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African, southern Indian and South American cratons were not part of the Rodinia supercontinent: evidence from field relationships and geochronology

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Abstract

We discuss the question whether the late Mesoproterozoic and early Neoproterozoic rocks of eastern, central and southern Africa, Madagascar, southern India, Sri Lanka and South America have played any role in the formation and dispersal of the supercontinent Rodinia, believed to have existed between about 1000 and 750 Ma ago. First, there is little evidence for the production of significant volumes of ~1.4–1.0 Ga (Kibaran or Grenvillian age) continental crust in the Mozambique belt (MB) of East Africa, except, perhaps, in parts of northern Mozambique. This is also valid for most terranes related to West Gondwana, which are made up of basement rocks older than Mesoproterozoic, reworked in the Brasiliano/Pan-African orogenic cycle. This crust cannot be conclusively related to either magmatic accretion processes on the active margin of Rodinia or continental collision leading to amalgamation of the supercontinent. So far, no 1.4–1.0 Ga rocks have been identified in Madagascar. Secondly, there is no conclusive evidence for a ~1.0 Ga high-grade metamorphic event in the MB, although such metamorphism has been recorded in the presumed continuation of the MB in East Antarctica. In South America, even the Sunsas mobile belt, which is correlated with the Grenville belt of North America, does not include high-grade metamorphic rocks. All terranes with Mesoproterozoic ages seem to have evolved within extensional, aulacogen-type structures, and their compressional deformation, where observed, is normally much younger and is related to amalgamation of Gondwana. This is also valid for the Trans-Saharan and West Congo belts of West Africa.

Third, there is also no evidence for post-1000 Ma sedimentary sequences that were deposited on the passive margin(s) of Rodinia. In contrast, the MB of East Africa and Madagascar is characterized by extensive structural reworking and metamorphic overprinting of Archaean rocks, particularly in Tanzania and Madagascar, and these rocks either constitute marginal parts of cratonic domains or represent crustal blocks (terranes or microcontinents?) of unknown derivation. This is also the case for most terranes included in the Borborema/Trans-Saharan belt of northeastern Brazil and west-central Africa, as well as those of the Central Goiás Massif in central Brazil and the Mantiqueira province of eastern and southeastern Brazil.

Furthermore, there is evidence for extensive granitoid magmatism in the period ~840 to < 600 Ma whose predominant calc-alkaline chemistry suggests subduction-related active margin processes during the assembly of the supercontinent Gondwana. The location of the main Neoproterozoic magmatic arcs suggests that a large oceanic domain separated the core of Rodinia, namely Laurentia plus Amazonia, Baltica and West Africa, from several continental masses and fragments now in the southern hemisphere, such as the São Francisco/Congo, Kalahari and Rio de La Plata cratons, as well as the Borborema/Trans-Saharan,

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Central Goiás Massif and Paraná blocks. Moreover, many extensional tectonic events detected in the southern hemisphere continental masses, but also many radiometric ages of granitois that are already associated with the process of amalgamation of Gondwana, are comprised within the 800–1000 age interval. This seems incompatible with current views on the time of disintegration of Rodinia, assumed to have occurred at around 750 Ma.
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Keywords: Rodinia; Gondwana; Grenville-age belt; Kibaran-age belt; Irumide-age belt; African cratons; South American cratons; Southern India; Sri Lanka

1. Introduction

There has been much debate in recent years on the existence, configuration and evolution of the supercontinent Rodinia (McMenamin and McMenamin, 1990; for a historical review and literature, see Hoffman, 1999). Formation of Rodinia around a core of Precambrian North America (Laurentia) during the Grenvillian event (~1300–1000 Ma ago) is supposed to have occurred by accretion of all existing continental fragments of that time, including those that now make up the cratonic components of the Gondwana continents. Continental aggregation, in this view, was marked by collision of these fragments with each other and with Laurentia, giving rise to several broad mobile (orogenic) belts, the deep, eroded parts of which are now exposed in high-grade terranes such as the Grenville belt of North America (Rivers et al., 1989; Hoffman, 1991) and the Eastern Ghats belt of India (Ramakrishnan et al., 1998) (Fig. 1). Mobile belts of central and southern Africa that formed during the above time interval (e.g., Kibaran, Irumide, Namaqua, Natal, Lurio) are collectively known as Kibaran (Cahen et al., 1984; Tack et al., 1994) (Fig. 2). However, the Kibaran orogeny in east-central Africa took place at ~1400–1350 Ma, clearly in pre-Grenvillian times, whereas the typical Grenville-age event is reflected by the Irumide belt of Zambia (Fig. 2) that formed at ~1150–950 Ma (Rivers and Johnson, in press).

In South America, mobile belts that formed in Mesoproterozoic times were described from within the Amazonian craton where they constitute the Rondonian–San Ignacio Province (1500–1300 Ma) and the Sunsas belt (1200–1000 Ma, Tassinari et al., 2000) (Fig. 3). Within the São Francisco Craton, there occurs the Espinhaço Folded System (da Costa and Inda, 1982), where sediments as old as 1700 Ma where deformed by the Espinhaço orogeny at about 1300 Ma (Cordani et al., 1992). All these belts have featured strongly in the Rodinia debate since they were considered to have played a major role in the assembly of the supercontinent (Unrug, 1992, 1996, 1997; Kampunzu, 2001) (Fig. 4). This also includes the Namaqua–Natal belt of South Africa (Jacobs et al., 1997; Dalziel et al., 2000; but see alternative view of Powell et al., 2001a,b), Madagascar (Hoffman, 1991; see Fig. 1) and parts of East Africa (Unrug, 1992) where apparent Kibaran ages have been reported in the past (see summary in Cahen et al., 1984 and the review by Rivers and Johnson, in press).

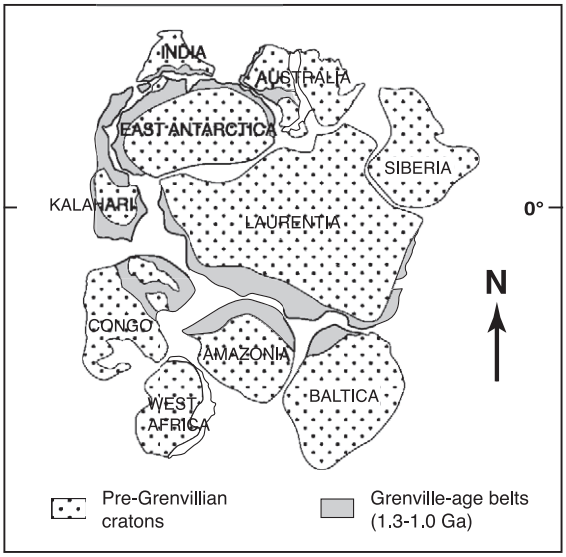


Fig. 1. Map of Rodinia after Hoffman (1991). Note Grenville–Kibaran mobile belts, interpreted to have resulted from collision of Archaean to Palaeoproterozoic continental blocks with Laurentia.

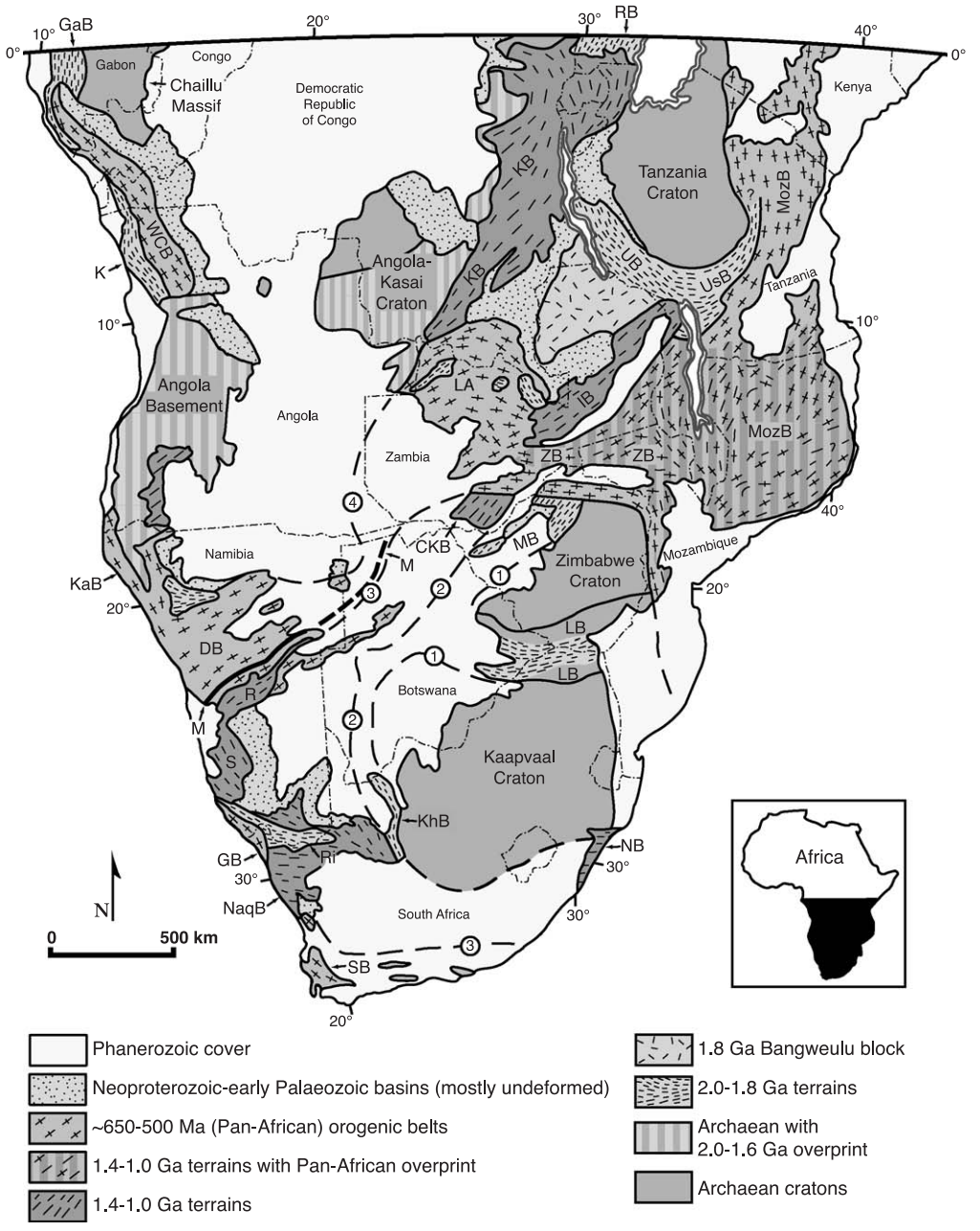


Fig. 2. Precambrian tectonic framework of southern Africa, slightly modified from Hanson (2003). Structural trends are shown schematically. Dashed lines with circled numbers in Kalahari region indicate inferred positions of major subsurface tectonic boundaries. (1) Western edge of Archaean cratons; (2) boundary between 2.05–1.8 Ga orogenic belts and 1.35–1.0 Ga belts; (3) boundary between 1.35–1.0 Ga belts and Pan-African belts; (4) western edge of Pan-African belts. Abbreviations for orogenic belts and other geological features: CKB—Choma–Kalomo block; DB—Damara belt; GaB—Gabon belt; GB—Gariep belt; IB—Irumide belt; K—Kimezian belt; KaB—Kaoko belt; KB—Kibaran belt; KhB—Kheis belt; LA—Lufilian arc; LB—Limpopo belt; M—Matchless amphibolite; MB—Magondi belt; MozB—Mozambique belt; NaqB—Namaqua belt; NB—Natal belt; R—Rehoboth Inlier; RB—Ruwenzori belt; Ri—Richtersveld terrane; S—Sinclair sequence; SB—Saldania belt; UB—Ubendian belt; UsB—Usagaran belt; WCB—West Congo belt; ZB—Zambezi belt.

