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Research Paper

Contemporaneous assembly of Western Gondwana and final Rodinia break-up: Implications for the supercontinent cycle

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ABSTRACT

Geological, geochronological and isotopic data are integrated in order to present a revised model for the Neoproterozoic evolution of Western Gondwana. Although the classical geodynamic scenario assumed for the period 800–700 Ma is related to Rodinia break-up and the consequent opening of major oceanic basins, a significantly different tectonic evolution can be inferred for most Western Gondwana cratons. These cratons occupied a marginal position in the southern hemisphere with respect to Rodinia and recorded subduction with back-arc extension, island arc development and limited formation of oceanic crust in internal oceans. This period was thus characterized by increased crustal growth in Western Gondwana, resulting from addition of juvenile continental crust along convergent margins. In contrast, crustal reworking and metacratonization were dominant during the subsequent assembly of Gondwana. The Río de la Plata, Congo-São Francisco, West African and Amazonian cratons collided at ca. 630–600 Ma along the West Gondwana Orogen. These events overlap in time with the onset of the opening of the Iapetus Ocean at ca. 610–600 Ma, which gave rise to the separation of Baltica, Laurentia and Amazonia and resulted from the final Rodinia break-up. The East African/Antarctic Orogen recorded the subsequent amalgamation of Western and Eastern Gondwana after ca. 580 Ma, contemporaneously with the beginning of subduction in the Terra Australis Orogen along the southern Gondwana margin. However, the Kalahari Craton was lately incorporated during the Late Ediacaran–Early Cambrian. The proposed Gondwana evolution rules out the existence of Pannotia, as the final Gondwana amalgamation postdates latest connections between Laurentia and Amazonia. Additionally, a combination of introversion and extroversion is proposed for the assembly of Gondwana. The contemporaneous record of final Rodinia break-up and Gondwana assembly has major implications for the supercontinent cycle, as supercontinent amalgamation and break-up do not necessarily represent alternating episodic processes but overlap in time.

1. Introduction

The term *Gondwana* was first used in the geological literature to refer to a plant-bearing series in India and was afterwards extended to the Gondwana system (Feistmantel, 1876; Medlicott and Blanford, 1879, and references therein). Based on similarities in the Paleozoic–Mesozoic geological and fossiliferous record of India and other continental masses, Suess (1885) proposed the existence of a supercontinent and coined the name *Gondwanaland*, which was extended to South America, Australia and Antarctica by Wegener (1915).

The amalgamation of Gondwana started at ca. 630 Ma and extended to ca. 550–530 Ma, when subduction along its proto-Pacific margin was already established (Dalziel, 1997; Cordani et al., 2003; Meert and Torsvik, 2003; Cawood, 2005; Collins and Pisarevsky, 2005; Cawood and Buchan, 2007). Likewise, Pannotia was considered as a Late Neoproterozoic “short-lived” supercontinent that included Laurentia and Gondwanan domains, prior to Gondwana final configuration (Powell and Young, 1995; Dalziel, 1997). On the other hand, the break-up of Rodinia took place in two phases at ca. 800–700 Ma and after ca. 600 Ma, being the later coeval with the timing of Gondwana assembly (Cordani et al., 2003; Cawood, 2005; Li et al., 2008). Late Neoproterozoic paleogeography

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thus resulted from major geodynamic processes that might be a priori linked to the evolution of Rodinia, Pannotia and/or Gondwana (Fig. 1).

Two end-member processes were proposed for the formation of supercontinents, namely introversion and extroversion, depending on whether internal or external oceans of a supercontinent are closed to form the next one, respectively (Fig. 2; Nance et al., 1988; Hartnady, 1991; Hoffman, 1991; Murphy and Nance, 2003, 2005, 2013; Mitchell et al., 2012; Evans et al., 2016). Internal oceans comprise juvenile crust that forms after break-up of the previous supercontinent and external ocean crust predates the previous supercontinent (Murphy and Nance, 2003, 2005). In terms of paleogeography, introversion implies that the location of a supercontinent is the same of its predecessor, whereas the successor is located in the opposite hemisphere in the case of extroversion (Mitchell et al., 2012). In the particular case of Gondwana, an extroversion model has been typically considered (Hoffman, 1991; Murphy and Nance, 2003, 2005, 2013; Evans et al., 2016).

Based on a review of geological, geochronological and isotopic evidences, a revised model for the Neoproterozoic evolution of Western Gondwana is presented in this work. Relationships with Rodinia break-up and the evolution of the Terra Australis Orogen are discussed, and implications for the supercontinent cycle are analyzed as well.

2. Pre-Gondwana configuration

Many contributions attempted to elucidate Late Mesoproterozoic–Early Neoproterozoic paleogeography (e.g., Powell et al., 1993; Dalziel et al., 2000; Kröner and Cordani, 2003; Pisarevsky et al., 2003; Tohver et al., 2006; Li et al., 2008; Evans, 2009), which represents a key point to understand the history of Gondwana amalgamation. The assembly of Rodinia took place at ca. 1.1–1.0 Ga and, although most authors agree on the fact that the Amazonian Craton was part of Rodinia (Dalziel et al., 2000; Tohver et al., 2006; Li et al., 2008; Evans, 2009), the participation of other western Gondwanan blocks is still under discussion. Kröner and Cordani (2003) and Cordani et al. (2003) indicated that the Río de la Plata, Kalahari and Congo-São Francisco cratons were not part of Rodinia, which was further supported by Tohver et al. (2006) and Rapalini et al. (2013). In contrast, Evans (2009) included all blocks within Rodinia.

