Palaeoproterozoic to Neoproterozoic growth and evolution of the eastern Congo Craton: Its role in the Rodinia puzzle

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Abstract

The Central African Cratons comprise various Archaean and Palaeoproterozoic blocks, flanked or truncated by orogenic belts ranging in age from Palaeoproterozoic (Rusizian, Ubendian and Usagaran Belts) to Mesoproterozoic (Kibaran and Irumide Belts). These various orogenic systems map out the progressive nucleation of the Central African Cratons to form the Congo Craton, which during late Neoproterozoic times participated in various collisional processes to form part of the Gondwana supercontinent. Subsequently, the opening of the South Atlantic separated a small portion from the Congo Craton, which now forms part of the South American cratonic assemblage and is referred to as the S\textsuperscript{˜}ao Francisco Craton. The original continuity of the S\textsuperscript{˜}ao Francisco and Congo Craton is supported by similarities in basement ages and craton stabilisation during Eburnean-aged tectonothermal events and the recognition of the original unity of the Arac\textsuperscript{¸}uai and West Congo Belts and the Sergipane and Oubanguide Belts across the Atlantic. The nucleation of the Congo Craton from its composing cratonic blocks, which include the Angola-Kasai Block, the NE-Congo-Uganda Block and the Cameroon-Gabon-Congo-S\textsuperscript{˜}ao Francisco Block to the west and northwest of the Mesoproterozoic Kibaran Belt, and the Bangweulu Block and Tanzania Craton, to the east and southeast, was at the latest completed after peak compressional tectonism in the Kibaran Belt at 1.38 Ga. Late Mesoproterozoic tectonism along the southern margin of this proto-Congo Craton, in a region called the Irumide Belt, marks compressional tectonism at ca. 1.05–1.02 Ga, which produced extensive reworking along this margin, possibly linked to the participation of the Congo Craton in the Rodinia Supercontinent. At present, insufficient evidence is available to support or deny the participation of the Congo Craton in Rodinia.

During the early Neoproterozoic, several rifting events occurred along the southern margin of the Congo Craton, in the Luflitan and Zambesi Belts, with localised volcanism and deposition of clastic sequences (Roan and Mwashya Groups), and followed by passive margin sedimentation (Nguba and Kundelungu Group). These sequences also contain large diamictite horizons (Grand and Petit Conglom\textsuperscript{é}rat). At ca. 570–530 Ma, convergence with the Kalahari Craton to the south and the Malagasy-Indian Cratons to the East culminated in collisional processes that formed the Damara-Luflitan-Zambesi and the East African Orogens, and led to the formation of Gondwana.

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1. Introduction

The Congo Craton is defined as the amalgamated central African landmass at the time of Gondwana assembly (ca. 550 Ma). Along its southern and eastern margins it comprises various Archaean blocks including the Angola-Kasai Block and the Tanzania Craton (Fig. 1). These Archaean units have been extensively affected by Palaeoproterozoic events between 2.2 and 1.9 Ga, collectively referred to as the “Eburnean” events, and which may also have reworked a cryptic Archaean terrane to form the Palaeoproterozoic Bangweulu Block (De Waele, 2005). By Palaeoproterozoic times, the various blocks of the proto-Congo Craton had stabilised and were subsequently affected by Mesoproterozoic convergent tectonism forming the Kibaran, Irumide and Southern Irumide Belts. During the Mid-Neoproterozoic, rift successions and subsequent passive margin deposits developed along the southern margin of the Congo Craton prior to collisional events leading to the amalgamation of Gondwana (Johnson et al., 2005). In this paper, we will review the Palaeo- to Neoproterozoic tectonic evolution of South-Central Africa where (1) Palaeoproterozoic events formed the Rusizian-Ubendian Belt along western and south-western
Fig. 1. Simplified geological map of Sub-Saharan Africa (based on Hanson, 2003). Abbreviations as follows: CKB, Choma Kalomo Block; Gab. Belt, Gabon Belt; LM, Lake Malawi; LT, Lake Tanganyika; LV, Lake Victoria; MB, Magondi belt; NE Kib. Belt, Northeast Kibaran Belt; R. Belt, Ruwenzori Belt; Rus., Rusizian; SF, São Francisco Craton; S.M., Southern Malawi; Ub. Belt, Ubendian Belt; Us. B, Usagaran Belt. Note that the Kilimafedha and Sukumaland Greenstone Belts are not distinguished from the Tanzania Craton.

The Angola-Kasai Block comprises two poorly known regions of central Africa, the Angola Block in southern and central Angola, and the Kasai Block, straddling the borders between Angola, the Democratic Republic of Congo (hereafter DRC) and Zambia (Fig. 1). These two blocks are separated by a region of extensive Phanerozoic cover of the Congo Basin, which together with the scarcity of data for either region, makes correlation impossible.

The Angola Block comprises a magmatic complex of orthogneisses and a metasedimentary complex of quartzite and schist, affected by granulite facies metamorphism at ca. 2.8–2.7 Ga (Trompette, 1994). In the south of the complex, in northern Namibia and southern Angola, these basement rocks comprise Archaean to Siderian protoliths (2645–2464 Ma, Seth et al., 1998; Delor et al., 2006) extensively affected by migmatisation events between 2.29 and 1.85 Ga (Cahen et al., 1984; Seth et al., 1998), and accompanied by intrusion of granitoids dated in southern Angola using zircon U–Pb sensitive high-mass resolution ion microprobe (SHRIMP) at 2038 ± 28 and 1959 ± 6 Ma (McCourt et al., 2004), 1987–1968 Ma (Delor et al., 2006) and in Namibia by U–Pb or Pb–Pb isotope dilution thermal ionisation mass spectrometry (ID-TIMS) at 1985–1960 Ma. Sedimentation follows these events with the deposition of the Chela Group, in which an ignimbrite yielded a zircon U–Pb SHRIMP age of 1790 ± 17 Ma (McCourt et al., 2004).
The Kasai Block consists of a complex of granitic rocks yielding whole-rock Rb–Sr dates of between 3.49 and 3.33 Ga, which have been metamorphosed in the granulite facies at ca. 2.8 Ga (Cahen et al., 1984). In northwestern Zambia, Key et al. (2001) reported zircon U–Pb SHRIMP crystallisation ages of 2561 ± 10 and 2538 ± 10 Ma for granitic gneisses, providing the only reliable U–Pb age constraints for the Kasai Block. The same authors also reported a zircon crystallisation age of 2058 ± 7 Ma for a porphyritic granite intruding these Archaean gneisses, indicating that an Eburnean-aged event, similar to that affecting the Angola Block to the West, also affected the Kasai Block.