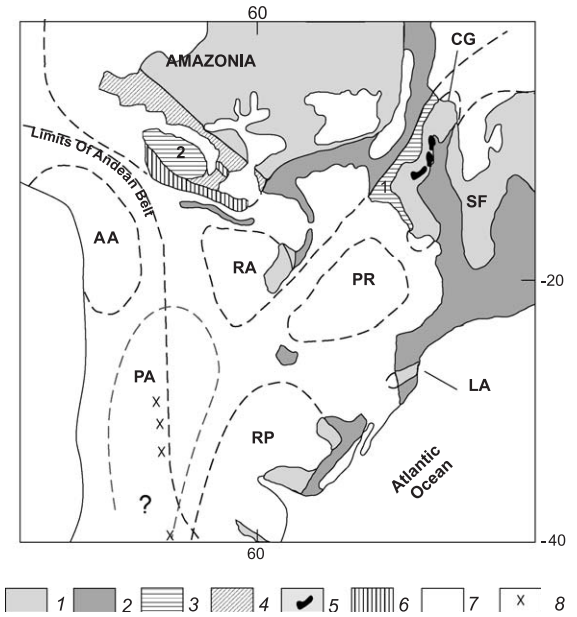


Fig. 3. Precambrian tectonic framework of Central South America. (1) Cratonic masses and fragments, dashed lines indicate inferred positions of major cratonic fragments below Phanerozoic cover: SF—São Francisco; RA—Rio Apa; PR—Paraná; RP—Rio de la Plata; PA—Pampia; AA—Arequipa–Antofalla; LA—Luiz Alves; CG—Central Goiás. (2) Late Neoproterozoic mobile belts. (3) Juvenile granitoid rocks of intra-oceanic magmatic arcs: 1—Goiás; 2—São Gabriel. (4) Rondonian–San Ignácio mobile belt. (5) Mafic–ultramafic layered complexes within Central Goiás Massif: from north to south, Cana Brava, Niquelândia and Barro Alto. (6) Sunsás orogenic belt, including Aguapeí formation. (7) Phanerozoic cover. (8) Outcrops of Neoproterozoic granitoids of the Sierra de Córdoba magmatic arc.

Of particular relevance to this debate are the high-grade rocks of the Mozambique belt (MB, Holmes, 1951) or East African Orogen (EAO, Stern, 1994), extending from Ethiopia through Kenya, Tanzania and Malawi into northern Mozambique (Shackleton, 1996; Pinna et al., 1993) and also including Madagascar (Windley et al., 1994), southernmost India (Braun and Kriegsmann, 2003; Jacobs et al., 1998, 1999; Fitzsimons, 2000) (Fig. 5). New isotopic, petrologic and structural data from southeast Africa and Madagascar place constraints on the evolution of these rocks and their significance for the formation and dispersal of Rodinia and subsequent amalgamation of the supercontinent Gondwana.

The history of accretion and crustal amalgamation during the Grenvillian–Kibaran–Sunsas event was followed by a period of supercontinent disintegration and dispersal, considered by most authors to have begun some 750–800 Ma ago (Hoffman, 1991, 1999; Pisarevsky et al., 2003). Powell et al. (1993) suggested on palaeomagnetic grounds that Rodinia did not break apart before 725 Ma, a view that is incompatible with the evidence from East Africa and Madagascar (Kröner, 2002a), Sri Lanka (Kröner et al., in press) and the Brazilian shield (Cordani et al., 2000). By about 550 Ma, most of the dispersed terranes had reassembled, probably in a different configuration, in the supercontinent Gondwana (Meert and Van der Voo, 1997). This occurred after closure of several oceans of different sizes. For instance, the Mozambique ocean between East and West Gondwana, conventionally interpreted to have been consumed and ending in collision of these two large continental masses (Burke and Dewey, 1972) is now seen as the site of Neoproterozoic arc and microcontinent accretion (Kröner et al., 2001; Kröner, 2002b; Powell and Pisarevsky, 2001). Likewise, the South American and African continental blocks of Amazonia, West Africa, Rio de la Plata, São Francisco/Congo and Borborema assembled into West Gondwana in several stages of ocean closure and collision between 900 and ~550 Ma (see, e.g., Brito-Neves and Cordani, 1991; Campos-Neto, 2000; Alkmim et al., 2001, Cordani et al., 2003).

2. How do we recognize continental accretion, collision and dispersal?

Accretion may occur in intra-oceanic domains by the formation and collision of island arcs, oceanic plateaus and microcontinents. One of the best documented examples is the evolution of the Arabian–Nubian Shield (ANS) in the Neoproterozoic (Johnson, 2001, and references therein), where a number of island arc complexes containing calc-alkaline magmatic rocks with primitive isotopic signatures have been identified which are separated from each other by linear ophiolite belts. ‘‘Soft’’ collision of these arc terranes did not lead to extensive crustal thickening during the accretion process so that metamorphic grades remained relatively low, and no extensive uplift occurred. This resulted in the exposure of upper to middle crustal levels and preservation of typical plate tectonic signatures such as arc assemblages,

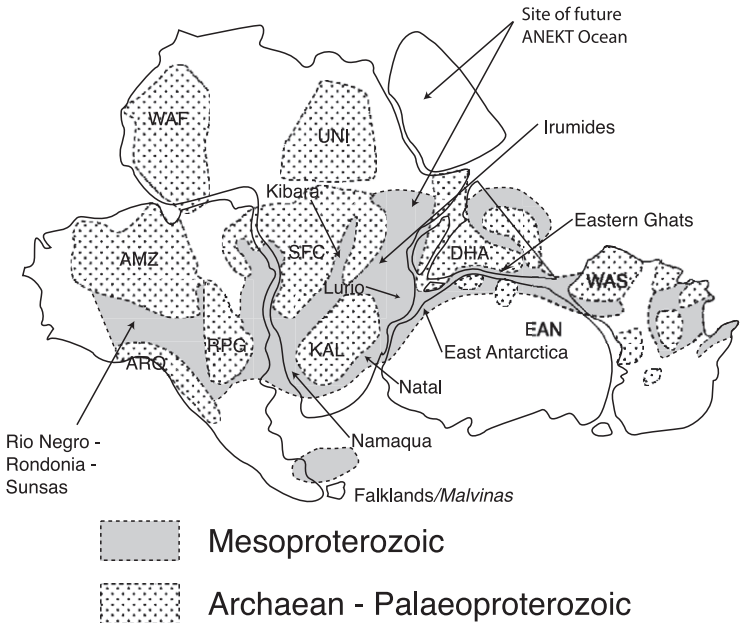
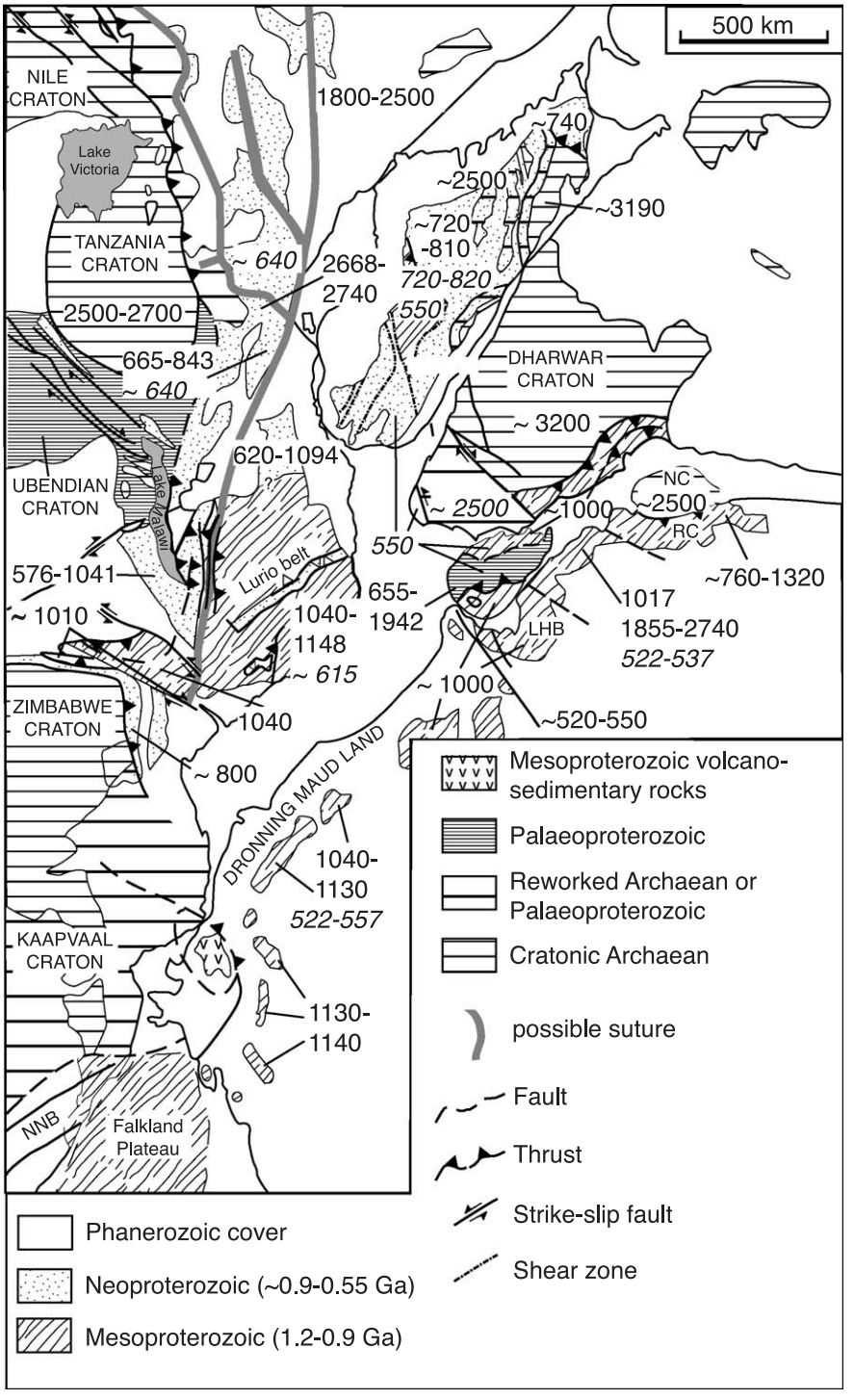


Fig. 4. Gondwana supercontinent with Kibaran (~1400–1000 Ma) mobile belts in the interpretation of Unrug (1992). Cratonic domains in stipples: AMZ = Amazon; ARQ = Arequipa; RPG = Rio de la Plata–Goiás; WAF = West Africa; UNI = Uweinat–Nile; CSF = Congo/São Francisco; KAL = Kalahari; DHA = Dharwar; WAS = West Australia; EAN = East Antarctica. Site of Neoproterozoic ANEKT Ocean (Unrug, 1996) is indicated.

accretionary complexes, ophiolite suites, thrust-and-fold belts and so on (for summaries, see Kröner et al., 1987; Johnson, 2001). The entire arc terrane was ultimately accreted to the African continent and became part of the cratonic domain (Gass, 1984; Kröner, 1985). Similar processes have operated in the vast terrane of the Central Asian Mobile Belt (Altaids) in late Neoproterozoic to Palaeozoic times (e.g., Sengör et al., 1993; Sengör and Natal'in, 1998; Badarch et al., 2002) and can be observed in the SW Pacific today and can be observed in the SW Pacific today (Hamilton, 1979, 1988; Petterson et al., 1999).

