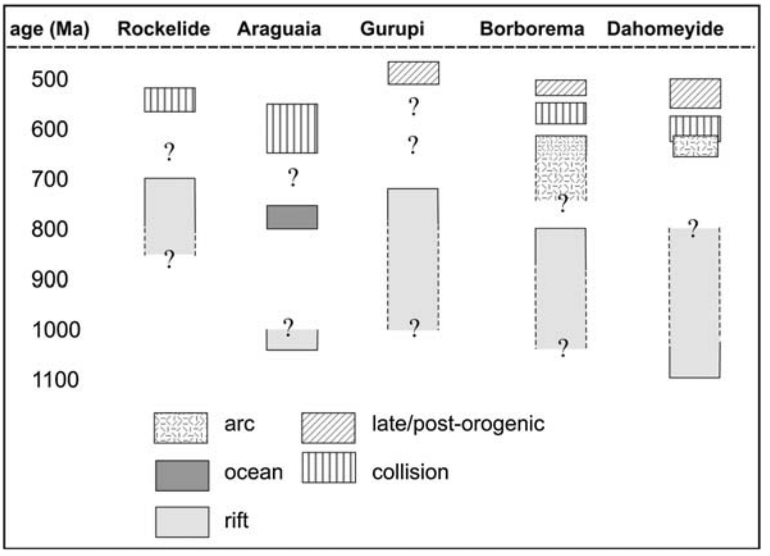

Fig. 7. Tectonic correlation chart for Brasiliano/Pan-African belts. See text for references and discussion.

and a third block located between these two cratons, possibly the Parnaíba block. In fact, convergence between the São Luís–West African craton and the Parnaíba block is a possible hypothesis to explain origin of the Gurupi Belt (Klein *et al.* 2005*b*). This is, however, a working hypothesis, since not even those works dealing with the basement of the Phanerozoic cover have provided any explanation for the continuation of the Gurupi Belt to the west (Fig. 2).

The closest and apparently least debatable link between north-northeastern Brazil and western Africa appears to be between the Médio Coreaú domain and the Dahomeyide Belt along the Hoggar 4°50' –Transbrasiliano lineaments (Figs 1 and 2). Brito Neves *et al.* (2002), however, stated that many features of the Dahomeyide Belt described by Caby (1998) are not found in the Médio Coreaú domain. This includes the tectono-sedimentary record and the absence on the Brazilian side of collisional linearity and of the Neoproterozoic suture. Brito Neves *et al.* (2002) argued that this could result from the drag effect related to the Hoggar 4°50' –Transbrasiliano lineaments, and to the shape of the Neoproterozoic ocean that developed off the eastern margin of the West African Craton. We understand that this comment is also valid for the whole southeastern margin of the West African–São Luís craton, i.e., we also include the area that encompasses the Gurupi Belt.

Positive gravity anomalies have been interpreted as reflecting the concealed suture zone between the Neoproterozoic belts (Gurupi, Borborema) and the West African–São Luís craton (Lesquer *et al.* 1984). This is a possibility since mafic–ultramafic and granulite-facies rocks have been documented in the eastern margin of the West African Craton. However, the lithological record of this suture has been found neither in the Gurupi Belt nor in the Médio Coreaú domain, and metamorphic conditions attained in the Gurupi Belt are only of amphibolite facies. Moreover, the Tentugal strike-slip shear zone (Fig. 3) represents only the geochronological (Rb–Sr, K–Ar) boundary between the São Luís Craton and the Gurupi Belt and not a suture, since the same rock sequences are found on both sides (Klein *et al.* 2005*b*). An alternative location for the suture could be south of the Tentugal shear zone, approximately near the gneissose nepheline syenite (Fig. 3), since deformed alkaline rocks are good indicators of the proximity of suture zones (see Burke *et al.* 2003). Therefore, the gravity contrasts described by Lesquer *et al.* (1984) may still indicate the presence of denser rocks (granulite facies?), but these would underlie the exposed Gurupi Belt.