2.2. The Tanzania Craton

The Tanzania Craton is almost entirely constituted of granitoids and greenstone belts, the latter of which are referred to in the south (central Tanzania) as the Dodoma, and in the north (northern Tanzania, Kenya and southeast Uganda) as the Nyanzian and Kavirondian systems (Cahen et al., 1984). The granitoids and gneisses yielded whole-rock Rb–Sr dates ranging from 2.7 to 2.5 Ga, with some scattered reports of age determinations up to 3.12 Ga possibly attesting to Mesoarchaean components (Cahen et al., 1984). Recent zircon U–Pb SHRIMP determinations on granulite facies granitoids and gneisses in the Mozambique Belt of central Tanzania confirmed emplacement ages of 2.7 Ga, with xenocrystic components of ca. 3.0 and 3.5 Ga (Muhongo et al., 2001; Johnson et al., 2003; Cutten et al., 2006), while 2.7 Ga zircon from an orthogneiss has also been reported from rocks within the Usagaran Belt to the south (Reddy et al., 2003). U–Pb dating by laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) of igneous zircons extracted from rhyolites and granitoids within the Kilimafedha Greenstone Belt (eastern Tanzania Craton) indicate that volcanism and granitoid intrusion occurred between ca. 2.72–2.71 and 2.69–2.65 Ga, respectively (Wirth, 2004). Positive initial \( ^{147} \text{Sm} / ^{144} \text{Nd} \) whole rock Sm–Nd isotopes of both the rhyolites and granitoids indicate derivation from an upper mantle-like source without significant involvement of older basement, i.e., an oceanic arc setting. These ages and initial isotopic ratios are very similar to those obtained for the upper parts of the Sukumaland Greenstone Belt further west, which have been dated by isotope dilution thermal ionisation mass spectrometry (ID-TIMS) on zircon at ca. 2.65 Ga (Borg and Krogh, 1999), although the lower parts of this greenstone belt appear to be significantly older at ca. 2.8 Ga based on a whole-rock Sm–Nd age on mafic volcanics (Manya and Maboko, 2003). Positive initial \( ^{147} \text{Sm} / ^{144} \text{Nd} \) whole rock Sm–Nd isotopes for the basaltic lithologies (Manya and Maboko, 2003) also support a supra-subduction origin for this greenstone belt. Nd isotopic ratios of the Sukumaland metasedimentary lithologies provide evidence for an older crustal sedimentary provenance of ca. 3.0–3.1 Ga (Cloutier et al., 2005).

2.3. The Bangweulu Block

The Bangweulu Block forms a crystalline unit of granitoids and coeval volcanic rocks, unconformably overlain by a continental clastic succession of fluvial and lacustrine conglomerates, quartzites and siltstones called the Mporokoso Group, with minor volcanic intercalations near the base (Fig. 2). The block adjoins the Archaean Tanzania Craton along the Palaeoproterozoic Ubendian Belt (Fig. 1), in which granulite facies metamorphism, constrained through zircon U–Pb and Ar–Ar dating at ca. 2.0 Ga (Ring et al., 1997; Boven et al., 1999), was interpreted to mark collisional processes between the Bangweulu and Tanzania Blocks. Despite this, lithologies on the Bangweulu Block yielded dates no older than 1.9 Ga (Brewe et al., 1979; Schandelmeier, 1980, 1983), indicating that the terrain underwent significant crustal reworking during Ubendian tectonism. Rainaud et al. (2003) reported a significant component of xenocrystic zircons with U–Pb SHRIMP crystallisation ages of ca. 3.2 Ga in a tuff within the Neoproterozoic Lufilian Belt to the West, and proposed that metasediments in the Zambian and DRC Copperbelt region, and possibly the Bangweulu Block, are underlain by a Mesoarchaean terrane they termed the Likasi Terrane. One granite gneiss, which forms part of the basement within the Irumide Belt along the southern margin of the Bangweulu Block, yielded a zircon U–Pb SHRIMP crystallisation age of 2726 ± 36 Ma, and so far represents the only direct evidence of Archaean crust in the region (De Waele, 2005; De Waele et al., 2006a,b).

3. Palaeoproterozoic stabilisation of the proto-Congo Craton

A system of Palaeoproterozoic belts can be traced from the south-east margin of the Tanzania Craton (Usagaran Belt), passing between the Bangweulu Block and Tanzania Craton (Ubendian Belt), and northwards into the Mesoproterozoic Kibaran Belt (Rusizian Rise), marking an important tectonothermal event that could record the amalgamation of the Tanzania Craton and the Bangweulu Block (Fig. 1). Similar high-grade thermal events have been reported for the Kasai and Angola Blocks (Cahen et al., 1984), but because of a lack of reliable data neither of these regions are further discussed below.