Accretion in now deeply eroded crustal domains is much more difficult to recognize because most of the signatures summarized above are eroded away, leaving a "sea" of complexly and ductilely deformed amphibolite- to granulite-facies migmatitic gneisses. Distinct crustal domains (terranes?) can sometimes be demarcated on the basis of structural discontinuities (shear zones) and isotopic characteristics such as precise zircon or Nd mean crustal residence ages as in the Central Goiás Massif of central Brazil (Pimentel et al., 2000) or in the MB of Madagascar (e.g., Collins and Windley, 2002), Tanzania (e.g., Muhongo et al., 2001) and Sri Lanka (Milisenda et al., 1988). Kröner (2002b) has recently interpreted the Mozambique belt of East Africa, Madagascar and Sri Lanka as a giant Neoproterozoic accretionary orogen.

Signatures resulting from continental collision need not be elaborated on here since they have been described extensively in the literature (Burke and Dewey, 1972; Windley, 1995). Collision frequently leads to crustal thickening and subsequent orogenic collapse, finally resulting in uplift and exposure of high-grade gneiss assemblages such as those exposed in the MB or in the southern Tocantins orogenic system of central Brazil (Campos-Neto and Caby, 1999). High-grade gneisses resulting from continental collision typically display clockwise PT paths (Harley, 1989). Many of the features characteristic of collision belts such as deformed accretionary wedges, ophiolite-decorated sutures, blueschists, nappe stacking and so on are no longer preserved in the deeply eroded orogens, making reliable reconstructions of the collisional and accretionary process almost impossible (e.g., Key et al., 1989).


500 km
NILE CRATON
Lake Victoria
TANZANIA CRATON
2500-2700
665-843
~ 640
UBENDIAN CRATON
Lake Malawi
576-1041
~ 1010
ZIMBABWE CRATON
1040
~ 800
KAAPVAAL CRATON
NNB
Falkland Plateau
1800-2500
~ 640
2668-2740
620-1094
Lurio belt
1040-1148
~ 615
DRONNING MAUD LAND
1040-1130
522-557
1130-1140
740
~2500
~720-810
720-820
550
~3190
DHARWAR CRATON
~ 3200
~ 2500
1000
550
655-1942
~ 1000
LHB
NC
~2500
RC
~760-1320
1017
1855-2740
522-537
~520-550
Mesoproterozoic volcano-sedimentary rocks
Palaeoproterozoic
Reworked Archaean or Palaeoproterozoic
Cratonic Archaean
possible suture
Fault
Thrust
Strike-slip fault
Shear zone
Phanerozoic cover
Neoproterozoic (~0.9-0.55 Ga)
Mesoproterozoic (1.2-0.9 Ga)

Dispersal of supercontinents is exemplified by the break-up history of Pangaea (Nance et al., 1988; Cox, 1992). Dispersal is preceded by extensional processes, resulting in crustal thinning and the emplacement of bimodal dyke swarms before rifts form that eventually evolve into small ocean basins, such as the Red Sea (Bailey, 1992; Girdler, 1983; Pallister, 1987). This is generally accompanied by the emplacement of alkali granitoids and associated felsic dykes typically found in evolving continental rift zones (Williams, 1982) or during ocean opening (Nielsen, 1987). Once a small ocean basin has formed, sedimentary debris is shed into the ocean, and a passive continental margin evolves with deposition of clastic and carbonate sequences onto the older crust that was previously deformed. During the process of dispersal, continental fragments may also get involved in the subduction of oceanic crust and the formation of magmatic arcs and/or marginal basins along the edges of such terranes. This can be observed today in Indonesia where fragments of the Precambrian Australian continent are engulfed by subduction-related magmatic rocks (e.g., Pigram and Panggabean, 1984). All these rock assemblages may be modified by deformation and metamorphism once the displaced terranes collide again with a continent and participate in the next cycle of supercontinent aggregation.

Crustal domains situated in the interior of an accreting supercontinent, such as North America during the formation of Rodinia (Hoffman, 1999), leave virtually no evidence of the accretionary process and are usually also devoid of isotopic ages characteristic of the period during which supercontinent formation occurred. Thus, the lack of isotopic ages in the bracket 1300–1000 Ma is no evidence per se for noninvolvement of a given crustal domain in the formation of a supercontinent.

The following account will summarize new data from the Mozambique belt of East Africa, central Madagascar and Sri Lanka as well as the Rondonian–San Ignacio and Sunsas Provinces in South America in order to determine what role these terranes played, if any, in the Rodinia–Gondwana evolutionary history between ~1300 and ~550 Ma ago. Some comments will also be given for other pertinent regions of Brazil and Africa, where events in the 1300–1000 Ma age interval have been described and were attributed to the making up of the late Mesoproterozoic supercontinent. For purposes of illustration, simplified maps are used, where the various continental fragments are placed in their most likely position prior to the break-up of Gondwana (Figs. 5 and 6).

3. The Mozambique belt of East Africa and Madagascar

Unrug (1992) assumed the existence of a system of Mesoproterotozoic (Kibaran–Grenvillian) mobile belts surrounding the Kalahari and Sao Francisco/Congo cratons (Fig. 3) that he thought were responsible for the formation of what he named the supercontinent Pangaea Y, now generally known as Rodinia. He considered the basement of the Neoproterozoic Mozambique belt to be Kibaran in age and to have formed by collisional processes during the assembly of Pangaea Y (=Rodinia). In contrast, he interpreted the proper Kibaran as well as the Irumide belts of east-central Africa (Fig. 2) as having evolved from intracratonic rifts accommodated by transverse shear zones (Unrug, 1992, p. 219). Kampunzu (2000) followed similar reasoning but considered the Kibaran and Irumide belts to have played a major role in the formation of Rodinia. He also considered the earliest Neoproterozoic extensional magmatism in east-central Africa at around 900 Ma to have resulted from collapse of the crust thickened during supercontinent amalgamation and linked this to Rodinia break-up.

The following summaries of the various segments of the MB in East Africa and Madagascar critically examine the above hypothesis and discuss whether there are any real Kibaran-age rocks and what role they played, if any, in the evolution of the belt. Furthermore, the question is addressed whether the presence of Kibaran-age rocks in East Africa can generally be

Fig. 5. Early Mesozoic Gondwana reconstruction (mostly after Lawyer et al., 1998) showing main geologic units in east and southeast Africa, Madagascar, southern India, Sri Lanka and East Antarctica (after Shackleton, 1996, with modifications of Madagascar from Collins et al., 2000; Collins and Windley, 2002; Kröner et al., 2000a). LHB=Lützow–Holm Bay; NNB=Namaqua–Natal belt; NC=Napier Complex; RC=Rayner Complex. Normal numbers are ages for magmatic events; numbers in italics are ages for metamorphic events.

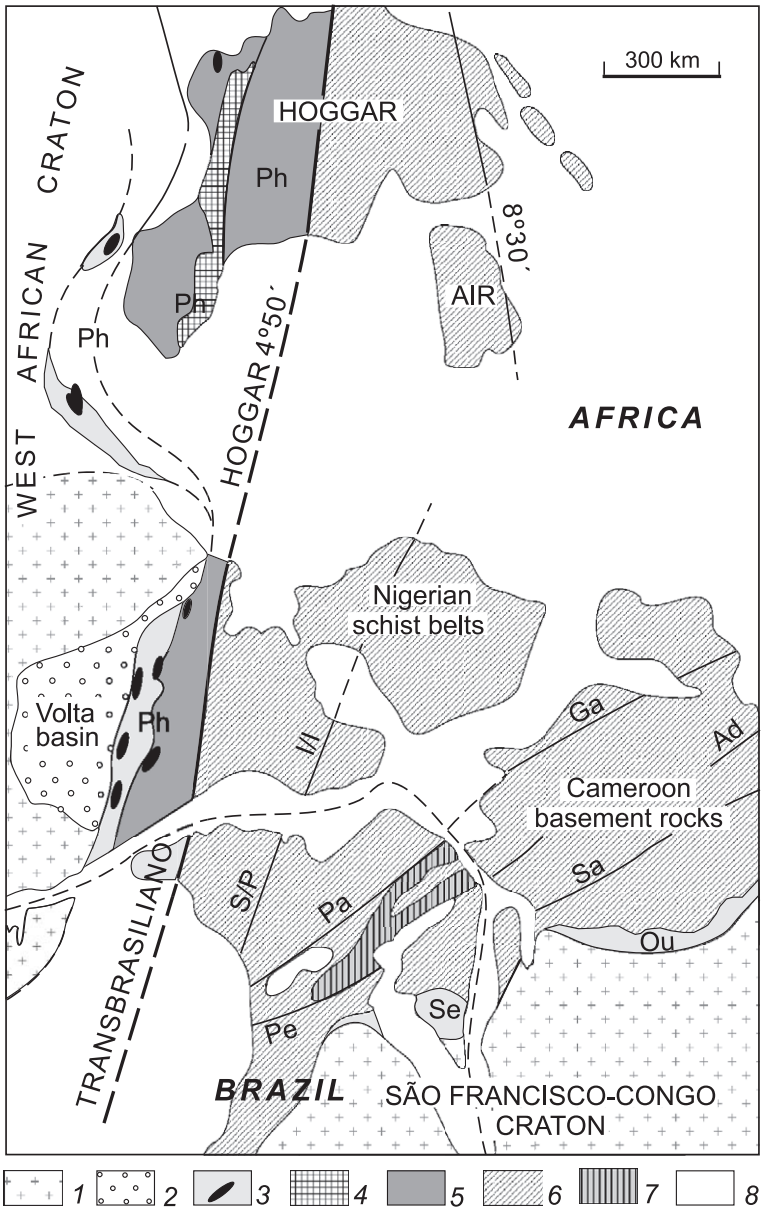


Fig. 6. Schematic tectonic framework of the Brasiliano/Pan-African domain of NE Brazil and Central-West Africa around the Gulf of Guinea (mostly after Trompette, 1994; Caby, 1989). (1) Cratonic domains; (2) Neoproterozoic rocks of the Volta basin; (3) marginal fold belts of Neoproterozoic age: Ph—Pharusian, with mafic–ultramafic bodies; Se—Sergipano; Ou—Oubanguides; (4) reactivated Eburnean high-grade basement rocks within Pharusian belt; (5) orthogneisses and related rocks of granitoid magmatic arcs within Pharusian belt; (6) reactivated polymetamorphic basement of the Borborema/Trans-Saharan domain, intruded by Brasiliano/Pan-African granitoids; (7) Cariris Velhos orogenic belt; (8) Phanerozoic sedimentary covers. Main lineaments, usually major shear zones: S/P—Senador Pompeu; Pa—Patos; Pe—Pernambuco; I/I—Ile–Ife; Ga—Garoua; Ad—Adamaua; Sa—Sanaga.

related to the formation of Rodinia or whether alternative tectonic configurations are possible.