A major problem with any model for the Neoproterozoic tectonic evolution (Klein *et al.* 2005*b*) and the discussion of the internal architecture of the Gurupi Belt is its incomplete lithological record. Most of the rocks that crop out in the belt are Palaeoproterozoic rocks of the São Luís Craton that have been reworked during the Neoproterozoic orogenic events, i.e., they represent the cratonic margin and the external domain of the orogen.

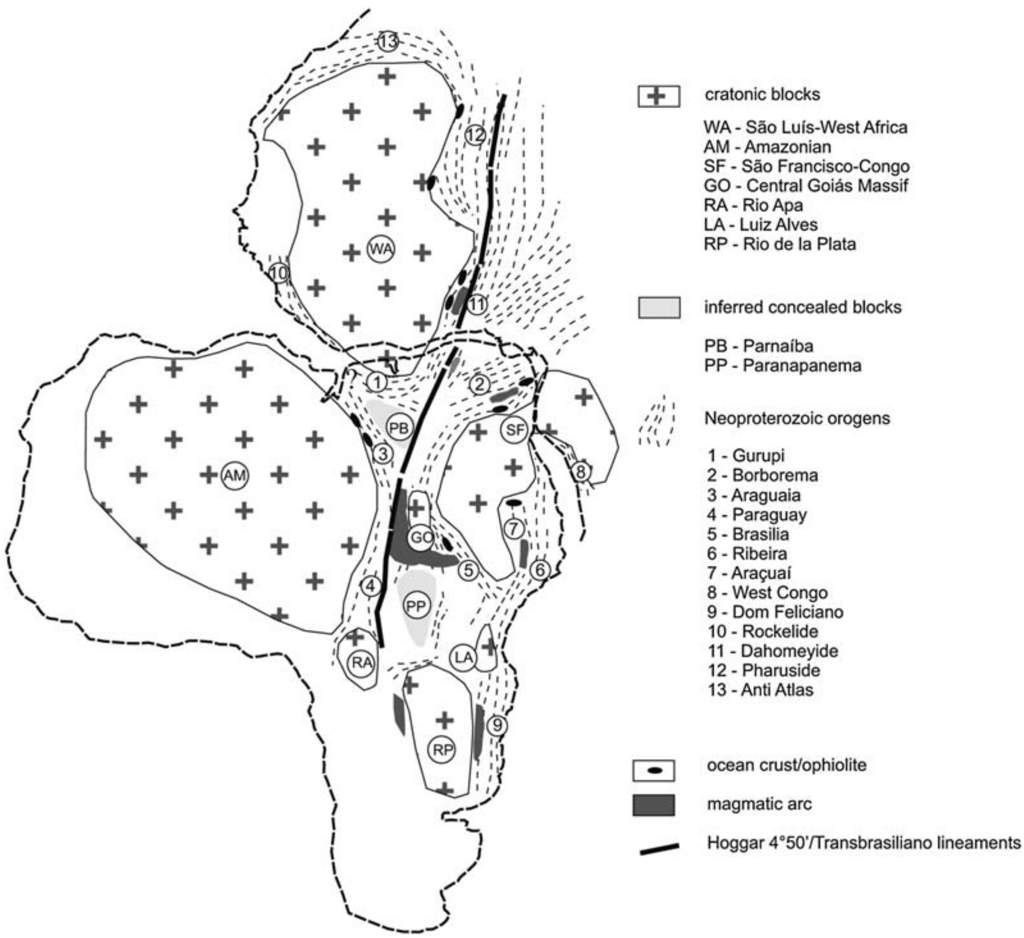

Fig. 8. Cartoon (not to scale) showing the tectonic elements of South America and northwestern Africa that participated of the West Gondwana assembly (references in the text). Although based primarily on geological comparisons and correlations, this reconstruction is essentially compatible with that of Tohver *et al.* (2006), which is based on palaeomagnetic control (albeit not for the São Luís craton). Both reconstructions differ from that of Nomade *et al.* (2003), which is also based on palaeomagnetic data, but without taking into account the position of the São Luís craton.