In the case of the Río de la Plata Craton, the pre-Brasiliano geological record (i.e., older than ca. 650 Ma) is restricted to the Paleoproterozoic. Basement rocks comprise mainly Rhyacian–Orosirian gneisses and granitoids, which show Late Paleoproterozoic K–Ar muscovite cooling ages and are intruded by Statherian mafic dykes (Cingolani, 2011; Oyhantçabal et al., 2011). Paleoproterozoic rocks are covered by Neoproterozoic metasedimentary rocks and only show local overprinting related to the Brasiliano Orogeny (Martínez et al., 2013; Oriolo et al., 2016a), pointing to lack of Mesoproterozoic events and isolation of the Río de la Plata Craton during Rodinia evolution. In a similar way, the West African Craton is made up of Archean and Paleoproterozoic nuclei and lacks in rocks yielding ages between ca. 1.7 and 1.0 Ga (Ennih and Liégeois, 2008, and references therein). Although most paleogeographic reconstructions placed this block attached to the Amazonian Craton (e.g., Cordani et al., 2003; Tohver et al., 2006; Li et al., 2008), isolation of the West African Craton during the Mesoproterozoic seems to be a more plausible scenario (Fig. 1a).

The Congo-São Francisco Craton, in turn, shows a protracted Mesoproterozoic evolution. In Brazil, the São Francisco Craton records intraplate magmatism and sedimentation related to several extensional events throughout the Late Paleoproterozoic–Mesoproterozoic (Chemale et al., 2012; Ribeiro et al., 2013, and references therein), whereas the Congo Craton in Africa exhibits several distinct Mesoproterozoic magmatic events at ca. 1.50, 1.38 and 1.10 Ga (Ernst et al., 2013). Despite being almost coeval with the timing of Rodinia assembly, the youngest event comprises gabbro–norite dykes that resulted from intraplate magmatism (Ernst et al., 2013).

A different situation can be observed in the Kalahari Craton, which presents an Archean–Paleoproterozoic nucleus surrounded by Late Mesoproterozoic mobile belts. The northern (present coordinates) Sinclair–Ghanzi–Chobe Belt recorded arc-related magmatism resulting in collision and subsequent post-collisional magmatism at ca. 1.1 Ga (Kampunzu et al., 1998; Becker et al., 2006). To the southwest, collision-related deformation and metamorphism in the Namaqua Belt has been placed at ca. 1.20–1.18 Ga (e.g., Miller, 2008; Colliston et al., 2015; Cornell et al., 2015), although extension in a back-arc setting was alternatively considered for this major tectonothermal event (Bial et al., 2015a, b). The southeastern margin of the Kalahari Craton, in turn, is bounded by the Natal Belt, which recorded accretion of island arc complexes (e.g., Thomas, 1989; Jacobs and Thomas, 1994; Spencer et al., 2015). On the other hand, the Umkondo LIP intruded the Kalahari Craton at ca. 1.1 Ga (Hanson et al., 2004) and was correlated with coeval intraplate extensional magmatism in the Congo Craton (Ernst et al., 2013). In contrast, Becker et al. (2006) interpreted this intrusion as the result of post-collisional processes.

Although it seems to be clear that the Río de la Plata, West African and Congo-São Francisco cratons were not part of Rodinia (Fig. 1a), the position of the Kalahari Craton is still uncertain. However, even if being part of Rodinia, the Kalahari Craton might already rift away at ca. 700 Ma (Jacobs et al., 2008). Rifting at ca. 800–700 Ma gave rise to the opening of a major oceanic basin between Laurentia and eastern Gondwanan cratons and was succeeded by a second rifting event starting after ca. 610–600 Ma (Fig. 1a) that triggered the opening of the Iapetus Ocean between Laurentia, Baltica and Amazonia (e.g., Cawood et al., 2001; Meert and Torsvik, 2003; Li et al., 2008). In any case, most reconstructions do not consider Mesoproterozoic connections of the Congo-São Francisco and Kalahari cratons (e.g., Cordani et al., 2003; Meert and Torsvik, 2003; Tohver et al., 2006), which is further supported by detrital zircon data from the Damara Belt indicating different provenance for passive margins of both blocks till the Early Ediacaran (Foster et al., 2015). Hence, the Río de la Plata, West African, Congo-São Francisco and Kalahari cratons did not interact with each other during the Mesoproterozoic and probably occupied a marginal position with respect to Rodinia during the Early Neoproterozoic (Fig. 1a).

3. The assembly of Gondwana

Comparison of available data allows characterizing the amalgamation of the Río de la Plata and Congo cratons as one of the earliest collisional events, which is constrained at ca. 630 Ma by $^{40}Ar/^{39}Ar$ and K–Ar amphibole and mica data from the Dom Feliciano Belt in Uruguay (Figs. 3 and 4a; Oriolo et al., 2016b). This collisional event implied the docking of the Nico Pérez Terrane to the Río de la Plata Craton margin along the Sarandí del Yí Shear Zone (Oriolo et al., 2015, 2016a). Consequent regional metamorphism, crustal shortening and exhumation is recorded along the belt between ca. 630 and 600 Ma (da Silva et al., 1999; Chemale et al., 2011; Oriolo et al., 2016b; Philipp et al., 2016). Peraluminous syncollisional leucogranites resulting from crustal anatexis at 740–820 °C and 8–9 kbar were succeeded by voluminous post-collisional magmatism between ca. 630–580 Ma (Oyhantçabal et al., 2007; Florisbal et al., 2009, 2012; Basei et al., 2011; Philipp et al., 2013, 2016).