3.1. Southern Ethiopia

Most of the low-grade rocks of western and southern Ethiopia are considered to be part of the Arabian–Nubian Shield, whereas the higher-grade gneissic rocks adjacent to these assemblages were considered as older (Archaean to Mesoproterozoic) basement (Kazmin, 1971). However, recent zircon dating (Teklay et al., 1998; Yibas et al., 2002, and references therein) has demonstrated that there are no Archaean to Mesoproterozoic rocks in western and southern Ethiopia, and ages range from ~900 to 580 Ma for pre- and syntectonic, arc and collisional, granitoid gneisses, followed by ~550–530 Ma post-orogenic granites. This age range is similar to that observed in the ANS of Arabia and Sudan and also to high-grade granitoid rocks in Tanzania and Madagascar, and favour the interpretation that the Neoproterozoic high-grade rocks are lower crustal equivalents of the arc/ophiolite assemblages in the ANS. There is no record of Kibaran-age (1300–1000 Ma) igneous, metamorphic or tectonic activity in this part of Africa.

3.2. Kenya

The extensive review on the evolution of the MB in Kenya by Mosley (1993) still holds true, and the only additional information comes from an Austrian group working in the Taita Hills of SE Kenya near the Tanzania border (Loizenbauer et al., 2000). The oldest rocks in the MB appear to be the Mukogodo nebulitic migmatites in north-central Kenya for which Key et al. (1989) obtained a three-point Rb–Sr isochron age of 1206 ± 96 Ma, which is of questionable significance and was interpreted to reflect a Kibaran high-grade event. Additional evidence for the possible existence of a Kibaran-age "basement" comes from Nd mean crustal residence ages of 1100–1125 Ma for the Sekerr metapelites (Harris et al., 1984) exposed near the Archaean craton margin in NE Kenya. The extent and nature of this "basement" is not known, and the assumption of Shackleton (1986) that a pre-Pan-African basement underlies much, if not all, of the MB in Kenya remains open to debate.

There are no precise zircon ages for the MB of Kenya, but considering the sequence of events recorded in southern Ethiopia by Yibas et al. (2002), it is unlikely that there are rocks older than about 900 Ma. Key et al. (1989) reported Rb–Sr isochron and errorchron "ages" of 826 ± 33 and 818 ± 48 Ma respectively for gneisses and a granite, and Mosley (1993) suggested that a first phase of granitoid magmatism, deformation and metamorphism lasted from ~830 to ~730 Ma ago. This is similar to the situation in southern Ethiopia, Tanzania (see Section 3.3) and Madagascar (see Section 3.6). In summary, there is as yet no reliable zircon age substantiating the existence of a Kibaran-age basement in the MB of Kenya.

3.3. Eastern and southern Tanzania

The upper amphibolite to granulite-facies gneiss terrane of the Mozambique belt in east-central Tanzania contains rocks ranging in age from late Archaean to late Neoproterozoic (Muhongo et al., 2001; Sommer et al., in press). Nd model ages from an E–W traverse across the MB in northern Tanzania are all late Archaean (Maboko, 2000) and probably represent reworked remnants of the Tanzania craton, whereas in the Usambara Mountains of NE Tanzania, they are Neoproterozoic (Fig. 5) and similar to the Arabian–Nubian Shield in NE Africa and Arabia (Maboko and Nakamura, 2002). The boundaries of the Archaean rocks with the 840–720 Ma Neoproterozoic gneiss terrane of eastern Tanzania are not known because of poor outcrop conditions but are thought to be tectonic, probably defined by narrow ductile shear zones. The Neoproterozoic assemblage has remarkable similarities in age and tectono-metamorphic evolution with the basement of central Madagascar (Kröner et al., 2000a; Collins et al., 2003) and southern Ethiopia. The peak of granulite-facies metamorphism was determined at ~625–640 Ma on metamorphic zircons (Muhongo et al., 2001; Sommer et al., in press) and monazite (Möller et al., 2000), and the PT evolution of the high-grade rocks seems to have followed both clockwise (Sommer et al., in press) and counterclockwise PT trajectories (Appel et al., 1998), whereas the petrology of rocks in southern Kenya is more compatible with a clockwise PT path (Hauzenberger et al., 2000).

Recent fieldwork along a traverse from Morogoro to Iringa has revealed the presence of a major flat-lying ductile shear zone of predominantly Archaean and Palaeoproterotoic rocks with consistent top-to-the-west shear sense (Sommer et al., in press). This is more than 1 km thick and probably constitutes the western margin of the MB against virtually undeformed granites of the Mesoproterozoic Usagaran domain of central Tanzania. This boundary is also present, though less pronounced, in southwestern Tanzania and may link up with the Mwembeshi shear zone in central Malawi. No rocks in the age bracket 1300–1000 Ma have so far been discovered in northern and central Tanzania.

The protoliths of several granitoids and migmatitic granitoid orthogneisses in southern Tanzania around the town of Masasi have magmatic emplacement ages in the range 630–799 Ma (Fig. 5) and are thus comparable with those farther north and in Madagascar (Kröner et al., 2003a). However, there are also rare gneisses with primary magmatic crystallization ages of 1009 and 1094 Ma, and these may belong to an Irumide-age cycle of magmatic activity that is also found farther south in neighbouring Mozambique (Kröner et al., 2000b; see Section 3.4). There are no marked structural boundaries between these isolated Irumide-age rocks and the younger Pan-African assemblages since all rocks in the Masasi area share the same intense ductile fabric.

3.4. Northern Mozambique

Reconnaissance zircon dating of Precambrian rocks from the Tete and Manica Provinces in northwestern Mozambique (Kröner et al., 2000b) shows that granitic gneisses and charnockites north of Tete and at Lake Cabora Bassa have magmatic emplacement ages between 1004 and 1046 Ma (Fig. 5). These ages correspond well with U–Pb and Rb–Sr ages for other rocks in the Tete region (Evans et al., 1999). The Irumide-age province seems to end SW of Tete, since emplacement of strongly deformed and partly mylonitized pink granitic gneisses at the Ruenya River south of Changara was dated at 802 Ma, and the shear fabric is cut by pegmatites dated at 551 Ma (Kröner et al., 2000b). However, farther south, along the road from Manica to Chimoio, ~1110 Ma granitoid gneisses confirm the presence of further Kibaran-age magmatic rocks (Manhica et al., 2001).

Jamal et al. (1999) dated zircons from igneous and metamorphic assemblages in the Lurio belt of NE Mozambique where previous zircon and Rb–Sr dating had established the presence of Kibaran-age granitoid rocks in the age range 1000–1100 Ma (Sacchi et al., 1984; Costa et al., 1994; Kröner et al., 1997). The zircon data of Jamal et al. (1999) scatter considerably in the concordia diagram but suggest an age range of 956–1082 Ma for the emplacement of granitoid and pegmatitic rocks, whereas a strong metamorphic event is evident from lower concordia intercept ages of 538–606 Ma.

The above age data confirm previous suggestions for a major magmatic event of Irumide (not Kibaran) age in NW Mozambique (Pinna et al., 1993), but there is as yet no conclusive evidence for Irumide-age metamorphism and deformation in this region. The magmatic event may correlate with similar-age rocks in the central part of the Zambezi belt in northern Zimbabwe where granitic gneisses were emplaced at 1005–1018 Ma (Dirks et al., 2000), but, like in Mozambique, no Irumide-age metamorphism or deformation has been recognized. These data also suggest that there may be a closer genetic link between the Zambezi and Mozambique belts than hitherto assumed, and that both belts reveal Irumide and Pan-African components whose significance in terms of orogenic evolution is still poorly understood (see also Rivers and Johnson, in press).

Kröner et al. (2001) suggested that the Irumide-age igneous province of northern Mozambique may link up with the Wanni Complex of Sri Lanka (see below) in a Gondwana configuration, whereas Jacobs et al. (1998) proposed that these rocks link up with Kibaran-age gneisses in Dronning Maud Land of East Antarctica.

3.5. Southern Malawi

Single zircon U–Pb and Pb–Pb ages and Nd isotopic systematics for granitoid orthogneisses and metapelites from southern Malawi collected around Lilongwe and farther south in the region around Blantyre and Zomba (Fig. 5), document three distinct events: (1) an Irumide-age period of intrusion of calc-alkaline granitoids around 1040–929 Ma; (2) a Pan-

African period of intrusion of calc-alkaline granitoids around 710–555 Ma; and (3) a long-lasting thermal peak of Pan-African high-grade metamorphism around 571–549 Ma (Kröner et al., 2001). The chemistry of the dated granitoid gneisses is compatible with an origin of the protoliths in an Andean-type active continental margin setting, and Nd isotope data indicate overall juvenile compositions consistent with an arc environment, or emplacement into thinned continental crust with little involvement of older basement. However, the data of Kröner et al. (2001) are insufficient to determine whether the magmatic activity was quasi-continuous from ~1040 to ~555 Ma or whether there were two distinct plutonic events separated by ~250 Ma. Rocks with presumably rather different histories prior to the thermal peak were stacked together. An earlier Irumide-age granulite metamorphism, as postulated by Andreoli (1984), could not be verified. The late Irumide and Pan-African magmatic events in Malawi may be genetically linked to granitoid magmatism of similar ages in NW Mozambique, southern Tanzania and in the central Zambezi belt of northern Zimbabwe, but the geodynamic relationships between these terranes remain obscure.

3.6. Central and southeastern Madagascar

Several distinct terranes have recently been identified in Madagascar and are separated from each other by major tectonic boundaries (Collins et al., 2000b, 2003; Collins and Windley, 2002) (Fig. 5). The majority of high-grade rocks in central Madagascar consists of 820–720 Ma old granitoid orthogneisses (Kröner et al., 2000a), and there is a distinct lack of Kibaran or Irumide ages, making it unlikely that Madagascar played an active role in Rodinia aggregation. The similarity in rock types and ages between eastern Tanzania and Madagascar suggests that the island was part of, or close to, present East Africa during the accretion history of the Gondwana supercontinent. There is no similarity in rock type, structural evolution and geochronology between most of southern India and Madagascar, and any suture between East and West Gondwana must have been situated east of Madagascar. Peak metamorphic conditions in Madagascar were reached at 560–550 Ma, and there is no evidence of a 640–620 Ma metamorphic event as in East Africa. Nd isotopic data demonstrate that the 820–720 Ma granitoids were largely generated through melting of Palaeoproterozoic and Archaean crust (Kröner et al., 2000a).

3.7. Sri Lanka

In contrast to the terranes discussed above, the high-grade metamorphic basement of eastern and western Sri Lanka consists of ~895–1100 Ma old Grenville (=Irumide)-age calc-alkaline granitoid rocks that are exposed in the Wanni (western Sri Lanka) and Vijayan (eastern Sri Lanka) Complexes (Kehelpannala, 1997; Kröner, 1991, 2003b) (Fig. 5). This suggests that the island was involved in a subduction/accretion event at around 1000 Ma, but there is only scanty evidence for major deformation at this time since virtually the entire basement of Sri Lanka was ductilely deformed at high temperatures during the pervasive Pan-African event between ~600 and 550 Ma ago. It is possible, but by no means certain, that Sri Lanka may have played a key part in Rodinia aggregation, and that the voluminous granitoids produced during the ~1000 Ma event reflect arc formation and accretion onto the marginal part of the evolving supercontinent. In addition, granitoid rocks in the age range ~750–790 Ma are also present in the Wanni Complex (Kröner and Jaeckel, 1994, and references therein) and probably document the same important Pan-African magmatic event that is present in Madagascar and East Africa. The peak of granulite-facies metamorphism in Sri Lanka was dated at 560–550 Ma (Kröner and Williams, 1993; Kröner et al., 1994).

The similarity in ages between the Wanni Complex of Sri Lanka and the Lurio belt of northern Mozambique has given rise to speculations linking these two terranes (Kröner et al., in press). However, detailed studies comparing the lithologies and structural evolution of these terranes are not available, and reconstructions based on similarities in ages alone may be misleading.