Otherwise, only the pre-orogenic (or syn-rift) and late/post-orogenic magmatic stages have been recognised at *c.* 732 and 550 Ma, respectively (Fig. 7) and we can only speculate as to what happened between these events. For instance, the existence of more felsic to intermediate rocks (arc-related?) within, or in the proximity of, the Gurupi Belt has been inferred (Pinheiro *et al.* 2003; Klein *et al.* 2005*b*) from the presence of abundant 550 Ma old detrital zircons in an immature (proximal) arkose of the Igarapé de Areia Formation and from a Sm–Nd T_{DM} model age of 1.4 Ga found in the Neoproterozoic amphibolite-facies metasedimentary sequence of the Marajupema Formation (Klein *et al.* 2005*b*).

The tectonic setting of this amphibolite-facies sequence is not clear. However the abundance of quartz-rich rocks (quartzite, feldspar-rich quartzite and coarse-grained quartz–mica schist) and the absence of associated igneous rocks suggest a passive margin environment, without carbonate deposition. This passive margin probably developed from the continental rift that contained the nepheline syenite intrusion and then evolved to an active margin.

The age of metamorphism is still unconstrained. There are a number of K–Ar and Rb–Sr mineral ages in the 466–618 Ma interval (see Klein *et al.* 2005*b* for a review and primary references). It is uncertain if these ages represent metamorphism, cooling/uplift, or late extrusion tectonics. Since the Neoproterozoic peraluminous granite of 549 ± 4 Ma has not been affected by the low-angle deformation imparted to the supracrustal rocks and the nepheline syenite in the north-western portion of the belt, metamorphism and tangential deformation are interpreted as older than the emplacement of this granite (Klein *et al.* 2005*b*).

Implications for the assembly and break-up of supercontinents

São Luís Craton and Gurupi Belt in West Gondwana

There is reasonable geological, geochronological and isotopic evidence that in Neoproterozoic times a large branching system of ocean basins extended from western Africa to central Brazil, passing through north-western Brazil. The evidence is provided by (Fig. 8): (1) mafic–ultramafic rocks and calc-alkaline granitoids in the Pharusian–Dahomeyide belts (Caby 1998); (2) continental magmatic arc rocks of 650–620 Ma in the northwestern (Médio Coreaú domain) and eastern portions of the Borborema Belt (Brito Neves *et al.* 2000; Fetter *et al.* 2003); (3) mafic–ultramafic rocks of 757 ± 49 Ma that represent obducted oceanic crust in the Araguaia Belt (Paixão *et al.* 2002); (4) juvenile magmatic arc rocks of 890–600 Ma in the Brasília Belt (Pimentel *et al.* 2005).

The closure of these ocean basins via subduction and the attendant convergence of several continental blocks in Neoproterozoic/Early Cambrian times resulted in orogenic belts, including the Gurupi Belt, that amalgamated these blocks forming the western part of the Gondwana supercontinent. The ocean closure was not linear but it probably formed a branching system and the main suture is approximately outlined by the Hoggar 4°50'–Transbrasiliano lineaments (Fig. 8). The overall evolution of these orogenic belts involved the formation of continental rifts, marginal and oceanic basins, island arcs, subduction and collision, with the events occurring diachronously (Fig. 7) in different geographic areas of West Gondwana. In the study area, the continental blocks that participated in West Gondwana assembly include, among others, the Amazonian and West African–São Luís cratons, which are the best-preserved cratonic blocks, being affected only in their margins, and the concealed Parnaíba block (Fig. 8).

São Luís Craton in Rodinia

Geological, tectonic and geochronological aspects of the São Luís and West African cratons (and of the Palaeoproterozoic portion of the Gurupi Belt), along with previous palaeogeographic and geochronological reconstructions, strongly support the interpretation that these terrains were contiguous in Palaeoproterozoic times.

Most of this Palaeoproterozoic landmass remained relatively stable until its incorporation into Rodinia (Condie 2002; Rogers & Santosh 2002). This is almost certainly true with respect to the São Luís Craton, where there is no known geological activity in the time interval between about 2000 and 750 Ma. Rifting of the (present-day) southern margin of the São Luís Craton occurred slightly before the intrusion of the nepheline syenite pluton (732 ± 7 Ma). This event marks an early stage of the Gurupi orogen and the beginning of Rodinia break-up in this region. At this time, and even before, ocean realms had already been developed in other parts of future West Gondwana.

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