Further north, the Ribeira Belt records a protracted history of minor terrane amalgamation (Figs. 3 and 4b; Heilbron et al., 2008).

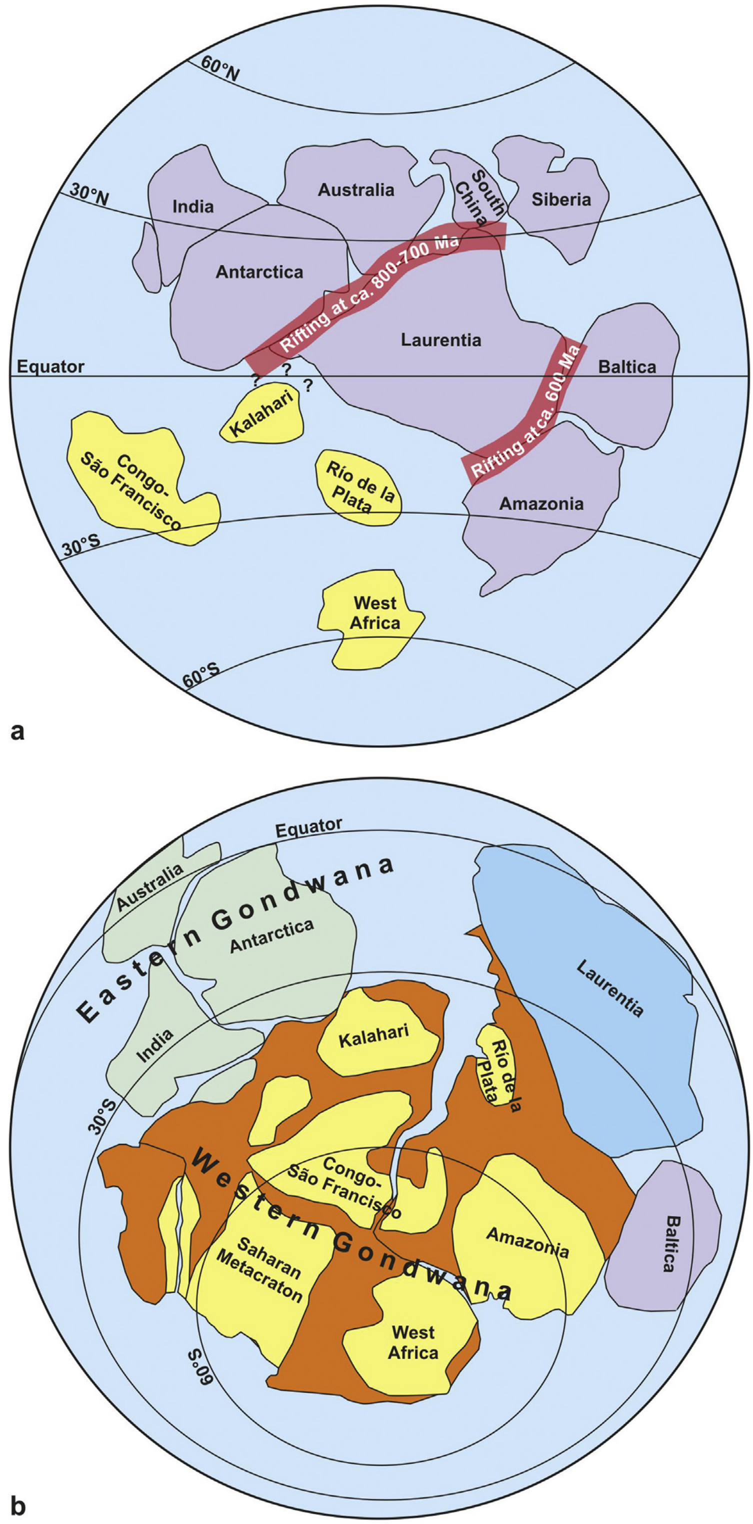

Figure 1. (a) Rodinia reconstruction (modified after Meert and Torsvik, 2003). “Non-Rodinian” blocks are shown in yellow. Question marks indicate uncertain position of the Kalahari Craton. Note that location of the Río de la Plata and West African cratons are tentative due to the lack of Stenian–Tonian rocks. (b) Pannotia reconstruction for the Late Ediacaran–Cambrian (modified after Dalziel, 1997). Western and Eastern Gondwana domains are shown.

A U–Pb SHRIMP zircon age of 612 ± 13 Ma constrains a metamorphic event related to an Early Ediacaran collisional event (Bento dos Santos et al., 2010), although peak metamorphic conditions of ca. 700–800 °C and 9.5–12 kbar were attained after ca. 590 Ma (Bento dos Santos et al., 2010; Faleiros et al., 2011). Subsequent terrane accretion took place at ca. 580–550 and 525–520 Ma as well (Schmitt et al., 2004; Heilbron et al., 2008; Fernandes et al., 2015), with peak metamorphic conditions of >780 °C and >9 kbar for the latter (Schmitt et al., 2004). Nevertheless, Meira et al. (2015) argued for a model of intracontinental deformation throughout the Ribeira Belt instead of multiple collisional events.