3.8. Southern India

The gneiss–granite–greenstone terrane of the Archaean Dharwar craton is separated from the high-

grade gneiss terrane of southernmost India, generally known as the Southern Granulite Terrane (SGT), by the Neoproterozoic crustal-scale Moyar shear zone (Meissner et al., 2002). The SGT is made up of two crustal domains of contrasting composition and evolution known as Madurai Block, predominantly of late Archaean to Palaeoproterozoic age, and the Kerala Khondalite belt, predominantly of Neoproterozoic (Pan-African) age (Braun and Kriegsmann, 2003). These domains consist of charnockitic to enderbitic gneisses and their retrogressed equivalents as well as gneiss–migmatite terranes, all interlayered with high-grade supracrustal sequences, intruded by granitoid plutons of Neoproterozoic age and transected by major shear zones (Drury et al., 1984; Naqvi and Rogers, 1987; Jayananda and Peucat, 1996; Meissner et al., 2002; Braun and Kriegsmann, 2003). The tectonic evolution of the SGT is still poorly understood because of lacking structural investigations and precise age data.

Braun and Kriegsmann (2003) reviewed the geochronology of the SGT and concluded that the Madurai Block and Kerala Khondalite belt experienced a common tectono-metamorphic evolution in Neoproterozoic times, culminating in granulite-facies metamorphism at 650–550 Ma and extensive granitoid magmatism at 590–525 Ma. It is therefore likely that this part of the Indian Peninsula was involved in terrane accretion to form the Mozambique belt (or East African Orogen) in Pan-African times. There is only one indication for late Grenville-age granitoid magmatism in southern India, and this comes from strongly discordant zircons of a granite–gneiss xenolith in the Kerala Khondalite belt with a poorly defined upper concordia intercept age of 987 ± 65 Ma (Gosh, 1999). This is in strong contrast to the ubiquitous Grenville-age magmatism in Sri Lanka and favours a Sri Lanka–Mozambique–East Antarctica connection over a connection with southern India (Kröner et al., in press).

We therefore conclude that the data for southern India provide no evidence for involvement in an accretion or collision event at the time Rodinia is supposed to have formed. In line with this conclusion, Pisarevsky et al. (2003) placed southern India into a position far removed from the core of Rodinia at the time of supercontinent amalgamation (Fig. 6).

4. The Kibaran and Irumide belts of east-central Africa

The Kibaran mobile belt extends for more than 1500 km from Katanga (SE Congo) to eastern Uganda (Fig. 2) and consists of multiply deformed supracrustal rocks and large volumes of granitoids ranging in age between 1.4 and 1.35 Ga (Kokonyangi et al., 2001a,b; Kampunzu, 2001; Rivers and Johnson, in press). The early granitoids, now orthogneisses, have geochemical features of continental arc magmas (Kampunzu et al., 1986). An intense deformation event at 1.2–1.1 Ga ago resulted in the formation of a fold-and-thrust belt and is generally ascribed to collision of the Congo/São Francisco Craton with ill-defined "East African cratons" such as the Tanzania craton, the poorly known Bangweulu Block of northern Zambia and the equally poorly defined West Nile craton in northern Uganda (Kampunzu, 2001) (Fig. 2). However, there are also views interpreting the Kibaran belt as entirely intracratonic (e.g., Pohl, 1992). A review of the Kibaran belt favouring a collisional origin is provided by Rivers and Johnson (in press).

Kampunzu (2001) related the 1.4–1.35 Ga Kibaran belt to Rodinia assembly but provides no information as to the exact nature of the tectonic setup and why this amalgamation should have begun 200 Ma earlier than elsewhere in the supercontinent. In contrast, Pisarevsky et al. (2003) show that, on palaeomagnetic grounds, the Congo/São Francisco Craton was far removed from the other continental masses making up Rodinia (see Fig. 6), and it is therefore unlikely that the formation of the Kibaran belt was related to supercontinent assembly.

The NE-trending Irumide belt is parallel to the Kibaran belt (Fig. 2) and evolved on the SE margin of the Palaeoproterozoic Bangweulu Block in Zambia and northern Malawi (Cahen et al., 1984; Daly, 1986; De Waele and Mapani, 2002). Its NE continuation in Tanzania is not known, likewise, its SW continuation in Zambia and NE Botswana is unknown, although Hanson (2003) and Kampunzu et al. (1998) argued for a connection with the Ghanzi–Chobe belt in NE Botswana via the Choma–Kalomo block (see below). This connection has now been shown untenable on the basis of new zircon ages

(De Waele et al., 2003). The SE margin of the Irumide belt is severely overprinted by Pan-African deformation and metamorphism.

Contrasting tectonic models have been proposed for the Irumide belt such as collisional (Daly, 1986; De Waele and Mapani, 2002) and accretionary (Rivers and Johnson, in press), but both are circumstantial and partly rely on data outside the Irumide belt proper. Magmatism is poorly constrained chronologically and seems to have lasted from ~1050 to 950 Ma (De Waele et al., 2003). As in the case of the Kibaran belt, no conclusive evidence exists for the generation of the Irumide belt as part of Rodinia assembly.

5. The Sinclair and Rehoboth magmatic provinces of Namibia

Volcanic and intrusive rocks of low metamorphic grade and interpreted as a ~1100 Ma calc-alkaline magmatic province known as the Sinclair Group are exposed along the western margin of the Kalahari craton (Fig. 2) on the fringe of the Namib Desert in southern Namibia (Watters, 1978; Hoal, 1993; Hoal and Heaman, 1995). These rocks have broadly been related to the Kibaran–Namaqua period of tectogenesis, but no genetic relationship with either the high-grade Namaqua belt farther south or the granitoid-dominated Kibaran belt of east-central Africa has been established. The little deformed Sinclair Province appears to have developed on the edge of the Kalahari craton and was not involved in subsequent collisional processes.

Farther north, in south-central Namibia, a belt of NE–SW-trending mixed sedimentary and magmatic rocks known as Rehoboth Basement Inlier (Ziegler and Stoessel, 1993; see Fig. 2) has been considered to constitute the continuation of the Sinclair Province (Watters, 1978), but here again no genetic relationship has been established. On the contrary, the Rehoboth Inlier consists of highly variable rocks dating back to 2.6 Ga, intruded by 1.8–2.0 Ga calc-alkaline granitoids, a 1.4 Ga mafic complex and 1.0–1.25 Ga granitoids and rhyolites (Ziegler and Stoessel, 1993). Its link with the ~1100 Ma Ghanzi–Chobe belt of northwestern Botswana (Fig. 2) (Kampunzu et al., 1998) is equally uncertain.

6. Namaqua–Natal belt of southern Africa

The Irumide-age Namaqua and Natal belts (Fig. 2) have always been considered to constitute one single mobile belt, interrupted only by the thick cover of Phanerozoic deposits of the Cape and Karoo Supergroups. However, their rock composition, tectono-metamorphic evolution and the precise timing of metamorphic events are all different, and it is therefore uncertain whether both belts developed in the same tectonic setting. Eglington and Armstrong (in press), for example, concluded from zircon and Nd model ages obtained for basement rocks in deep boreholes penetrating the Phanerozoic cover in Cape Province that there must be a terrane boundary separating the western (Namaqua) from the eastern (Natal) sector of the belt.

The Natal belt, for instance, is characterized by thin-skin tectonics, contains remnants of ophiolitic rocks and has been thrust over the Kaapvaal craton during a major shortening event beginning prior to 1135 Ma (Jacobs et al., 1997). The peak of granulite-facies metamorphism has been dated at 1050–980 Ma ago (Thomas et al., 1993; Jacobs et al., 1997).

In contrast, the Namaqua belt is a broad zone of predominantly high-temperature and ultra-high-temperature granitoid gneisses and interlayered metasediments of shallow-water origin. Metamorphic temperatures reached 1050 °C at places (Waters, 1990), a feature that cannot be due solely to crustal stacking as a result of collisional tectonics (Harley, 1989). The peak of UHT metamorphism was reached 1060–1030 Ma ago (Robb et al., 1999).

The position of the Kalahari craton (Kalaharia), to which the Namaqua–Natal belt is attached, is uncertain with respect to the other continental masses that make up Rodinia. At least four different positions have been shown in the last few years (for discussions, see Powell et al., 2001a; Pisarevsky et al., 2003). Dalziel et al. (2000) suggested that the Namaqua–Natal belt resulted from collision of Kalaharia with the Llano Uplift in Texas (Laurentia) during Rodinia assembly. However, Powell et al. (2001a,b) and Pisarevsky et al. (2003) have shown that there was a latitudinal separation of some 30° between Kalaharia and Laurentia at about 1100 Ma (see Fig. 6) and that the Natal–Namaqua belt was facing outward and away from Laurentia. An alternative model was proposed by Bruguier et al. (1999) and

Powell et al. (2001b) who suggested that Kalaharia collided with the western margin of Australia at around 1080 Ma. In summary, the role of Kalaharia and the Namaqua–Natal belt in the formation of Rodinia, if any, is uncertain, and models relating the Namaqua–Natal belt to collisional tectonics during Rodinia assembly (e.g., Unrug, 1992; Bruguier et al., 1999; Dalziel et al., 2000) are entirely speculative.

7. Amazonia and Laurentia

In most reconstructions of Rodinia (e.g., Hoffman, 1991; Unrug, 1996; Sadowski and Bettencourt, 1996), Amazonia is placed against the eastern side of Laurentia by matching the Grenville and Sunsas mobile belts, tectonically active between 1300 and 1000 Ma (Figs. 1 and 4). They are considered as products of continental collision during the process of amalgamation of Rodinia. Whereas the high-grade Grenville belt is well developed in eastern North America, no high-grade rocks of Grenvillian age are found in the Sunsas belt of western Brazil and Bolivia (Fig. 3). Moreover, the Sunsas belt is better described as having formed in an extensional environment and consists of a passive margin sedimentary sequence that was deformed under low to medium grade metamorphic conditions and intruded by syn- to late-tectonic granitoids (Litherland et al., 1989). The Grenville/Sunsas link is far from resolved. For instance, there are other reconstructions than discussed above in which, although keeping the general correlation for the ca. 1000 Ma mobile belts, Laurentia and Amazonia are placed in different positions (Starmer, 1996; Keppie and Ortega-Gutierrez, 1999) in order to accommodate other smaller cratonic fragments such as Oaxaquia of southern Mexico, Arequipa–Antofalla of western South America and the Garzón Massif of Colombia.

In their interpretation of the Laurentia–Amazonia link, Sadowski and Bettencourt (1996) considered a possible correlation between the Elsonian and Rondonian–San Ignacio events (approximately 1.5–1.3 Ga) in addition to the already proposed correlation between the younger Grenville and Sunsas orogenies. Their proposed reconstruction for the latter leaves the main thrust belt with high-grade rocks over the Superior Province of Laurentia, but also considers extensional systems over the Amazonian craton, including a magmatic arc affecting the Sunsas system and a back thrust represented by the Aguapei belt. Their conception of the regional tectonic evolution includes a former Mesoproterozoic link of Laurentia and Amazonia within a large continental mass prior to 1.5 Ga. Its break-up, by means of a first Wilson cycle, formed an ocean floor between Laurentia and Amazonia, which was later consumed in a subduction event (Rondonian–San Ignacio), with the ocean disappearing towards the Amazonian side. A very large granitoid domain (the Pensamiento granites) is recognized as the magmatic arc produced in that orogeny (Fig. 3). It consists of dominantly potassic calc-alkaline granitoids, with ages roughly between 1.4 and 1.3 Ga, and a mixed Sr isotopic signature, indicating a significant participation of mantle material in their petrogenesis (Tassinari et al., 2000). A second Wilson cycle is envisaged, with the formation of another ocean floor, consumed by subduction once more towards Amazonia, represented by the Sunsas orogeny. However, only a small amount of granitoid material is associated with it.