3.1. The Usagaran Belt

The Usagaran Belt is comprised of granitoid gneisses, metasedimentary rocks and high-pressure eclogite facies rocks that occupy the southern and southeastern corner of the Tanzania Craton (Fig. 1). The Konse Group (formally Konse Series) is comprised of various greenschist-grade metasediments and metavolcanics that directly overlie the Tanzania Craton and the adjacent Isimani Suite (Mruma, 1995). They are tentatively correlated with the Ndembera volcanics that have a zircon U–Pb SHRIMP age of 1921 ± 14 Ma (Sommer et al., 2005). The Isimani Suite lies to the east of the Konse Group and is comprised of upper amphibolite, granulite and eclogite grade metasedimentary gneisses, orthogneisses and mafic rocks with N-MORB-like geochemical compositions (Möller et al., 1995). Detrital zircon analyses of the metasedimentary gneisses indicate they were sourced from the Tanzania Craton and a 2.6–2.4 Ga source region similar to that of reworked rocks within the East African Orogen (Collins et al., 2004) and an orthogneiss within the
Fig. 2. Simplified geological map of the region outlined in Fig. 1. Abbreviations are as follows: inset political boundary map: DRC, Democratic Republic of Congo; MAL, Malawi; MOZ, Mozambique; TAN, Tanzania; ZIM, Zimbabwe; main map: Chp, Chipata Terrane; C.I, Chewore Inliers; C-R, Chewore-Rufunsa Terrane; HGM, Hook Granite Massif; KnG, Kanona Group; KsF, Kasama Formation; LG, Lusaka Granite; L-N, Luangwa-Nyimba Terrane; LTN, Lake Tanganyika; MfG: Mafingi Group; MrG, Manshya River Group; M.S.Z, Mwembeshi Shear Zone; NgG, Ngoma Gneiss; N.S.Z, Nyamadzi Shear Zone; P-S, Petauke-Sinda Terrane. Southern Irumide Belt divisions after Mapani et al. (2001) and Johnson et al. (2006).
sequence has been zircon-dated by SHRIMP at 2705 ± 11 Ma (Reddy et al., 2003). Most importantly, the Isimani Suite contains N-MORB-type mafic rocks that have been metamorphosed under eclogite facies conditions of 750 °C and 1.8 GPa (Möller et al., 1995) and which were quickly cooled and exhumed to the amphibolite facies (Collins et al., 2004). This subduction-related tectonic event is robustly dated at 1999.5 ± 1.4 Ma by ID-TIMS on monazite (Möller et al., 1995) and at 1999.1 ± 1.1 Ma by SHRIMP on zircon (Collins et al., 2004). The Isimani Suite was exhumed to mid crustal levels by 1996 ± 2 Ma and was exposed at the surface by 1991 ± 2 Ma (Collins et al., 2004). The age of the post-tectonic Kidete Granite, dated by zircon U–Pb SHRIMP at 1877 ± 7 Ma (Reddy et al., 2003), confirms that tectonometamorphism was complete by this time.

The Usagaran Belt to the south and southeast is dominated by granitoid orthogneisses that were partially derived from the reworking and recycling of the Tanzania Craton. Few of these granitoid gneisses have been precisely dated but the majority have whole rock Sm–Nd isotopes consistent with the mixing of mantle-derived material with Archaean crust of the Tanzania Craton. Post-tectonic volcanics of the Ndembera Series and various post-tectonic granitoids have been dated using zircon U–Pb SHRIMP at 1921 ± 16, 1910 ± 11, 1824 ± 17 and 1817 ± 12 Ma (Sommer et al., 2005) indicating that Usagaran convergent-margin tectonometamorphism was complete by ca. 1920 Ma. However, the similarity in ages of crustal components on either side of this supposed suture zone, i.e. Tanzania Craton and Isimani Suite to the west of the eclogitic rocks, and the Western Mozambique Belt to the east, led Reddy et al. (2003) to suggest that only a relatively narrow, Red Sea-type oceanic basin was closed and subducted, leading to a similar pre-rift configuration of crustal components.

3.2. The Ubendian Belt

The Ubendian Belt forms an elongate stretch of granulite–amphibolite facies gneisses and metasedimentary rocks in between the Tanzania Craton and the Bangweulu Block, and is exposed in northern Malawi, along the Zambia–Tanzania border, and along the shores of Lake Tanganyika (Fig. 1). The belt has been subdivided into various elongated blocks, which are bound by NW–SE oriented faults or shear zones, and which have contrasting lithologies and structural features (McConnell, 1972; Daly, 1988; Daly et al., 1989; Lenoir et al., 1994; Theunissen et al., 1996). The structural trends within these blocks generally follow a NW–SE trend, and developed under amphibolite facies conditions (Sklyarov et al., 1998). In some blocks, an E–W trending structure associated with granulite facies metamorphism has been reported, which is truncated by, and thus predates the NW–SE shear zones. Local occurrences of eclogite-facies assemblages have also been reported (Sklyarov et al., 1998). The Ubendian Belt is therefore considered to have undergone a two-stage tectonic evolution, which, although diachronous, could be closely linked to the development of the more easterly Usagaran Belt (Daly, 1988; Lenoir et al., 1994). The earliest deformation occurred between 2.1 and 2.0 Ga, coeval with collisional tectonics in the Usagaran Belt, and resulted in granulite-grade metamorphism, preserved in the southern part of the Ubendian Belt (Lenoir et al., 1994; Ring et al., 1997). A second event occurred at around 1.86 Ga, and involved the emplacement of numerous granitoids (Lenoir et al., 1994), and resulted in the exhumation of high-pressure granulites and eclogites for which a 40Ar–39Ar barroisite cooling age of 1848 ± 6 Ma was reported (Boven et al., 1999). Widespread dextral and sinistral strike-slip shearing has been interpreted as being: (1) Palaeoproterozoic, Daly (1988) suggested that east–west thrusting in the Usagaran Belt was linked to shear zone development in the Ubendian Belt; (2) Late Mesoproterozoic, coeval with the Irumide Orogeny (Ring, 1993); (3) Late Neoproterozoic and an intra-continental response to the East African Orogeny (Ring et al., 2002).

