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Depositional age, provenance, and tectonic and paleoclimatic settings of the late Mesoproterozoic–middle Neoproterozoic Mbuji-Mayi Supergroup, Democratic Republic of Congo

Franck Delpomdor [a,*], Ulf Linnemann [b], Ariel Boven [c], Andreas Gärtner [b], Aleksey Travin [d], Christian Blanpied [e], Aurélien Virgone [e], Hielke Jelsma [f], Alain Préat [a]

[a] Biogeochemistry & Modeling of the Earth System, Université libre de Bruxelles, 1050 Brussels, Belgium
[b] Museum für Mineralogie und Geologie, Sektion Geochronologie, GeoPlasma Lab, Senckenberg Naturhistorische Sammlungen, 01109 Dresden, Germany
[c] Section of Isotope Geology, Royal Museum for Central Africa, 3080 Tervuren, Belgium
[d] United Institute of Geology, Geophysics and Mineralogy, Siberian Branch, Novosibirsk 630090, Russia
[e] Total SA, 92078 Paris, France
[f] De Beers Exploration, Southdale 2135, Johannesburg, South Africa

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ABSTRACT

The late Mesoproterozoic–middle Neoproterozoic period (*ca.* 1300 Ma–800 Ma) heralded extraordinary climatic and biological changes related to the tectonic changes that resulted in the assembly (~1.0 Ga) and the break-up of Rodinia (880 Ma–850 Ma). In the Democratic Republic of Congo, these changes are recorded in the Mbuji-Mayi Supergroup which was deposited in the SE–NW trending siliciclastic-carbonate failed-rift Sankuru-Mbuji-Mayi-Lomami-Lovoy Basin. New LA-ICP–MS U–Pb laser ablation data on detrital zircon grains retrieved from the lower arenaceous-pelitic sequence (BI group) together with C and Sr isotopic data on carbonates from the upper dolomitic-pelitic sequence (BII group) and an $^{40}Ar/^{39}Ar$ age determination on a dolerite give a new depositional time frame between 1174 ± 22 Ma and *ca.* 800 Ma for the Mbuji-Mayi Supergroup. The upper age limit is based on the assumption that the transition between the BIIb and BIIc subgroups recorded the Bitter Springs anomaly. In terms of tectonic and paleoclimatic settings, the BII group was deposited in the eastern passive margin of the Congo Craton during warm periods interlaced with temporarily dry and wet seasons, suggesting greenhouse conditions during the fragmentation of Rodinia.

1. Introduction

Since the demise of the Rodinia Supercontinent (McMenamin and McMenamin, 1990) to the assembly of the eastern Gondwana (eastern Africa, Arabian–Nubian Shield, Seychelles, India, Madagascar, Sri Lanka, East Antarctica and Australia) in the ~750 Ma to ~530 Ma interval (Fitzsimons, 2000; Meert, 2003), it has long been argued that the extraordinary biogeochemical and climatic fluctuations are deeply intertwined with global tectonics (Kaufman et al., 1993, 1997; Halverson et al., 2007). The low-latitude distribution of continents in the late Neoproterozoic (Kirschvink, 1992; Hoffman et al., 1998; Hoffman and Schrag, 2002; Li et al., 2008) would have favored a cool climate due to the nature of continental chemical weathering, atmospheric and oceanic circulation, and the magnitude of planetary albedo (Halverson et al., 2010). These effects led to several icehouse climate periods throughout the Neoproterozoic times. These dramatic climatic changes, global carbon cycling and atmospheric oxygen budget (Knoll et al., 1986; Derry et al., 1992; Knoll, 1992; Des Marais, 1994; Strauss, 1997; Hoffman et al., 1998; Canfield, 1999) were recorded in the well-pronounced fluctuations of C, Sr and S isotopic compositions observed in the worldwide mid- (Cryogenian, i.e. 850 Ma–635 Ma) to late (Ediacaran, i.e. 635 Ma–541 Ma) Neoproterozoic series (Kaufman and Knoll, 1995; Asmerom et al., 1991; Halverson et al., 2005, 2007; Kaufman et al., 2006; Tewari and Sial, 2007).