Therefore, the preferred model of Sadowski and Bettencourt (1996) is the onset of two Wilson cycles, virtually on the same site, producing a reasonably large oceanic floor during the first cycle (about 1.5 Ga) and only small ocean opening during the second cycle (about 1.25 Ga). The final closing of the latter ocean produced an oblique collision, associated with the Grenville and Sunsas orogenies, the former producing a high-grade collisional belt on Laurentia's side and the latter a series of transtensional features over the border of Amazonia. The above authors also envision a left-lateral transpressional component, which is consistent with many tectonic features observed on both continents (Cordani et al., 2003).

It is apparent from the above account that the Grenville and Sunsas mobile belts are very different in lithology and tectonic evolution, and their proposed correlation is based solely on similarities in age and on the possibility of a correlation as proposed by various authors, in particular Sadowski and Bettencourt (1996). According to the latter, Laurentia and Amazonia were very close together, since at least 1.5 Ga, when they were part of the same continental mass, until about 580 Ma, when Laurentia separated from Gondwana. This is not incompatible with the present

palaeomagnetic evidence (e.g., Pisarevsky et al., 2003), but we must consider that the available poles from South America are very scanty (Cordani et al., 2003). A correlation of the Sunsas tectonic province of Amazonia with the cratonic fragment Pampia is made possible by the Grenville-type ages obtained on igneous and metamorphic rocks located along the present western side of Pampia (Ramos et al., 2000).

8. Central Brazil

The Tocantins Province of Central Brazil (Fig. 3) is considered as a mosaic of tectonic blocks that were fused together in the process of amalgamation of Gondwana (Brito-Neves and Cordani, 1991). The main tectonic element in the region is a very large fault zone, considered a megasuture that was active during the entire Neoproterozoic, the Trans-Brasiliano lineament, which crosses the entire region from NE to SW (Fig. 3). It separates the tectonic units related to the Amazonian and São Francisco cratons, and, farther south, it separates the Rio de la Plata craton from the Pampia terrane (Ramos et al., 2000).

Marginal belts are identified at the borders of both continental masses, evolving from passive margins with thick sedimentary piles: the Araguaia belt on the border of the Amazonian craton, and the Brasilia belt on the border of the São Francisco Craton. In addition, a series of juxtaposed terranes with variable lithologies, bounded by shear zones of different sizes, make up a unit formed by tectonic blocks, named the Central Goiás Massif and considered as a single heterogeneous cratonic fragment by Brito-Neves and Cordani (1991). Ultramafic bodies interpreted as remnants of oceanic lithosphere occur within the Araguaia marginal belt (Alvarenga et al., 2000) and also at the western margin of the Pampia cratonic fragment (Ramos et al., 2000).

In close association with the Trans-Brasiliano megasuture, Pimentel and Fuck (1992) characterized a large and elongated region with calc-alkaline granitoid orthogneisses and associated rocks, the Goiás Magmatic Arc (Fig. 3). The principal rock types of this unit are plutonic, ranging in composition from tonalite to granodiorite. They are associated with several supracrustal belts in which volcano-sedimentary sequences with mafic and ultramafic bodies, considered as possible ophiolite fragments, were described (Pimentel and Fuck, 1987; Pimentel et al., 1997). Many Neoproterozoic post-orogenic granitic intrusions are also found in this region. The geochronological systematics of the Goiás Magmatic Arc indicate an age interval between 930 and 530 Ma. The juvenile signature of the granitoid rocks is demonstrated by their low initial Sr isotopic ratios, their positive $\varepsilon_{Nd(t)}$ values and their Sm–Nd T_{DM} mean crustal residence ages between 0.9 and 1.2 Ga. (Cordani and Sato, 1999; Pimentel et al., 2000). The interpretation of the latter authors is that the Goiás Magmatic Arc formed, between 930 and 600 Ma ago, at the roots of a series of juvenile, intra-oceanic, island arcs that resulted from consumption of oceanic lithosphere. The Goiás arc implies the existence of a large oceanic domain, the Goiás Ocean, at least since 930 Ma, between the Amazonian and the São Francisco/Congo cratons.

The following domains are recognized within the Central Goiás Massif: the late Archaean terranes of Crixás, the Palaeoproterozoic gneisses and supracrustal belts of Almas–Dianópolis, the Anápolis–Itauçu high-grade complex and the three large layered mafic–ultramafic complexes of Cana Brava, Niquelândia and Barro Alto. These mafic–ultramafic units are remarkably similar and are considered to have originally formed in a continuous magmatic structure, possibly a large continental rift. They comprise an older, lower layered series of gabbro–norites, pyroxenites and dunites, and a younger, upper layered series in which gabbro–anorthosites predominate. The three mafic–ultramafic complexes are in contact with medium grade metavolcano-sedimentary sequences in their western part and were affected in their eastern part by high-grade metamorphism.

Fuck et al. (1989) obtained Rb–Sr whole-rock isochron ages of about 1.3 Ga from the Barro Alto Complex and interpreted these as reflecting the high-grade metamorphic event. However, similar Mesoproterozoic ages were obtained from the Barro Alto Complex by Suita et al. (1994) by dating magmatic zircons and by Ferreira-Filho and Pimentel (2000) for the Niquelândia Complex, using the Sm–Nd whole-rock isochron method. These authors interpreted the above ages to reflect igneous crystallisation of the mafic–ultramafic rocks. The complete evolutionary history of the mafic–ultramafic complexes is far from resolved, considering that Fuji (1989) obtained a Sm–

Nd whole-rock isochron age of about 2000 Ma for the older series of the Cana Brava Complex, and Correia et al. (1997) obtained SHRIMP zircon ages of ca. 2000 Ma for the Niquelândia Complex.

In contrast, the age of metamorphism seems to be better constrained. Different authors, employing several radiometric methods, established the main regional metamorphic event to have occurred between 750 and 790 Ma (e.g., Ferreira-Filho et al., 1994, 1998). Pimentel et al. (2000) considered this high-grade metamorphic event observed at the eastern portion of the mafic–ultramafic bodies to be related to the collision between the Central Goiás Massif and the São Francisco Craton.

In the southern part of the Central Goiás Massif, there is the high-grade metamorphic complex of Anápolis–Itauçu where Fischel et al. (1998) obtained Neoproterozoic Sm–Nd garnet/whole-rock metamorphic ages in the range 630–610 Ma. These ages are probably related to collision between the São Francisco Craton and the Paraná cratonic fragment, as suggested by Campos-Neto (2000).

Therefore, the known examples of high-grade metamorphic rocks all appear to be Neoproterozoic, probably due to collisional events linked to the amalgamation of Gondwana. As a corollary, these units do not seem to have had any role in the making up of Rodinia (Cordani et al., 2003). The Paraná cratonic fragment mentioned here is kept separate from the larger Rio de La Plata craton (Fig. 3). Recent work (Cordani et al., 2001a) indicated that the southern Precambrian area of Paraguay at the Tebicuary River has the same geochronological pattern as the Brasiliano/Pan-African Ribeira belt (Trouw et al., 2000) and should therefore be considered as an extension of this belt below the sedimentary rocks of the Paraná Basin. Thus, the attachment of the Paraná block to the Rio de La Plata craton, as envisaged by Campos-Neto (2000) and Alkmin et al. (2001) among others, is not confirmed for Neoproterozoic times.

9. Northeastern Brazil and its extension to West Africa

The Borborema Province of northeastern Brazil (Fig. 6) is a mosaic of Archaean and Palaeoproterozoic blocks that were involved in the Neoproterozoic tectono-magmatic events of the Brasiliano/Pan-African orogeny during the process of amalgamation of West Gondwana (Brito-Neves et al., 2000, and references therein). Petrological, structural and geochronological features support a strong correlation with the Trans-Saharan belt in West Africa, a complex region consisting of an old basement strongly affected by the Pan-African orogeny (the *ghost craton* or *metacraton* of Abdelsalam et al., 2002). This correlation was already convincingly demonstrated by Hurley et al. (1967), de Almeida and Black (1968), Torquato and Cordani (1980), Caby (1989) and Trompette (1994). The combined Borborema/Trans-Saharan continental mass (Fig. 6) is very large and consists essentially of ancient basement, for which recent isotopic data indicate that Palaeoproterozoic material predominates, including some Archaean nuclei (Caby, 1987; Brito-Neves et al., 2000). This large terrane is intruded by numerous and voluminous Brasiliano/Pan-African granitoids, for which the Sr and Nd isotopic signatures indicate derivation by melting of older crustal material. This old terrane, about the same size as the West African or Rio de La Plata cratons, was almost always forgotten in published reconstructions of Rodinia (e.g., Figs. 1 and 4). However, it was considered in the recent reconstructions for the Neoproterozoic by Alkmin et al. (2001) and Cordani et al. (2001b).

With the possible exception of the Cariris Velhos orogenic belt, to be considered below, confirmed Mesoproterozoic terranes have not been described from the Borborema/Trans-Saharan domain. The metamorphic belt in Nigeria, suggested by Grant et al. (1972) as possibly Kibaran, has been demonstrated by U–Pb zircon ages as being much older (Ajibade et al., 1987). Other metamorphic belts in the Borborema/Trans-Saharan domain are either related to the ubiquitous Brasiliano/Pan-African orogeny or to metamorphic events older than 1700 Ma (Caby, 1987; Brito-Neves et al., 2000, Trompette, 1994).

However, the Cariris Velhos belt in northeastern Brazil could be an exception. It occupies the so-called Transversal Zone, located between the Patos and Pernambuco shear zones (Fig. 6). It exhibits Grenvillian radiometric ages (1000–950 Ma) and was considered by Van Schmus et al. (1995) as possible evidence for the Borborema Province to have been

part of Rodinia. The belt has a Palaeoproterozoic basement and consists of a series of volcano-sedimentary, plutonic and metamorphic units along a 750 km long terrane, which reaches the Atlantic coast near the cities of João Pessoa and Recife. The geochemistry of the igneous rocks points to a mature magmatic arc environment. U–Pb zircon and Rb–Sr whole-rock ages are generally in the 980–950 Ma range. The main granitoid rocks are syenogranites and monzogranites with Nd T_{DM} mean crustal residence ages between 1.2 and 1.7 Ga, and positive to slightly negative $\varepsilon_{Nd(t)}$ values, indicating juvenile, mantle-derived material, mixed with some older crust (Cordani and Sato, 1999; Cordani et al., 2000). The metamorphic rocks are mainly amphibolite-facies orthogneisses, in association with migmatites, but some high-grade sillimanite schists have also been described, making up the Sertania Complex on the eastern side of the belt (Brito-Neves et al., 2000). These authors consider the Cariris Velhos orogeny as resulting from the operation of a short-lived Wilson cycle, affecting an old continental terrane that may have been part of Rodinia.

It must be emphasized that the entire region has been through strong deformation and reworking by the Brasiliano/Pan-African event. In our view, the Cariris Velhos magmatic rocks were generated during an extensional event at about 1.0 Ga, the time of the Grenville orogeny, when Rodinia is considered as having been in the process of amalgamation. However, since the main structural trends of the Cariris Velhos belt are coherent with the general structural trends of the Borborema Province, deformation and metamorphism of the belt may well have occurred much later, during amalgamation of Gondwana.