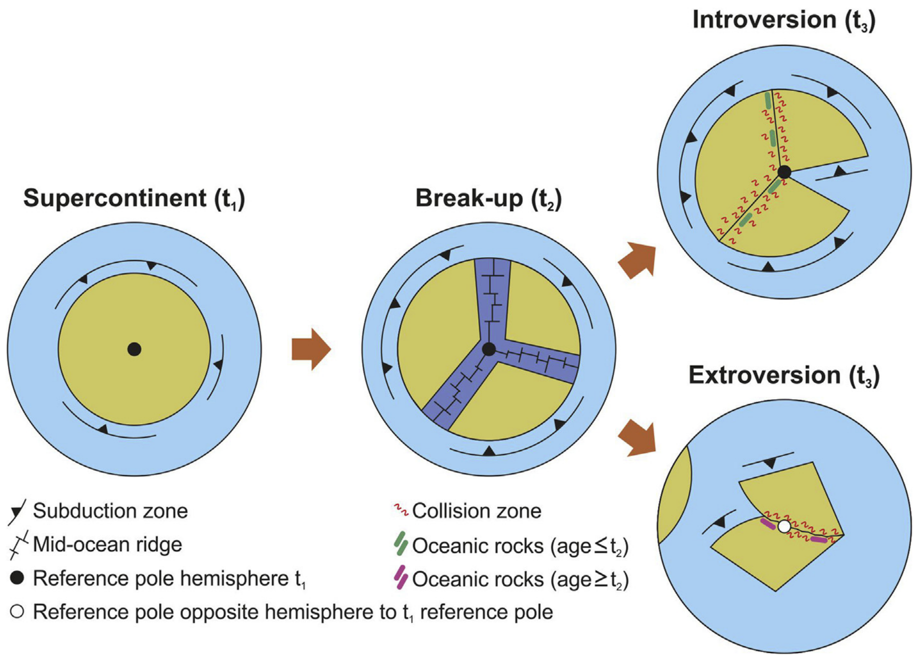

Figure 2. Cartoons illustrating two end-member processes for the formation of supercontinents (modified after Murphy and Nance, 2003).

The Brasilia Belt, in turn, represents the western and southern margin of the São Francisco Craton. Peak metamorphic conditions of ca. 825 °C and 12 kbar were achieved at 617.7 ± 1.3 Ma (ID-TIMS monazite) along the southern branch of this belt, thus constraining the collision of the São Francisco Craton and the Paranapanema Plate (Figs. 3 and 4b; Campos Neto et al., 2010).

Contemporaneous collisional processes are recorded along the Transbrasiliano Lineament (Figs. 3 and 4b), which separates the Amazonian Craton from the Borborema Province and the São Francisco Craton. Eclogite remnants reveal UHP metamorphism and associated crustal anatexis at ca. 625–615 Ma with peak metamorphic conditions of ca. 770 °C and 17 kbar resulting from continental collision (dos Santos et al., 2009; Ganade de Araujo et al., 2014a). Dextral shearing along the Transbrasiliano Lineament and post-collisional magmatism were afterwards recorded (Ganade de Araujo et al., 2014b).

Likewise, the Kandi Lineament separates the Saharan Metacraton and the West African Craton and represents the African counterpart of the Transbrasiliano Lineament. Eclogites located to the west of the Kandi Lineament indicate peak metamorphic conditions of ca. 700–800 °C and 30–33 kbar at ca. 617–605 Ma in the Gourma region (Ganade de Araujo et al., 2014c), whereas peak metamorphic conditions of ca. 650–750 °C and 28–30 kbar at ca. 615–603 Ma were reported for eclogites of the Dahomey Belt (Figs. 3 and 4b; Ganade et al., 2016). On the other hand, eclogitic relics exposed in the Tuareg Shield to the east of the Kandi Lineament record UHP metamorphism as well and yield peak metamorphic conditions of ca. 650 °C and 20–22 kbar achieved at ca. 625–620 Ma (Figs. 3 and 4b; Berger et al., 2014).

Despite being slightly diachronous, all collisional orogens between 630 and 600 Ma present some similarities (Fig. 3). Crustal-scale dextral shear zones, i.e. the Transbrasiliano-Kandi Lineament and the Sarandí del Yí Shear Zone, separate cratonic areas to the west from metacratonic areas to the east (Fig. 5; Section 4). Likewise, Cryogenian magmatism and subsequent arc/back-arc magmatism with subduction to the east after ca. 660 Ma are recorded up to the collisional phase (Section 4; Goscombe and Gray, 2007, 2008; Rapela et al., 2011; Berger et al., 2014; Ganade de Araujo et al., 2014a; Konopásek et al., 2014; Ganade et al., 2016; Oriolo et al., 2016a). The subsequent collision gave rise to the birth of Western Gondwana already at ca. 600 Ma and the consequent amalgamation of African-derived crustal blocks to the South American Archean–Proterozoic nuclei (Fig. 4b; Rapela et al., 2011; Oriolo et al., 2016a, c). The West Gondwana Orogen, which was previously defined for northwestern Africa and central Brazil (Ganade de Araujo et al., 2014b), can be thus extended to the Uruguayan sector (Fig. 5). On the other hand, the amalgamation of Western Gondwana was contemporaneous with the last events of Rodinia break-up, which were related to the opening of the Iapetus Ocean at ca. 610–600 Ma (Figs. 3 and 4c; Cawood et al., 2001; Hartz and Torsvik, 2002; Li et al., 2008; O'Brien and van der Pluijm, 2012).