4. Mesoproterozoic tectonic record on the Congo Craton

4.1. The Kibaran Belt

The Kibaran Belt forms a linear NE–SW oriented terrane of amphibolite-grade rocks in between the Tanzania Craton/Bangweulu Block to the south and east, and the Kasai/Ngongo–Uganda Block to the west and north (Fig. 1). The recognition of a basement culmination in the eastern DRC, referred to as the Rusizian (Fig. 1) (Gerards et al., 1971; Cahen et al., 1984; Lavreau, 1985), and partial truncation by the Cenozoic rift of Lake Tanganyika, led to the subdivision of the belt into the Kibaran Belt s.s. of DRC, and the NE Kibaran Belt of eastern DRC, Rwanda, Burundi and northwestern Tanzania (Fig. 1) (Tack et al., 1994). In the past, four separate magmatic events were considered to have occurred during the evolution of the Kibaran Belt, based on whole-rock Rb–Sr data (Klerkx et al., 1984; Fernandez-Alonso et al., 1986; Klerkx, 1987). However, subsequent zircon U–Pb geochronological data have refuted these earlier subdivisions, and have increasingly demonstrated that both parts of the belt underwent a broadly parallel development. Both sections of the Kibaran Belt contain at least two metasedimentary successions, the oldest of which was intruded by S-type granitoids in the north and by intermediate-felsic I-type plutons in the south both of which are dated between 1.38 and 1.37 Ga (Tack et al., 2002; Kokonyangi et al., 2004a). In the Kibaran Belt s.s. this magmatic phase has been linked to supra-subduction magmatism (Kokonyangi et al., 2004a,b, 2005, 2006) while in the NE Kibaran Belt this magmatic/thermal event has been linked to extensional detachment/collapse based on the noted absence of a compressional phase in the structural record (Klerkx, 1987; Klerkx et al., 1993; Fernandez-Alonso and Theunissen, 1998). A second tectonic event affected both sedimentary successions, and is recorded in the Kibaran Belt s.s., with P–T conditions of 740–780 °C and 6.0–6.5 kbar (Kokonyangi et al., 2004a). A zircon U–Pb SHRIMP age of 1079 ± 14 Ma for a syn-kinematic granitoid associated with this event, was interpreted by Kokonyangi et al. (2004a,b) to date this compressional tectonic phase. In the NE Kibaran Belt, limited shear-bounded A-type magmatism has been recorded in the Kabanga-Musongati mafic and ultramafic
alignment (KM hereafter), for which for one intrusion a previous bulk zircon U–Pb age determination of ~1275 Ma (Tack et al., 1994) has recently been revised by a zircon U–Pb SHRIMP age of 1205 ± 19 Ma for the same intrusion (Tack et al., 2002). This latter age could indicate a genetic link between the emplacement of the KM along lateral shear zones in the NE Kibaran Belt and compressional tecnotonics recorded in the Kibaran Belt s.s. by Kokonyangi et al. (2004a,b). Also in the NE Kibaran Belt, post-kinematic Sn-bearing granitoids were emplaced along these ca. 1200 Ma shear zones, which have been recently dated by SHRIMP at 987 ± 6 Ma (Tack, pers. comm.) refining earlier age estimations of ~1.0 Ga by the Rb–Sr method (Cahen and Ledent, 1979; Lavreau and Liégeois, 1982).