10. Eastern and southeastern Brazil and corresponding parts of Africa

The entire Mantiqueira Province in eastern and southern Brazil is considered to be part of a large Mesoproterozoic mobile belt in the reconstruction of Unrug (1992) (Fig. 4). The province includes the Araçuai and Ribeira belts in Brazil and is correlated with the West Congo belt in western equatorial Africa (Trompette, 1994). These structural domains comprise the northern branch of a large orogenic system connected by the closure of the Adamastor Ocean and formation of West Gondwana and were strongly affected by tectonic and metamorphic episodes of the Brasiliano/Pan-African orogeny.

The Araçuai and West Congo belts evolved in what appears to be a large cratonic embayment carved into the São Francisco/Congo continental mass in early Neoproterozoic times. It formed by crustal thinning and extension between the Brazilian and African parts, but the São Francisco and Congo cratonic fragments were not separated by a large ocean, and the Araçuai and West Congo belts are ensialic (Pedrosa-Soares and Wiedemann-Leonardos, 2000). Their basement is almost entirely Palaeoproterozoic or older. In addition, it appears that several volcano-sedimentary basins, filling extensional aulacogen-type structures, occur in many places, and some yielded late Mesoproterozoic radiometric ages. In Africa, the 1000–910 Ma Zadinian and Mayumbian Groups of Lower Congo were described in that position below the sedimentary succession of the Pan-African West Congo belt (Trompette, 1994; Tack et al., 2001). In Brazil, such structures are related to the Espinhaço belt, which is exposed over large areas within the São Francisco Craton in central Bahia State of Brazil (da Costa and Inda, 1982). The extensional events began at about 1770 Ma, and filling of the Espinhaço basin produced a thickness of more than 5 km of sedimentary material. The unit was affected by low-grade and locally medium-grade metamorphism, dated at about 1300–1400 Ma (Cordani et al., 1992). However, high-grade Grenvillian-age belts are not found in the Mantiqueira–West Congo provinces. Trompette (1994), Unrug (1996) and Tack et al. (2001) interpreted the Mesoproterozoic sedimentary basins on both sides of the Atlantic as aulacogen-type structures that were deformed in the Neoproterozoic.

The southern part of the Mantiqueira Province includes the Ribeira belt, which is exposed in southeastern Brazil. It is a large and complex orogenic system that developed in response to the closure of the Adamastor Ocean by convergence of the São Francisco/Congo, Kalahari and Rio de La Plata cratons, as well as the smaller cratonic fragments of the Paraná and Luiz Alves blocks (Campos-Neto, 2000). This author described the tectonic evolution of the Ribeira belt as comprising a series of successive collisional

events, all occurring in the Neoproterozoic. High-grade belts of Grenvillian age are completely absent. Some local tectonic slices within this belt were attributed to the Mesoproterozoic, but their ages are disputed, and they are essentially described as metasedimentary basins and cover successions affected by the Brasiliano orogenic cycle (Trouw et al., 2000).

Much like the Araçuai belt, the Ribeira belt also seems to have evolved on older crust (Trouw et al., 2000) and exhibits an old Palaeoproterozoic to Archaean continental basement. It is also invaded by numerous granitoids of Brasiliano/Pan-African age. According to Cordani et al. (2000), the Brasiliano/Pan-African granitoids of the Mantiqueira Province generally exhibit negative $\varepsilon_{Nd(t)}$ values, indicating at least partial derivation from ancient crustal material and suggesting that the Adamastor Ocean must have been relatively narrow.

11. Southern Brazil and Uruguay

The Neoproterozoic Dom Feliciano belt of southern Brazil and Uruguay is developed at the eastern margin of the Rio de La Plata craton. The corresponding orogens in southern Africa are the Kaoko–Damara and Gariep belts (Fig. 2), developed at the western margin of the Congo and Kalahari cratons (Torquato and Cordani, 1980; Trompette, 1994).

The Namaqua–Natal high-grade belt of Grenvillian age constitutes the basement of the Gariep belt at the western margin of the Kalahari craton. Within the Rio de La Plata craton, a Mesoproterozoic domain has not yet been fully characterized. Tectono-thermal events at about 1250 Ma were indicated for the Nico Perez terrane (Basei et al., 2000). However, within the Dom Feliciano belt in Uruguay, the Punta del Este terrane is made up mainly of 900–1000 Ma granitoid orthogneisses that were strongly reworked during the Brasiliano orogeny (Preciozzi et al. 1999). These rocks may correlate with the Namaqua belt of South Africa.

Moreover, most Pan-African granitoids in the Dom Feliciano belt yielded Nd T_{DM} mean crustal residence ages between 1.6 and 1.2 Ga and moderately negative $\varepsilon_{Nd(t)}$ values, suggesting derivation from magmas in which reworked crustal material was always present in significant proportions (Babinski et al., 1997; Basei et al., 2000). As a corollary, this again confirms that the Adamastor Ocean between the Rio de La Plata and Kalahari cratons could not have been very wide.

12. Grenville-age crustal fragments within Phanerozoic belts of Mexico and the Andean Cordillera

Many of the basement units of Mexico and the Andean Cordillera of South America yielded Grevillian radiometric ages of about 1000 Ma, a fact that induced their correlation with, and the consequent palinspastic reconstruction of, a large collisional belt produced during the formation of Rodinia. These fragments had thus been considered as extensions of the Grenville Province of Laurentia (Ballard et al., 1989) or the Sunsas belt of Amazonia (Restrepo-Pace et al., 1997).

The relatively small basement outcrops in southern Mexico (Oaxaca and Guichicovi complexes) comprise layered mafic and felsic granulites that are supposed to form a contiguous granulite-facies metamorphic terrane of about 1000 Ma. The Pb isotopic composition of the overlying Mesozoic and Cenozoic volcanic rocks indicates a close affinity with the underlying basement (Keppie and Ortega-Gutierrez, 1995; Ruiz et al., 1999). This similarity in Pb isotopic systematics suggested derivation of the Pb from an extensive Grenvillian basement, and the coherent Pb isotopic signature over a large area was taken as the main evidence for the existence of Oaxaquia, a crustal terrane whose blocks shared a common U–Pb history in the late Proterozoic (Ruiz et al.,1999). Moreover, the same authors favoured a close correlation in age and lithology between Oaxaquia and some of the northernmost basement outcrops of the Andean Cordillera. The correlative units are the high-grade Santa Marta and Garzón massifs in Colombia, exhibiting Grenvillian radiometric ages (Tschanz et al., 1974; Alvarez, 1981; Kroonenberg, 1982; Restrepo-Pace et al., 1997). In particular, anorthosites, relevant in the Oaxaca Complex, are also found in both Colombian units.

Following on a former idea of Kroonenberg (1982), Restrepo-Pace et al. (1997) suggested that Grenvillian basement occurs below the entire Eastern Cordillera of Colombia, and such a large terrane could be the extension of the Sunsas belt of Amazonia (Litherland

et al., 1989). Restrepo-Pace et al. (1997) also suggested possible links with the southern terranes of the Arequipa massif in Peru (Wasteneys et al., 1995) and the Precordillera in Argentina (Ramos, 1988).

Such correlations are based essentially on similarities in radiometric age and include a great deal of speculation. Ramos and Aleman (2000), for instance, described a number of small basement terranes with Grenvillian ages in the Andean regions of southern Peru, western Bolivia, northern Chile and northwestern Argentina. They interpreted them as parts of Rodinia that were left behind in South America when Laurentia separated from Gondwana in the latest Proterozoic during opening of the Iapetus Ocean. However, the radiometric evidence is very scanty, based on a few age determinations from outcrops or from basement xenoliths in younger volcanic rocks. A review of the available evidence on the South American fragments can be found in Ramos and Aleman (2000) and Ramos (2000).

A correlation of all Grenville-age terranes scattered in Mexico as well as in the Andean basement of northwestern South America, and their possible comparison with the Grenvillian-age belts of Laurentia, Amazonia and Baltica, was attempted by Keppie and Ortega-Gutierrez (1999). Their conclusion was that there is a close similarity between the tectonic evolution of terranes occurring in southern Mexico, Colombia and in the Adirondack Mountains of North America. Following the model of Starmer (1996), these terranes may initially have formed at 1200 ± 100 Ma within a large magmatic arc that was generated by consumption of a Mesoproterozoic ocean located between Laurentia on one side and Amazonia plus Baltica on the other. Subsequently, these terranes collided during the Grenvillian orogeny.

Even if the model of Keppie and Ortega-Gutierrez (1999) is accepted, the reconstruction of the original position of the fragments is made difficult by the complex regional tectonic history during the Neoproterozoic and Phanerozoic. If all of these terranes were part of Rodinia, they were kept relatively stable after amalgamation at about 1000 Ma, until about 600 Ma. Then they were involved in the latest Proterozoic continental break-out, when Laurentia separated from Gondwana and Iapetus was formed. They also took part in the closure of Iapetus, forming Pangaea in Palaeozoic times, and, finally, they took part in the disintegration of Pangaea and were affected by all the tectonic processes related to the construction of the Andes that partially consisted of terrane accretion in recent times. Precise palaeogeographic reconstructions are not possible and would require much additional geological, geochronological, palaeomagnetic and geophysical data. However, we question whether all of these terranes actually took part in the collisional events that produced Rodinia, because many of them seem to have formed during extensional events at passive margins or in plate interiors. Only the southern Mexican sites (Oaxaca and Guichicovi complexes) and some Colombian basement complexes, such as the Garzon Massif and parts of the Santa Marta Massif, are high-grade units of confirmed Grenvillian age, that could have been involved in collisional events during amalgamation of Rodinia.

13. Palaeomagnetic evidence

Meert and Powell (2001), in an editorial to a special volume on the assembly and break-up of Rodinia, noted that the few available palaeomagnetic tests for the existence of a supercontinent at ~1100 Ma provide mostly negative evidence. Powell et al. (2001b) arrived at a similar conclusion after careful evaluation of published palaeomagnetic data and noted that the African and South American continental blocks, perhaps with the exception of the Kalahari craton, were not involved in the formation of Rodinia. In particular, the Congo/São Francisco Craton was far removed from all the other continental blocks at ~1100–1000 Ma. Pisarevsky et al. (2003) presented a series of palaeogeographic maps showing a palaeomagnetically permissible reconstruction of Rodinia at the end of the Mesoproterozoic (Fig. 7). In their reconstruction, Congo/São Francisco and India are positioned independently and quite distant from Rodinia at 1100 and 1060 Ma. Moreover, they interpret the Irumide belt of east-central Africa (southeastern margin of Congo) as a collage of sedimentary and island-arc-related intrusive rocks with zircon ages between c. 1150 and 1000 Ma and consider this region as facing an open ocean undergoing passive arc accretion during the Mesoproterozoic until ~1000 Ma. Although Pisarev-