In contrast to the late Neoproterozoic, the late Mesoproterozoic–early Neoproterozoic successions (*ca.* 1300 Ma–850 Ma) are less well isotopically constrained. This interval exhibits moderately $\delta^{13}C$ variations between −2‰ and +4‰ (Fairchild et al., 1990; Knoll et al., 1995; Podkovyrov et al., 1998; Kah et al., 1999; Bartley et al., 2001), with pronounced negative shifts of up to 5‰ in average. The second large negative $\delta^{13}C$ anomaly (Bitter Springs stage) is recorded at *ca.* 800 Ma and appears to be unrelated to glaciation, but, in fact, to be related to eustatic changes (Halverson et al., 2005). The most common explanation of such negative excursions is that they result of organic matter burial with oxidation and input of massive volumes of ^{13}C-depleted carbon from reduced reservoirs, such as sedimentary methane clathrates, labile organic matter, or dissolved organic carbon (Rothman et al., 2003). In contrast, the $\delta^{13}C$ rise reflects increased biomass, e.g. along continental

* Corresponding author at: Avenue F.D. Roosevelt, 50 CP260/2 1050 Brussels, Belgium. Tel.: +32 2 650 29 01.
E-mail address: franck.delpomdor@ulb.ac.be (F. Delpomdor).

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margins during periods of rifting, and/or proportional burial of ^{13}C-depleted organic matter, potentially attributable to one of several tectonic forcing mechanisms (Knoll, 1992; Hoffman, 1999; Bartley et al., 2001).

Secular change in Sr isotopic composition is a useful indicator of global-scale changes of seawater related to tectonics and climatic regimes throughout the Neoproterozoic times. The late Mesoproterozoic seawater rose from early Mesoproterozoic lows (~0.7040) to values as high as 0.7060–0.7065 at 1300 Ma, evolving to 0.7051–0.7055 at the Mesoproterozoic–Neoproterozoic boundary (1000 Ma; Gorokhov et al., 1995; Bartley et al., 2001), and 0.7060 for the early Neoproterozoic (Veizer et al., 1983; Asmerom et al., 1991; Melezhik et al., 2001; Halverson et al., 2005, 2007). $^{87}Sr/^{86}Sr$ ratios have risen from 0.7070 to 0.7063 with the onset of the *ca.* 800 Ma Bitter Springs anomaly to >0.7085 for the late Neoproterozoic (Burke et al., 1982; Jacobsen and Kaufman, 1999; Shields, 1999; Brasier et al., 2000; Halverson et al., 2007) and ~0.7093 (Montanez et al., 1996) for the Late Cambrian.

In this paper, we present new age constraints for the late Mesoproterozoic–middle Neoproterozoic Mbuji-Mayi Supergroup (Democratic Republic of Congo) through various methods such as LA-ICP-MS U–Pb geochronology on detrital zircon grains, C–O–Sr chemostratigraphy and $^{40}Ar/^{39}Ar$ dating. Here, we attempt to better constrain the provenance of the Mbuji-Mayi sediments and to establish correlations with other late Meso- and middle Neoproterozoic units in Central Africa, and to constrain the minimum age of the Mbuji-Mayi Supergroup.

2. Paleogeography

The span of geologic time that stretches from the late Mesoproterozoic through the early–middle Neoproterozoic (1300 Ma–800 Ma) heralded extraordinary climatic and biological changes related to the tectonic changes that resulted in the assembly and the break-up of Rodinia. Although the timing of the Rodinia assembly will continue to be debated for some time (Piper, 1976; Bond et al., 1984; McMenamin and McMenamin, 1990; Dalziel, 1991; Hoffman, 1991; Karlstrom et al., 1999), as the exact configuration of various elements surrounding Rodinia is still not clear (Dalziel, 1997; Piper, 2000; Sears and Price, 2000). The greatest certainty is that Laurentia occupied the center of a major landmass, within periphery a multitude of blocks, which were not yet coherently positioned on the margin of East Gondwana from ~1.0 Ga until all its constituents became assembled in the late Neoproterozoic and early Cambrian times (Meert, 2003).

The first Rodinia models placed the southwestern Laurentia margin adjacent to East Antarctica (SWEAT hypothesis; Dalziel, 1991, 1997; Hoffman, 1991; Moores, 1991; Weil et al., 1998). While a second model placed the southwestern Laurentia adjacent to Australia (AUSWUS, Karlstrom et al., 1999; Burret and Berry, 2000). However, these SWEAT and AUSWUS models were replaced based on new paleomagnetic data of Western Australia, by a configuration proposing the AUSMEX (Australia–Mexico) connection (Wingate et al., 2002).

The paleocontinental configuration of the Congo Craton in the assembly of Rodinia is therefore still debated. Dalziel (1997) placed the Congo Craton, associated with the Kalahari Cratons, in the western margin of Laurentia. According to paleomagnetic results, Pisarevsky et al. (2003) suggested that the Congo Craton was independent from Rodinia in the late Mesoproterozoic, and only joined along its south-eastern margin facing Laurentia at *ca.* 1000 Ma (Hoffman, 1991; Li et al., 2008). During this time-span, the Congo Craton was marked by the Kibaran (1.4–1.0 Ga) and Irumide (*ca.* 1.02 Ga) orogenic events (De Waele, 2005; Johnson et al., 2005; De Waele and Fitzsimons, 2007; Begg et al., 2009; Fernandez-Alonso et al., 2012), which result of the accretion of Congo–Tanzania–Bangweulu–Kalahari cratons (Pinna et al., 1996).