4.2. The Irumide Belt

The Mesoproterozoic Irumide Belt is situated along the southern margin of the Palaeoproterozoic Bangweulu Block (Figs. 1 and 2), and is comprised of a deformed basement, folded metasedimentary units and voluminous granitoid intrusions. Exposed parts of deformed basement have exposed parts of deformed basement yielded U–Pb SHRIMP crystallisation ages ranging from 2049 ± 6 to 1927 ± 10 Ma (Armstrong et al., 1999; Ranaud et al., 1999, 2002, 2003; De Waele, 2005; De Waele et al., 2006a,b), with one granite gneiss in the southwestern most part of the Irumide Belt yielding an Archaean age of 2726 ± 36 Ma (Fig. 2) (De Waele, 2005; De Waele et al., 2006a,b). Sm–Nd data on two of these basement gneisses yielded TDM model ages of 3.1–3.2 Ga, indicating a significant crustal residence (De Waele, 2005; De Waele et al., 2006b). This basement is unconformably, and in places structurally, over lain by a metassembled succession of shallow marine quartzites and pelites with minor conglomeratic horizons, referred to as the Manshya River/Kanona Group (Fig. 2) (De Waele and Mapani, 2002). Detrital zircon U–Pb age data for several quartzites from the Manshya River/Kanona Groups (Ranaud et al., 2003; De Waele and Fitzsimons, 2004, 2007; De Waele, 2005) and a quartzite from near the base of the Mporokoso Group on the Bangweulu Block to the north (Fig. 2), indicate the maximum age of deposition for these successions to be ca. 1.8 Ga, while discrete water-lain volcanic units within the Manshya River Group in the northeastern part of the Irumide Belt yielded direct constraints of deposition between 1879 ± 13 and 1856 ± 4 Ma (De Waele and Fitzsimons, 2004, 2007; De Waele, 2005). The TDM model ages of the volcanic units range between 2.8 and 2.2 Ga indicating a mixture of juvenile and reworked crustal material in their generation (De Waele, 2005; De Waele et al., 2006b). The Palaeoproterozoic basement and overlying Palaeoproterozoic supracrustal units were locally intruded by granitoids with a crust-dominated geochemical character with crystallisation ages between 1664 ± 4 and 1551 ± 33 Ma, and TDM model ages between 3.2 and 2.8 Ga indicating significant crustal recycling involved in their genesis (De Waele, 2005; De Waele et al., 2006b). The recognition of a small basin of supermature fluvial sandstones and minor siltstones on the Bangweulu Block, with maximum age of deposition at 1434 ± 14 Ma, based on the youngest concordant U–Pb SHRIMP age of detrital zircon, and detrital patterns similar to those recorded in the Mporokoso Group, indicates that the ~1.6 Ga magmatic event was succeeded by a second cycle of sedimentation resulting in the reworking of the Mporokoso Group (De Waele, 2005; De Waele and Fitzsimons, 2004, 2007). Shear-controlled intrusion of A-type granitoids between 1119 ± 20 and 1087 ± 11 Ma (zircon U–Pb SHRIMP, Ring et al., 1999) within the Ubendian Belt and K–Ar biotite ages as old as 1050 Ma along the Luongo Fold Belt (Daly, 1986) (Fig. 2) herald the stirring of tectonism among the southern margin of the Bangweulu Block. Peak metamorphism in the Irumide Belt is constrained at between 1021 ± 16 and 1018 ± 5 Ma through U–Pb SHRIMP dating of low Th/U zircon rim overgrowths (De Waele, 2005). This tectonic event was accompanied by the intrusion of numerous high-K granitoids between 1055 ± 13 and 1003 ± 31 Ma (De Waele et al., 2003, 2006a; De Waele, 2005) with the bulk of intrusions being contemporaneous with peak metamorphism at ca. 1020 Ma. The geochemistry of these Irumide-age plutons is dominated by crustal signatures, while Sm–Nd isotopic data indicate TDM crustal residence ages of up to 3.3 Ga confirming their recycled nature (De Waele et al., 2006b). The overall geochemical and Sm–Nd isotopic characteristics of all magmatic suites recognised in the Irumide Belt are remarkably similar, suggesting that a broadly homogenous and similar crustal source has been tapped for the generation of each of these magmas. The data therefore suggest that a cryptic Archaean terrane of fairly uniform composition underlies the Irumide Belt, while the oldest TDM model ages of 3.3 Ga give some idea of its possible age. Together with the 3.2 Ga xenocrystic zircon evidence presented by Ranaud et al. (2003), this increasingly supports the existence of a cryptic ca. 3.2 Ga basement terrane in the region possibly both underlying the Bangweulu Block and the Irumide Belt (Fig. 1). This cryptic basement may well represent the enigmatic block that collided with the Tanzania Craton between 2.0 and 1.8 Ga.

4.3. The Southern Irumide Belt

The Southern Irumide Belt (SIB) is a relatively new term (Johnson et al., 2005, 2006, 2007b) used to describe lithologies that occur to the south of, and that are distinct from, the monotonous granitoids of the Irumide Belt (s.s.). The contact between the two belts is obscured by Permo-Triassic grabens and, in places the SIB has been strongly overprinted by Neoproterozoic tectonometamorphism, where structural fabrics merge with the Zambezi and Mozambique Belts (Figs. 1 and 2). In Zambia, this belt has been subdivided into several distinct lithotectonic terranes (Mapani et al., 2001; Johnson et al., 2006, 2007b), some of which are bounded by discrete shear zones. The westernmost terrane, the Chewore-Rufunsa Terrane (Johnson et al., 2006), is comprised of a wide variety of mafic to felsic gneisses and metavolcanic rocks that have calc-alkaline chemistries and trace element, Rare Earth Element (REE) and Nd-isotope signatures that are consistent with them having formed in a continental-margin-arc setting (Oliver et al., 1998; Johnson and Oliver, 2000, 2004; Johnson et al., 2006, 2007b). Magmatic activity in the arc is constrained between ca. 1095 and 1040 Ma (Johnson and Oliver, 2004; Johnson et al., 2005).
although a distinct marginal basin ophiolite (the Chewore Ophiolite) has been recognised in the Chewore Inliers of northern Zimbabwe (Fig. 2) and dated at ca. 1393 Ma (Oliver et al., 1998). Geological investigations of the terranes that lie to the east, the Luangwa-Nyimba, Petauke-Sinda and Chipata terranes (Fig. 2) are at a rudimentary level, but preliminary geochemistry and lithofacies analyses suggest that these terranes may represent the components of an accreted island arc complex or a continuation of the Chewore-Rufunsia continental-margin-arc (Mapani et al., 2001; Johnson et al., 2006, 2007b). Reconciliation SHRIMP geochronology of granitoids and felsic gneisses and migmatites within these terranes indicates that Mesoproterozoic magmatism occurred in a similar time frame to that in the Chewore-Rufunsia Terrane, ca. 1075–1010 Ma but that there are significant older Palaeoproterozoic (ca. 1.9 Ga) and younger Pan African (ca. 0.7–0.5 Ga) components (Johnson et al., 2006, 2007b). SHRIMP dating of metamorphic zircon rims and TIMS dating of metamorphic monazite extracted from high-temperature (>850 °C), low-pressure (<5 kbar) granulite facies lithologies from all the SIB terranes yielded ages between ca. 1080 and 1046 Ma (Goscombe et al., 2000; Schenk and Appel, 2001; Johnson et al., 2006, 2007b) which is contemporaneous with the peak of magmatic activity in these terranes and consistent with magma loading of the crust (Schenk and Appel, 2001). At present there is no convincing evidence that any of these SIB terranes underwent late Mesoproterozoic compressional tectonism.