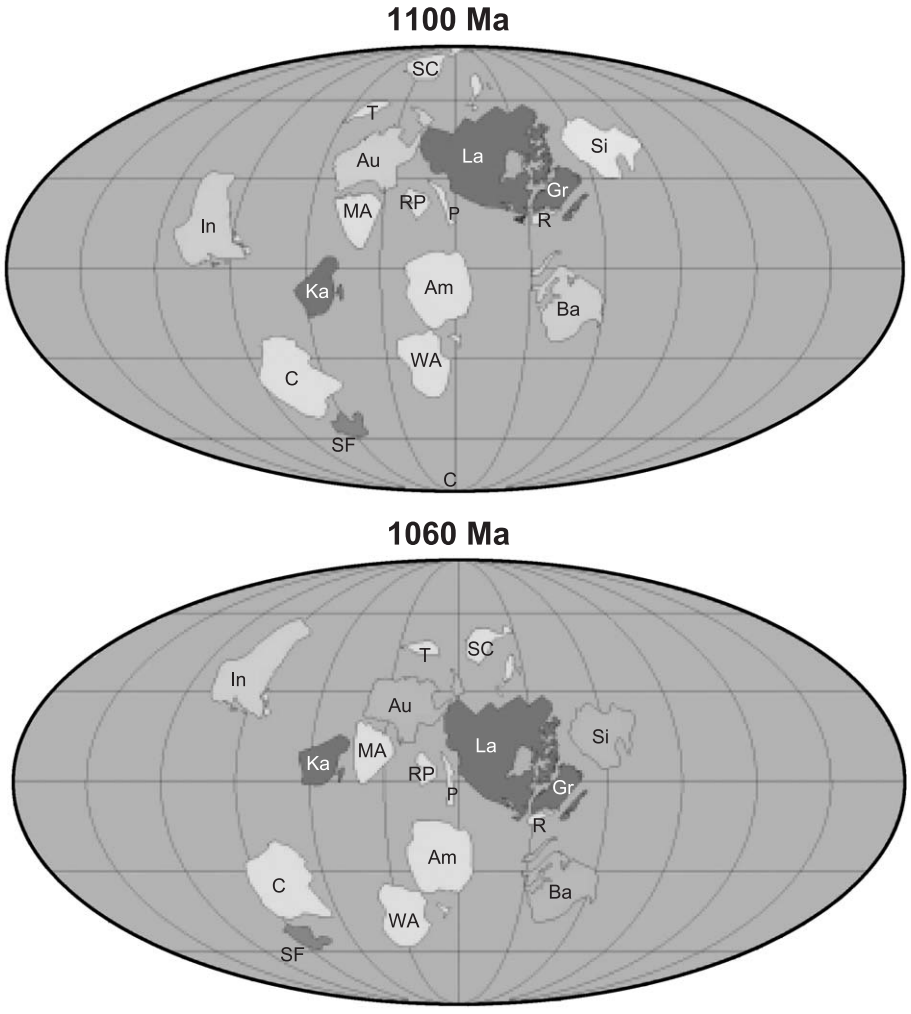


Fig. 7. Global palaeogeography at 1100 and 1060 Ma during formation of the supercontinent Rodinia (from Pisarevsky et al., 2003). Continental blocks positioned according to palaeomagnetic data and rotation poles in Powell and Pisarevsky (2001). Mollweide projections. C = Congo, SF = São Francisco, Ka = Kalahari, Au = Australia, Ma = Mawson craton, La = Laurentia, Gr = Greenland, S = Siberia, Am = Amazonia, W = West Africa, Ba = Baltica, SC = South China, In = India.

sky et al. (2003) prefer a configuration in which the Congo/São Francisco moved as an independent plate (Fig. 6), they specifically point out that this solution is poorly constrained.

Pisarevsky et al. (2003) proposed three separate South American blocks in a continental reconstruction for the period 1000–750 Ma, namely (1) Rio de La Plata, (2) the Pampean terrane and (3) the Paraná block. In a review of Brazilian palaeomagnetic and geochronologic data Cordani et al. (2003) concluded that a wide Brasiliano/Pan-African ocean separated the continental fragments of Africa and South America from Laurentia during the period ~1100–630 Ma.

It seems clear from the palaeomagnetic data that the African and South American cratonic blocks were far removed from the core of Rodinia during the formation of the Grenville-age mobile belts, making it unlikely, as already argued above on geological and geochronological grounds, that these belts resulted

from amalgamation of the supercontinent (Cordani et al., 2003).

14. Discussion and conclusions

14.1. The Brazil–Africa connection: tectonic significance of Neoproterozoic juvenile magmatic arcs and the Pampean–Goiás–Pharusian–Arabian–Nubian Ocean

As already mentioned, the Goiás Magmatic Arc of Brazil includes a series of 900–600 Ma granitoid rocks, formed during successive episodes of intra-oceanic subduction. This large amount of juvenile crustal material, located close to the Neoproterozoic megasuture of the Trans-Brasiliano lineament in central Brazil, is the main evidence for the existence of the Goiás Ocean, which must have had a considerable size. Therefore, in latest Mesoproterozoic to early Neoproterozoic times, at least since about 1000 Ma ago, the Goiás Ocean separated Amazonia from the São Francisco/Congo craton plus other smaller cratonic fragments such as the Central Goiás Massif and the Paraná block (Fig. 3). Campos-Neto (2000) considered the Goiás Ocean to have formed during what he called "Tonian-age taphrogeny" (about 1000 Ma ago), and he associated this event with the break-up of Rodinia. Obviously, this is incompatible with the current views on the time of disintegration of Rodinia, assumed to have occurred much later. Moreover, the radiometric ages of 1000–1100 Ma for mafic dykes within the São Francisco Craton (D'Agrella-Filho et al., 1990) also point to a major event of disruption that occurred at about the same time as the Grenville orogeny and formation of Rodinia.

To the northeast, the Trans-Brasiliano megasuture can be traced into the basement structures below the Palaeozoic Parnaiba sedimentary basin (Brito-Neves et al., 1984) and can be extended into West Africa where it correlates with the Hoggar 4°50′ lineament (Figs. 3 and 6). Adjacent to this tectonic zone, a large region consisting of granitoid orthogneisses formed within a magmatic arc environment (Caby, 1989), and a good example occurs within the Tilemsi intra-oceanic arc of Mali (Caby et al., 1989). These calc-alkaline rocks belong to the Pharusian belt and exhibit Neoproterozoic ages up to about 880 Ma. Their character indicates the existence of a large ocean (Caby, 2002), named here for convenience "Pharusian Ocean", that can be correlated with the Goiás Ocean of Pimentel et al. (2000). As already envisaged by Castaign et al. (1994), the Pharusian Ocean separated the West African craton from the Borborema/Trans-Saharan cratonic fragment in the Neoproterozoic.

To the southwest, the Trans-Brasiliano lineament is concealed below the Paraná sedimentary basin but is considered to extend at least into Paraguay and northern Argentina (Brito-Neves et al., 1984). Therefore, we think that it is permissible to consider the Neoproterozoic granitoid rocks occurring in the Pampean ranges of Argentina and making up a considerable part of the Pampia cratonic fragment (Fig. 3) as extensions of the Goiás Magmatic Arc and having formed above an intra-oceanic subduction zone. In this area, a large ocean separated Pampia from the Rio de La Plata craton (Ramos, 1988).

From the above, we envisage the existence of a very large oceanic domain, in the early Neoproterozoic, separating the core of Rodinia, namely Amazonia plus West Africa and Baltica, from a series of continental masses and fragments such as the São Francisco/Congo, Rio de La Plata and Kalahari cratons, plus the Borborema/Trans-Sahara, Central Goiás Massif and Paraná block. This large oceanic domain would include the well documented Goiás and Pharusian oceans as well as the less well-defined Pampean Ocean. Moreover, it is possible that the Pharusian Ocean may have been connected, below the Saharan territories of North Africa, with the large oceanic domain in which the Arabian–Nubian Shield evolved between ~850 and 650 Ma ago. Considering what has been referred to in Sections 3.1, 3.2 and 3.3, the Arabian–Nubian magmatic arcs extended into Ethiopia, Kenya and Tanzania, making up what has been named the Arabia–Nubia–Ethiopia–Kenya–Tanzania (ANEKT) Ocean (Unrug, 1996).

Most of the rocks that formed in the roots of intra-oceanic magmatic arcs are calc-alkaline granitoids (tonalites, granodiorites and potassic granites) exhibiting juvenile isotopic signatures. Their Sr, Nd and Pb isotopic data indicate material formed in intra-oceanic environments, by subduction of oceanic lithosphere. The oldest ages of such granitoids in the Goiás Magmatic Arc are in the 950–900 Ma interval (Pimentel and Fuck, 1992; Pimentel et al., 1997),

whereas in the Arabian–Nubian Shield, they are some 50–100 million years younger at 850–800 Ma (Kröner et al., 1992; Stern, 1994). This indicates that the oceanic lithosphere that was consumed in the production of the granitoid rocks must have been generated at least some 100 Ma earlier, at a time that is attributed to the "Grenville orogeny" and amalgamation of Rodinia. In other words, when Rodinia formed, a large ocean separated several continental masses from it. The Pampean–Goiás–Pharusian–Arabian–Nubian Ocean was only consumed later, in the Neoproterozoic, in the process of amalgamation of Gondwana.

In contrast, when examining the isotopic signatures of most Neoproterozoic granitoids in the Brasiliano/Pan-African provinces of eastern Brazil and Africa, they clearly demonstrate a significant proportion of remelting of older crustal material. Relevant mantle contributions are rare or absent (Cordani and Sato, 1999; Cordani et al., 2000). This is true for the Borborema/Trans-Saharan province, the Central Goiás and Paraná fragments and for the belts resulting from consumption of the Adamastor Ocean in Namibia (Damara–Kaoko, Fig. 2) and southeastern Brazil (Ribeira). In our view, this may be due to the relative proximity of the continental masses that collided in the Neoproterozoic, when the Brasiliano/Pan-African belts formed, or to magmatic arcs forming on continental margins adjacent to narrow ocean basins.

In other words, the Congo/São Francisco, Kalahari, Rio de La Plata, Central Goiás, Paraná, and Borborema/Trans-Sahara were always relatively close together since the latest Mesoproterozoic and through the entire Neoproterozoic, after the Espinhaço and Kibaran orogenies. As a consequence, we suggest that, at about 1000 Ma, the above continental masses and fragments, which are now parts of Brazil and Africa were separated from the other cratonic blocks making up the supercontinent Rodinia.

14.2. East Africa, Madagascar, southern India and Sri Lanka

Our review of geological and geochronological data for eastern, southern and central Africa (including Madagascar), southernmost India and Sri Lanka, as summarized above, supports the conclusion of Pisarevsky et al. (2003), and we see no compelling evidence for the involvement of the southern African cratons, Madagascar and southern India in the formation of a late Mesoproterozoic supercontinent. The only crustal domain within the broad Mozambique belt showing a *major* crust-forming event at Grenvillian times is Sri Lanka.

The ~1400 to ~950 Ma Kibaran and Irumide belts of east-central Africa most likely formed through collisional and/or accretionary processes, but there is no convincing evidence, certainly not from palaeomagnetism, that this was related to Rodinia amalgamation. Although there are Grenville-age magmatic rocks in parts of East Africa such as northern Mozambique, southern Tanzania and southern Malawi, it is not clear whether these rocks constitute remnants of a once continuous belt, possibly connected with the Irumides, and are thus directly related to a Grenvillian-age tectono-metamorphic event (Rivers and Johnson, in press) or whether they represent continental fragments resulting from Rodinia break-up and dispersal, or whether their formation is unrelated to Rodinia. The limited information available on these rocks and their isolated occurrence within the younger, Pan-African, gneisses makes it impossible, at this stage, to correctly determine their original tectonic setting. The vast majority of high-grade granitoid rocks in the Mozambique belt of East Africa are Pan-African (~850 to ~570 Ma) or early Mesoproterozoic to Archaean (~1600–2700 Ma), such as in Tanzania and Madagascar. Extensive granitoid magmatism in the time range ~850–650 Ma is characteristic of all East African, Malagassy and western Sri Lankan (Wanni) crustal domains discussed above and is primarily calc-alkaline. It may be related to processes of continental aggregation leading to the formation of the supercontinent Gondwana.

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