Although most Western Gondwana cratons were already amalgamated at ca. 600 Ma, the Kalahari Craton was lately incorporated (Fig. 4d and e). An early metamorphic event related to convergence along the Damara Belt was recorded at ca. 600–590 Ma by $^{40}Ar/^{39}Ar$ phengite data (Lehmann et al., 2016). Convergence also triggered sinistral shearing in the Dom Feliciano Belt (Oriolo et al., 2016a, b) and culminated with collision of the Congo and Kalahari cratons at ca. 530–520 Ma (Fig. 3; Gray et al., 2006; Schmitt et al., 2012). Peak metamorphic conditions of ca. 750–700 °C and 5 kbar were attained at ca. 525–505 Ma, which are constrained by U–Pb monazite and Sm–Nd garnet isochrone data (Jung and Mezger, 2003). Syncollisional intrusions are recorded at ca. 530 Ma (Schmitt et al., 2012), whereas regional exhumation and cooling below muscovite closure temperature are recorded up to ca. 460 Ma (Gray et al., 2006). Nevertheless, $^{40}Ar/^{39}Ar$ hornblende data from the Gariep Belt constrain an early collisional event along the western Kalahari Craton margin (Frimmel and Frank, 1998), giving rise to the closure of the Gariep-Rocha basin and deformation of post-collisional basins of the Dom Feliciano Belt at ca. 550–530 Ma (Frimmel and Frank, 1998; Basei et al., 2000, 2005; Frimmel et al., 2011; Oriolo et al., 2016b).

On the other hand, collisional processes at 580–550 Ma were reported in the East African–Antartic Orogen (Figs. 3 and 4d; Jacobs