4.4. Southern Malawi and the Lurio Foreland

Directly to the east of the SIB, Southern Malawi and northern Mozambique are dominated by Late Mesoproterozoic to Early Neoproterozoic calc-alkaline granitoids and gneisses. In Southern Malawi these lithologies have yielded Pb-evaporation ages of between 1040 and 999 Ma, have positive to mildly negative initial \( \varepsilon_{Nd} \) values and have been interpreted to have formed in a continental-margin-arc setting (Kröner et al., 2001). In the Lurio Foreland of northern Mozambique similar calc-alkaline gneisses have been dated by the SHRIMP and Pb-evaporation techniques between ca. 1148 and 1009 Ma (Costa et al., 1994; Kröner et al., 1997; Kröner, 2001). These lithologies lack inherited Archaean and Palaeoproterozoic zircons, and show positive to mildly negative initial \( \varepsilon_{Nd} \) values and have been interpreted to have formed in a mature island-arc-type setting (Jamal and De Wit, 2004). The lithofacies assemblage of the Zambezi and Lufilian Belt sedimentary sequences in both the Zambezi and Lufilian Belt rests unconformably upon these volcanics/basement gneisses (Mallick, 1966) indicating that there had been uplift and tilting shortly after, or accompanying, volcanism/plutonism (Fig. 3b and c). It is apparent that uplift and the formation of a dynamic topography, especially in the Copperbelt Region, was critical for controlling local oxidising–reducing conditions thus allowing the deposition of the extensive syn-sedimentary Cu–Co minerals (Binda, 1994; Cailleux et al., 2005; Sutton and Maynard, 2005). The lithofacies assemblage of the Zambezi and Lufilian Belt sedimentary sequences are remarkably similar (Fig. 3a–c) but the most striking feature are the presence of thousands of randomly oriented, isolated, variably metamorphosed but non-deformed gabbro, mafic and ultramafic blocks (Vrana et al., 1975) that occur within the upper marble sequence (the Cheta and Muzuma formations in the Zambezi supracrustal sequence and the Bancroft Group in the Lufilian Belt; Fig. 3a–c). In the Lufilian Belt this marble unit also contains large clasts of the underlying sedimentary succession that along with the mafic units can be interpreted as olistostromes (Johnson et al., 2005). In the Lufilian Belt these mafic blocks display continental-within-plate and E-MORB-type chemistries (Tembo et al., 1999) whilst in the Zambezi Belt they have mid-ocean-ridge (N-MORB) chemistries and isotopic signatures (John, 2001; John et al., 2003, 2004b). In the Lufilian Belt they have been interpreted

5. Neoproterozoic divergence

Neoproterozoic divergence along the southern margin of the Congo Craton is recorded within the Zambezi and Lufilian Belts as two-distinct rift-related volcano-sedimentary and passive margin sequences (Figs. 2 and 3a–c). The first rift to drift phase occurred between 880 and 820 Ma (Johnson et al., 2007a) and is coincident with the deposition of the extensive copper-bearing Roan strata of the Lufilian Belt. The second cycle lacks voluminous volcanic deposits but was initiated at ca. 765 Ma (Key et al., 2001) and is comprised of the Mwashya, Nguba and Kundelungu strata of the Lufilian Belt including two global diamicite horizons. This event records the formation and development of an extensive passive margin.

5.1. Rift cycle 1—the Roan and Zambezi supracrustal rift basins

The onset of rifting is recorded in the Zambezi Belt by the eruption of a 2500 m thick pile of felsic volcanics and volcanoclastics, the Kafue Rhyolite and Nazingwe Formations (Mallick, 1963; Smith, 1963). Three volcanic units from different parts of the volcanic stratigraphy have been dated by the SHRIMP at ca. 880 Ma (Fig. 3b) (Johnson et al., 2007a). The geochemical characteristics combined with negative \( \varepsilon_{Nd}(t) \) isotope ratios indicate that they were generated by mixing/assimilation of juvenile material with older basement gneisses (Johnson et al., 2007a), a scenario consistent with continental thinning and extension. Although currently undated, potential felsic volcanic equivalents have been identified from the Roan Group of the Lufilian Belt around the Luwishi Dome (Porada, 1989), as part of the R.A.T. (Cailleux et al., 1994) and from the lower Roan rocks from southeast Shaba in the Democratic Republic of Congo (Lefebvre, 1989). A granitoid gneiss with similar geochemical compositions to the Kafue and Nazingwe Formation volcanics (Katongo et al., 2004) has been dated by zircon U–Pb SHRIMP at ca. 883 ± 10 Ma (Fig. 3a) (Armstrong et al., 1999, 2005). The rift to drift sedimentary sequences in both the Zambezi and Lufilian Belts rest unconformably upon these volcanics/basement gneisses (Mallick, 1966) indicating that there had been uplift and tilting shortly after, or accompanying, volcanism/plutonism (Fig. 3b and c). It is apparent that uplift and the formation of a dynamic topography, especially in the Copperbelt Region, was critical for controlling local oxidising–reducing conditions thus allowing the deposition of the extensive syn-sedimentary Cu–Co minerals (Binda, 1994; Cailleux et al., 2005; Sutton and Maynard, 2005). The lithofacies assemblage of the Zambezi and Lufilian Belt sedimentary sequences are remarkably similar (Fig. 3a–c) but the most striking feature are the presence of thousands of randomly oriented, isolated, variably metamorphosed but non-deformed gabbro, mafic and ultramafic blocks (Vrana et al., 1975) that occur within the upper marble sequence (the Cheta and Muzuma formations in the Zambezi supracrustal sequence and the Bancroft Group in the Lufilian Belt; Fig. 3a–c). In the Lufilian Belt this marble unit also contains large clasts of the underlying sedimentary succession that along with the mafic units can be interpreted as olistostromes (Johnson et al., 2005). In the Lufilian Belt these mafic blocks display continental-within-plate and E-MORB-type chemistries (Tembo et al., 1999) whilst in the Zambezi Belt they have mid-ocean-ridge (N-MORB) chemistries and isotopic signatures (John, 2001; John et al., 2003, 2004b). In the Lufilian Belt they have been interpreted
as dykes or sills intruded during the initial stages of continental rifting but in the Zambezi Belt they are interpreted to be vestiges of oceanic crust either tectonically emplaced during Pan African collision tectonics (John et al., 2003) or as an ophiolitic mélange at the time of deposition of the marble units (Johnson et al., 2005). The lack of age data for any of the mafic units makes their interpretation difficult, but it is not beyond reason that all the mafic bodies are tectonically related and represent disrupted and boudinaged syn-rifting sills/dykes (Johnson et al., 2007a).