In the eastern margin of the Congo Craton, the early Neoproterozoic Era is characterized by the *ca.* 880–850 Ma continental break-up of Rodinia (Porada and Berhorst, 2000), which probably started at the newly eastern passive margin formed during the opening of the Mozambique ocean (Li et al., 2008). It gradually spread westwards into the area of the Zambezi rift-basin, and thereafter northwestwards into the area of the Katanga rift-basin (Porada and Berhorst, 2000), and finally in the Sankuru-Mbuji-Mayi-Lomami-Lovoy (SMLL) rift-basin with deposition of the Mbuji-Mayi Supergroup. In the eastern part of the Katanga rift-basin, a differential movement of Congo–Tanzania cratons caused the opening of Kundelungu rift (aulacogen), accompanied by felsic intrusions at *ca.* 765 Ma (Kafue rhyolites: 879 Ma; Nchanga Granite: 877 Ma; Lusaka Granite: 865 Ma; Armstrong et al., 2005) and the development of an extensional basin trending northeastwards along pre-existing Kibaran structures. This rifting propagation is not recorded in the SMLL Basin indicating that sedimentation related to the collision of Congo and Kalahari cratons along the Mwembeshi Shear Zone ceased with subsequent development of strike-slip systems. This induced clockwise rotation of the Lufilian Belt (Unrug, 1983) which can be correlated with a first phase of deformation in the Lufilian Belt ('Kolwezian phase'), between 790 Ma and 750 Ma, and with the collisional deformation in the Zambezi Belt at 820 Ma (Hanson et al., 1988).

The Pan-African tectonic episode was related to a *ca.* 560–550 Ma collision between the western margin of the Congo Craton and the São Francisco Craton, and between the Kasai–(Angola)–Kalahari cratons and Congo–Tanzania cratons along a southeast-northwest trending suture linking up the southern Mozambique Belt with the Araçuai–West Congo Belt (Porada and Berhorst, 2000; Pedrosa-Soares et al., 2001, 2007; Alkmim et al., 2006, 2007; Pedrosa-Soares and Alkmin, 2011). The Pan-African orogen ended with the polyphase assembly of Gondwana during the East Africa, Brasiliano, Kuungan and Damaran orogenic events that extended to at least the end of the Cambrian (~490 Ma; Meert, 2003; Gray et al., 2008; Begg et al., 2009; Fig. 1).

3. Geological setting

The SMLL Basin (Fig. 2) is an intracratonic failed-rift basin (Kadima et al., 2011) located between 6°S and 8°S latitude and 23°E and 26°E longitude, in the Kasai-Oriental Province (Democratic Republic of Congo). The 'System of Bushimay', herein renamed Mbuji-Mayi Supergroup, established by Raucq (1957) who mapped the geology of the Mbuji-Mayi area (geological map of 1/200,000, Raucq, 1958). Since 1977, petrographic and geochronological data for stratigraphic correlations in Central Africa have been acquired (Bertrand-Sarfati, 1972; Cahen et al., 1984). The stratigraphy may be summarized, following Raucq (1970), as follows.

3.1. Basement rocks

In the western part of the SMLL Basin, the basement consists of migmatitic gneisses with granitic to tonalitic compositions. Weakly deformed and undeformed granites (e.g. Malafudi granites) are present in the SE part of the Mbuji-Mayi area, and considered as the product of migmatization related to the Neoarchean Moyo deformation event (2680 Ma; Delhal, 1991). Rb–Sr ages of the Malafudi granites gave an age of 2590 (Delhal, 1991). Hf-isotope analyses indicated that this magmatic event was related to reworking of 2.8–3.0 Ga crust (Batumike et al., 2009). In northwestern Zambia, Key et al. (2001) reported zircon U–Pb SHRIMP crystallization ages of 2561 ± 10 Ma and 2538 ± 10 Ma in granitic gneisses.

In the southern part of the SMLL Basin, the Mbuji-Mayi Supergroup rests on Kibaran Belt lithologies including from base to top (i) the Kiaora (K1) or Mitwaba Group (~4300 m; Cahen, 1954; Van de Steen, 1959; Cahen et al., 1984; Kokonyangi et al., 2001); (ii) the Kataba Conglomerate (~3900–5500 m); (iii) the Nzilo Group (K2, ~400–600 m; Mortelmans, 1951; Cahen, 1954; Kokonyangi et al., 2001); (iv) the Hakansson Group (K3, ~1700 m; Cahen, 1954; Van de Steen, 1959; Cahen et al., 1984); and (v) the Lubudi Group (K4, ~1000–1300 m). The Kibaran