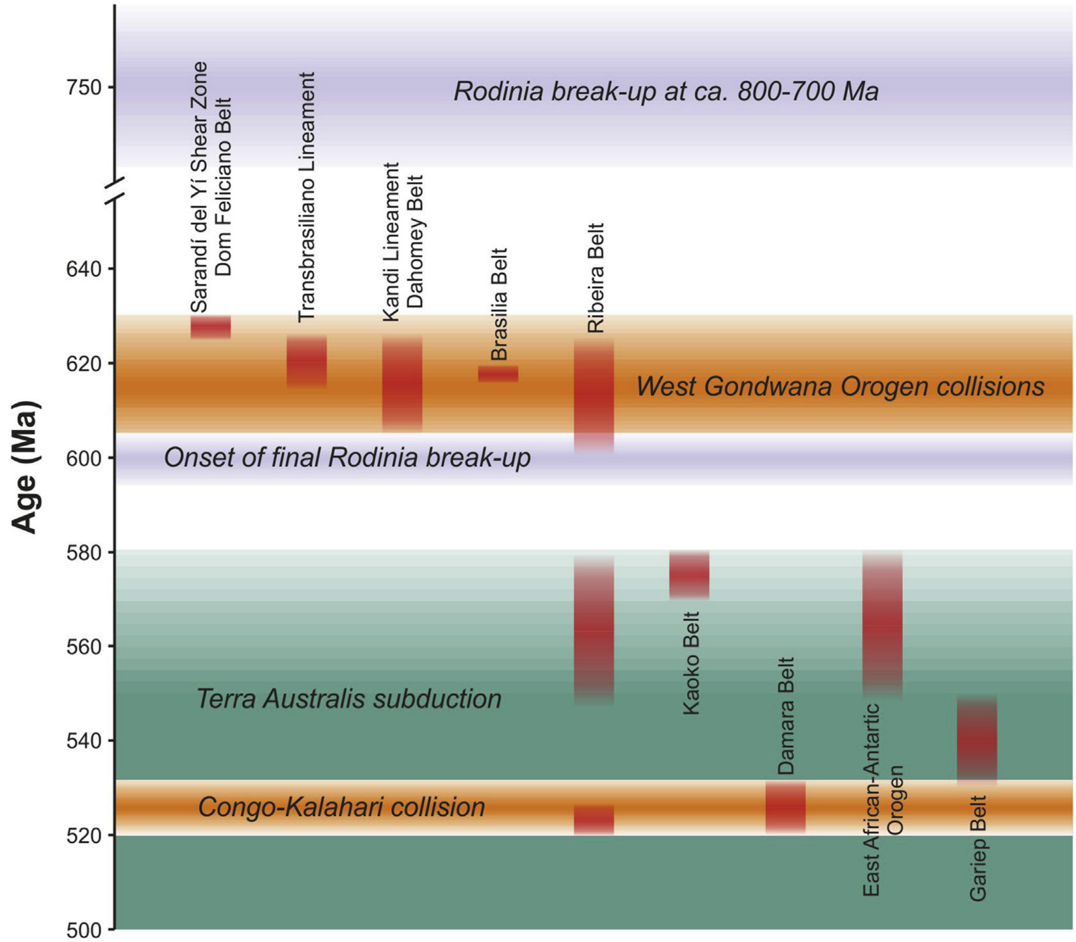

Figure 3. Timing of Gondwana-related collisional events (red) summarized from available data. Sarandí del Yí Shear Zone/Dom Feliciano Belt (Oriolo et al., 2016a,b), Transbrasiliano Lineament (Ganade de Araujo et al., 2014a,b), Kandi Lineament/Dahomey Belt (Berger et al., 2014; Ganade de Araujo et al., 2014c; Ganade et al., 2016), Brasilia Belt (Campos Neto et al., 2010), Ribeira Belt (Schmitt et al., 2004; Heilbron et al., 2008; Bento dos Santos et al., 2010), Kaoko Belt (Goscombe et al., 2005; Goscombe and Gray, 2007, 2008; Foster et al., 2009), Damara Belt (Gray et al., 2006; Schmitt et al., 2012), Gariep Belt (Frimmel and Frank, 1998; Frimmel et al., 2011), East African–Antartic Orogen (Jacobs and Thomas, 2004; Viola et al., 2008). Rodinia break-up after Meer and Torsvik (2003) and Li et al. (2008). Subduction along the Terra Australis Orogen after Cawood (2005) and Cawood and Buchan (2007). Note overlapping of the second stage of Rodinia break-up (i.e., opening of the Iapetus Ocean) and the assembly of Western Gondwana. See text for further explanation.

and Thomas, 2004; Viola et al., 2008; Fritz et al., 2013), indicating that the assembly of Western and Eastern Gondwana predates the final incorporation of the Kalahari Craton into the former. On the southern Gondwana margin, subduction is also recorded along the Terra Australis Orogen after ca. 580 Ma (Figs. 3 and 4d; Cawood, 2005; Cawood and Buchan, 2007). Along this margin, several peri-Laurentian blocks were rifted away during the opening of the Iapetus Ocean and subsequently juxtaposed to the Gondwana margin during the Paleozoic, as in the case of the Pampean Belt (Fig. 4e; Dalla Salda et al., 1992; Dalziel et al., 1994; Rapela et al., 1998, 2007, 2016; Siegesmund et al., 2010).

Hence, Late Neoproterozoic–Cambrian collisions leading to the final Gondwana assembly and coeval subduction along the Terra Australis Orogen suggest a coupling between internal collisional and marginal subduction processes, as indicated by Cawood and Buchan (2007). Likewise, these processes seemed to be strongly linked to the final stages of Rodinia break-up as well, particularly to the opening of the Iapetus Ocean. The evolution of Gondwana emphasizes the need to reevaluate the classical concept of the supercontinent cycle (Nance et al., 2014, and references therein), as supercontinent assembly and break-up do not necessarily represent alternating episodic processes but may overlap in time (Condie and Aster, 2013). On the other hand, the proposed Gondwana evolution (Fig. 4) rules out the existence of the supercontinent Pannotia (Fig. 1b; Powell and Young, 1995; Dalziel, 1997), as Laurentia, western and eastern Gondwana were not part of a single supercontinent during the Late Neoproterozoic. Indeed, Laurentia and Amazonia connections just represent remnants of the Rodinia assembly (Fig. 1a).

4. Gondwana crustal growth: from Tonian–Cryogenian island arc accretion to Ediacaran collision and metacratonization

Hf isotopic data from different Brasiliano–Pan-African belts were compiled in order to analyze the crustal growth history of Gondwana during the Neoproterozoic (Fig. 6a). In the last decade, most contributions that evaluate the crustal growth of Gondwana using isotopic data consider global databases. Though extremely useful, global databases may led to misinterpretations if the geological and tectonic framework is not clearly understood. For instance, basement inliers of the Andean chain present a Laurentian