In the Zambezi Belt the sediments are intruded by various metaluminous, monogranitic plutons (Katongo et al., 2004), two of which have been dated at ca. 820 Ma (Fig. 3b and c) (Hanson et al., 1988; Johnson et al., 2007a); thus providing the youngest age for the upper part of the rift succession. The geochemical and isotopic signatures of these granitoids indicate that they originated by the complete recycling of sialic continental crust, but their tectonic significance is unclear (Johnson et al., 2007a). Such granitoids have yet to be identified in the Lufilian Belt.

5.2. Rift cycle 2—the Mwashya-Nguba and Kundelungu rift basin

From 820 Ma there appears to have been a cessation in sedimentation until the extrusion of the Luakela volcanics of the Mwashya Sub-group at 765 ± 5 Ma (Key et al., 2001) (Fig. 3a), heralding renewed rifting in the Lufilian Belt (Porada and Berhorst, 2000). The Mwashya Sub-group is comprised of minor mafic volcanics and felsic tuffs (Fig. 3a) and a series of coarse clastics with abundant olistostromes (Wendorff, 2005a,b), followed by shallow-water argillites and carbonates (Porada and Berhorst, 2000). This group is directly overlain by a diamictite horizon known as the Grand Conglomérat, which is taken to be the base of the Nguba Group (Fig. 3a). This diamictite has a well-constrained age through volcanic horizons that occur in the overlying strata (735 ± 5 Ma zircon U–Pb SHRIMP age) and in the underlying Mwashya Sub-group (765 ± 5 Ma zircon U–Pb SHRIMP age, see Key et al., 2001). The remainder of the Nguba Group is comprised of an extensive and rapidly deepening passive margin sequence. The overlying Kundelungu Group contains a diamictite at its base known as the Petit Conglomérat, for which there are no direct age constraints; however, it has been suggested it may correlate with the global Marinoan Glacial deposited at around 635 Ma (Robb et al., 2002; Condon et al., 2005), which provides a tentative upper age limit for the sedimentary basin. Subduction of oceanic crust and ocean basin closure is dated by the age of eclogite facies metamorphism in the Zambezi Belt between 659 and 595 Ma (John et al., 2004a) with orogenesis culminating in collision between the Congo and Kalahari Cratons at 530 and 520 Ma (data summarised in Johnson et al., 2005) during the assembly of Gondwana.

6. Palaeomagnetic constraints for the position of Congo

The Precambrian palaeomagnetic database of the Congo Craton and its predecessors is relatively poor. A search in the latest version of the Global Palaeomagnetic Database (Pisarevsky, 2005) produced only few reliable data after filtering out poles with large age uncertainty and low reliability. Poles with
The positions of the Congo-São Francisco Craton at 800–750 Ma is constrained by the 795 Ma Gagwe and 748 Ma Mbozi poles (Meert et al., 1995; Deblond et al., 2001). Although detailed palaeomagnetic information is not of a fine-enough scale to determine actual plate velocities, the APWP provided by the few poles in the time period being considered (1100–750 Ma) is adequate to show that (1) the Congo Craton did not develop on the Congo Craton margin but represents the active margin of a separate microcontinental block or craton, along which extensive oceanic crust was subducted ultimately leading to the collision of this terrane/craton with the Congo Craton margin at ca. 1020 Ma. The peak of tectonometamorphism in the Irumide Belt was dominated by crustal thickening processes (~7–8 kbar) as documented by extensive isoclinal folding and thrusting, with large scale reworking of basement gneisses without the addition of juvenile crust (De Waele, 2005; De Waele et al., 2006b). The peak of this magmatic and metamorphic episode post-dates that in the Southern Irumide Belt by ca. 30 million years (i.e. ca. 1020 Ma, De Waele, 2005) and appears to have been initiated once magmatism had ceased in the Southern Irumide Belt. Our favoured interpretation of these events is that the Southern Irumide Belt did not develop on the Congo Craton margin but represents the active margin of a separate microcontinental block or craton, which is very similar to the 748 Ma Mbozi pole. However, a secondary Luakele remanence of uncertain age resembles the 795 Ma Gagwe remanence, casting some doubts on the assumed primary nature of the Gagwe remanence. The 547 Ma pole from the Sinyai Dolerite in Kenya suggests that the assembly of proto-Gondwana was complete by that time (Meert and Van der Voo, 1996).
was not stationary, (2) at the time of collision ca. 1020 Ma, the potential position of the craton within Rodinia can be assessed using the Olivenca and Ilheus palaeomagnetic poles and (3) between ca. 800 Ma (Gagwe pole) and ca. 750 Ma (Mbozi pole) the craton underwent a significant 90° spin, suggesting that during this time the Congo Craton was separate from Rodinia in order to allow such a rotation. With these constraints in mind, several options remain possible, including a detached...
and attached configuration with respect to Rodinia. These two contrasting tectonic scenarios will now be discussed in detail.