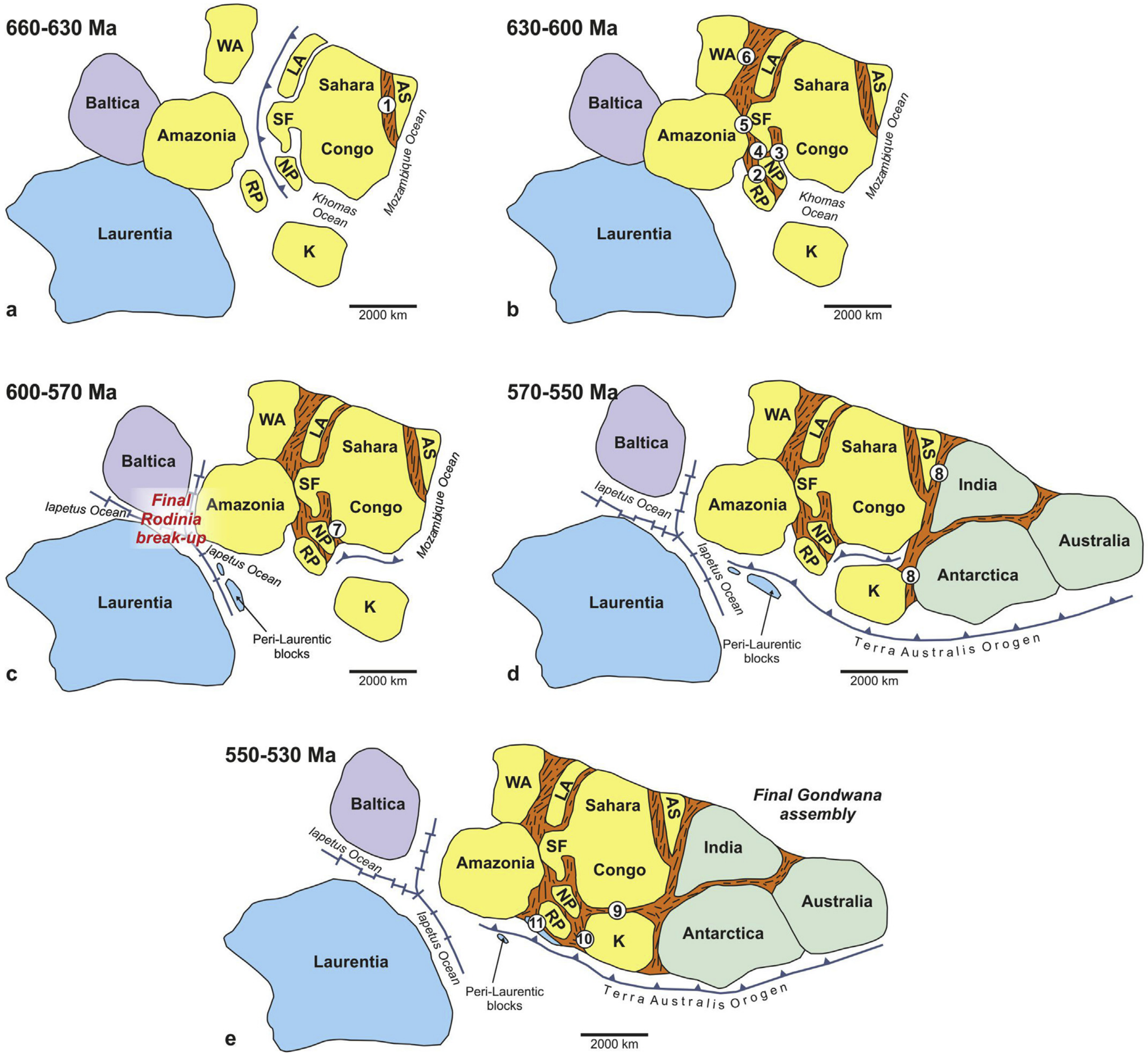

Figure 4. Schematic tectonic and paleogeographic evolution of main crustal blocks (WA: West Africa, RP: Río de la Plata, K: Kalahari, NP: Nico Pérez, SF: São Francisco, LA: LATEA, AS: Arabian–Nubian Shield). Gondwana reconstructions modified after Meert and Lieberman (2004), Collins and Pisarevsky (2005), Tohver et al. (2006), Li et al. (2008), Pisarevsky et al. (2008), Pradhan et al. (2009) and Johansson (2014). Evolution of Laurentia, Baltica and Iapetus Ocean after Cawood et al. (2001), Hartz and Torsvik (2002) and O'Brien and van der Pluijm (2012). Terra Australis Orogen after Cawood (2005) and Cawood and Buchan (2007). ① northern East African Orogen, ② Dom Feliciano Belt, ③ Ribeira Belt, ④ Brasilia Belt, ⑤ Transbrasiliano Lineament, ⑥ Kandi Lineament/Dahomey Belt, ⑦ Kaoko Belt, ⑧ East African–Antartic Orogen, ⑨ Damara Belt, ⑩ Gariep and Saldania belts, ⑪ Pampean Belt. As the onset of the final break-up of Rodinia (ca. 600 Ma) postdates the final assembly of Gondwana at ca. 550–530 Ma, the existence of Pannotia is ruled out. Late collisions in the Ribeira Belt (e.g., Schmitt et al., 2004; Heilbron et al., 2008) are not shown. See text for further explanation.

affinity and were juxtaposed to the Gondwana margin during the Paleozoic (e.g., Brito Neves and Fuck, 2014; Rapela et al., 2016). Hence, Ediacaran zircons from these areas do not record the assembly of Gondwana sensu strictu but a coeval process in Laurentia (e.g., opening of the Iapetus Ocean; Rapela et al., 2016). For this reason, the database of Fig. 6a comprises only isotopic data from Neoproterozoic zircons of Brasiliano–Pan-African belts.

Data reveal a fanning isotopic array (Fig. 6a) that points to increased continental loss towards the timing of Gondwana assembly, thus indicating dominance of crustal recycling processes as expected for collisional orogenies (Collins et al., 2011; Roberts, 2012). The isotopic array of Phanerozoic collisional orogens results from a negative correlation of ε_{Hf} vs. age, which represents increased continental loss, and a positive excursion arising from arc magmatism (Collins et al., 2011). While two main trends are also recognizable in the Gondwana array, both show a negative correlation of ε_{Hf} vs. age (Fig. 6a). Hence, addition of juvenile material was clearly subordinated to crustal reworking during the assembly of Gondwana, as further supported by dominant Archean to Mesoproterozoic Hf T_{DM} model ages of zircons yielding Brasiliano–Pan-African U–Pb ages