7.1. The Congo Craton as an independent craton

The preponderance of Late Mesoproterozoic supra-subduction related rocks in the Southern Irumide Belt and the Lurio Foreland indicate that prior to collisional tectonism in the Irumide Belt at ca. 1020 Ma, the southern margin of the Congo Craton faced an open ocean dominated by juvenile to mature island arcs with possible microcontinental fragments (Fig. 4a). If we presume that the Congo Craton developed independently from Rodinia, then Irumide orogenesis may have been the result of accretion of these arcs/microcontinents to the Congo margin (Fig. 4b), a scenario not too dissimilar to the Ordovician-Silurian arc accretion/collision events along eastern Laurentia, namely the Grampian and Appalachian events (e.g. Oliver, 2001; Murphy and Keppie, 2005). Continental-margin arc magmatism in Southern Malawi dated between 1040 and 999 Ma (Kröner et al., 2001), indicates that subduction-related magmatism continued after arc-accretion (Fig. 4b) supporting the hypothesis that Irumide orogenesis was not related to the collision of this margin with Rodinia. The Gagwe palaeomagnetic pole dated at ca. 795 Ma indicates that the Congo Craton had drifted from mid- to equatorial-latitudes (Fig. 4c) and then sometime before ca. 755 Ma, the Mbozi Complex palaeomagnetic pole indicates that, the craton had undergone a significant 90° rotation (Fig. 4d). This data is most compatible with a freely moving, Rodinia-independent Congo Craton between 800 and 750 Ma but says nothing about the pre-800 position of the craton. Rifting at ca. 880 Ma and again at ca. 765 Ma to form the Roan-Zambezi and Mwashya-Nguba-Kundelungu rift basins, respectively finally resulted in full continental separation by ca. 750 Ma, with the production of new oceanic crust and allowing sufficient space for the 90° rotation of the craton documented by the palaeomagnetic data (Fig. 4d).

7.2. The Congo Craton as part of Rodinia

The four palaeomagnetic poles from the São Francisco Craton generate a slightly curved APWP that is anchored at 1078 ± 18 and 1011 ± 24 Ma by the Olivenca and Ilheus dykes, respectively (Renne et al., 1990) (Fig. 4f). This APWP is similar to the Keeweenawan track of Laurentia (Pisarevsky et al., 2003) suggesting a common motion between the two and could be used to suggest that the Congo was an integral part of Rodinia. A comparison between the two APWP’s allows the possibility of constraining the Congo-São Francisco Craton’s position within Rodinia at the time of collision. Since both APWP’s have a low curvature, both polarity options (poles and antipoles) are permissible (Fig. 4f). If one polarity option is accepted (position A in Fig. 4f), the northern Congo-São Francisco margin is juxtaposed against Australia-Antarctica and causes overlap with the Kalahari Craton (Fig. 4e and I) with the southern margin of the Congo Craton facing an open ocean. If the alternative polarity option is accepted (position B in Fig. 4f) then the Irumide Belt is juxtaposed against the Rodinia margin but the craton itself lies directly on top of West Africa. Trompette (1994, 1997) proposed the existence of a single West Africa-Amazonia-Rio de la Plata mega Craton in the Mesoproterozoic and Neoproterozoic. However, through applying minor movements between the West Africa and Amazonia Cratons a permissible fit can be proposed between the Congo and Amazonia Cratons (compare Fig. 4a and Fig. 4g). If we accept this reconstruction then it is possible that the Southern Irumide Belt may represent a continental-margin arc developed on the Amazonia Craton margin during ocean closure (Fig. 4e). The Lurio Foreland arcs may have developed as intra-oceanic arcs between the Congo and Amazonia Cratons but equally they may represent a northerly extending peninsula of the Kalahari Craton as suggested by Collins and Pisarevsky (2005) (Fig. 4e). This position could also account for the post 1020 Ma supra-subduction magmatism recorded in Southern Malawi.

The formation of the Roan Rift Basin at ca. 880 Ma would represent the earliest record of attempted break-up of Rodinia (Fig. 4g). At present, rifting of this age has not been recognised in any of the other Rodinian blocks (e.g. Li et al., 2004). Although not conclusive, it is most likely that rifting at ca. 880 Ma never resulted in the production of oceanic crust (Johnson et al., 2006, 2007a,b), and we suggest that break-up of the Congo Craton from Rodinia occurred during Mwashya rifting (ca. 765 Ma), at a similar time to other recorded Rodinia break-up events (summarised in Pisarevsky et al., 2003). In order for the Congo-São Francisco Craton to reach equatorial latitudes and a rotational aspect as constrained by the Mbozi Complex pole at ca. 755 Ma, 10 million years after rifting (compare Fig. 4g with Fig. 4h), potentially unrealistic plate motions would be required.

8. Conclusions

Geological evidence indicates that the southern margin of the Congo Craton underwent compressional tectonics at ca. 1020 Ma, which could be used to infer that the combined Congo-São Francisco Craton collided with, and became an integral part of Rodinia. If the Congo-São Francisco Craton indeed did form part of Rodinia, then two potential positions can be constrained using available palaeomagnetic data. A careful consideration of these two positions either shows incompatibilities with geological evidence or overlap in position of other (better constrained) cratons, and we consider it unlikely that the Congo-São Francisco Craton ever formed part of Rodinia. The available palaeomagnetic data are also compatible with the scenario that the Congo-São Francisco Craton existed as an independent craton that was not associated with Rodinia at all. We propose that accretion of juvenile to mature island arcs and microcontinental fragments along the margin of the independent Congo Craton, could account for the formation of the 1020 Ma compressional Irumide Orogen. The current data do not allow us to unequivocally determine whether the Congo-São Francisco Craton formed an integral part of the Rodinia supercontinent or whether it acted as an independent craton.

More high-quality precisely dated palaeomagnetic poles from the Congo-São Francisco Craton could help resolve this uncertainty. As one can see from Table 1, it would be espe-
cially helpful to obtain new palaeomagnetic data for the time of “stable” Rodinia around 900–800 Ma.

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