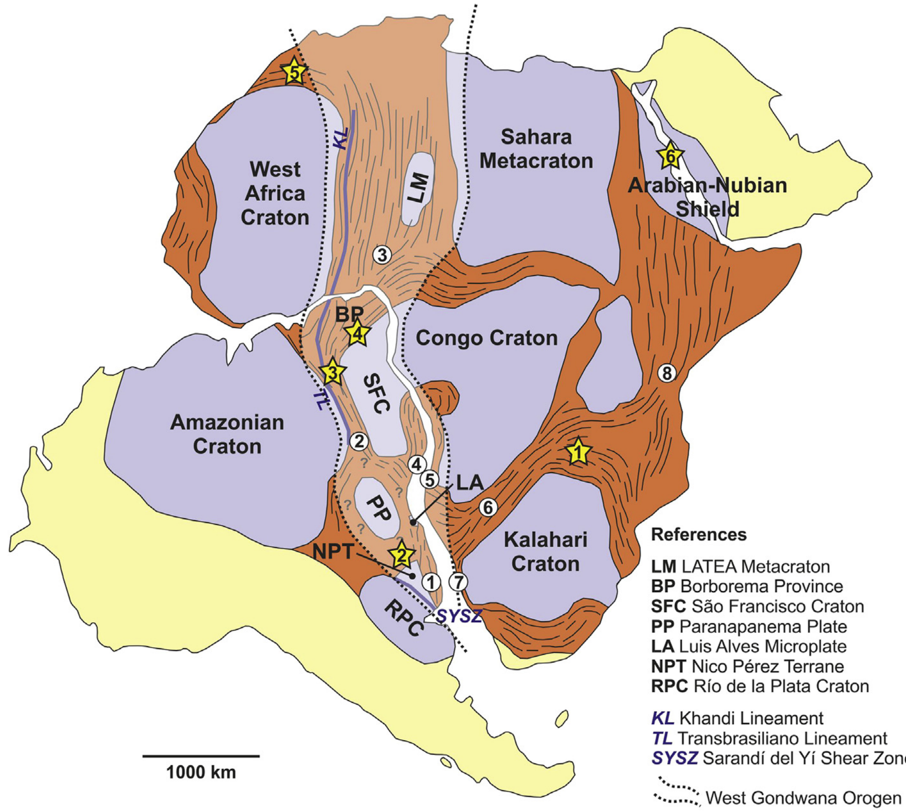

Figure 5. Main crustal blocks and Neoproterozoic orogenic belts in South America and Africa, including location of the West Gondwana Orogen (modified after Gray et al., 2008; Liégeois et al., 2013; Brito Neves and Fuck, 2014; Ganade et al., 2016). Brasiliano–Pan-African belts are shown with white dots (① Dom Feliciano Belt, ② Brasilia Belt, ③ Dahomey Belt, ④ Ribeira Belt, ⑤ Kaoko Belt, ⑥ Damara Belt, ⑦ Gariep Belt, ⑧ East African–Antarctic Orogen), whereas yellow stars indicate location of ophiolites and/or island arc complexes (1: Zambezi Belt, 2: São Gabriel Block, 3: Araguaia Belt, 4: Borborema Province, 5: Anti-Atlas Belt, 6: Arabian–Nubian Shield). Note that most crustal fragments within the West Gondwana Orogen were affected by metacratonization. See text for further explanation.

(<650 Ma; Fig. 7). Differences in the slope can be explained considering the tectonic evolution of the different orogenic belts and the age of the reworked crust. Although Lu–Hf T_{DM} model ages do not necessarily reflect crustal growth events (e.g., Roberts and Spencer, 2015; Payne et al., 2016), they show marked differences in the crustal evolution of different Western Gondwana domains (Fig. 7).

In the first place, the isotopic excursion towards negative ε_{Hf} values (i.e., evolved Hf signature) is essentially defined by data from belts related to the West Gondwana Orogen (Fig. 6a), which implies dominant reworking of Archean–Paleoproterozoic continental crust. Though present, basement remnants are significantly overprinted by deformation, magmatism and metamorphism during the Brasiliano–Pan-African Orogeny, thus indicating the importance of metacratonization processes during Gondwana assembly (Figs. 4b and 5; Liégois et al., 2013). A similar trend is also evident for the Damara Belt, resulting from recycling of Paleoproterozoic crust (Fig. 6a; Milani et al., 2015).

In the Dom Feliciano Belt, post-collisional Ediacaran magmatism records recycling of Archean-Paleoproterozoic basement rocks of the Nico Pérez Terrane and other minor crustal blocks, as indicated by inherited zircons yielding Archean and Paleoproterozoic crystallization ages and coeval whole-rock Sm–Nd and zircon Lu–Hf model ages (Oyhantçabal et al., 2007, 2009, 2012; Florisbal et al., 2012; Basei et al., 2013; Lara et al., 2016; Oriolo et al., 2016c). Th–U–Pb monazite, $^{40}Ar/^{39}Ar$ and K–Ar hornblende and mica ages also record shearing and metamorphism of the Nico Pérez Terrane basement at ca. 630–580 Ma (Oyhantçabal et al., 2009, 2011, 2012; Oriolo et al., 2016b). In a similar way, Archean–Paleoproterozoic gneisses of the São Francisco Craton show metamorphic overprinting at ca. 630–500 Ma (da Silva et al., 2005, and references therein). Coeval deformation and metamorphism of Mesoproterozoic sedimentary sequences (Süssenberger et al., 2014) as well as shear zone activity and associated hydrothermal alteration are also reported (Teixeira et al., 2010). Further north (present coordinates), Paleoproterozoic gneisses of the Borborema Province show intense overprinting due to Brasiliano magmatism and metamorphism (Neves, 2003, 2015; Ganade de Araujo et al., 2014b), which is further supported by whole-rock Sm–Nd and zircon Lu–Hf model ages of Neoproterozoic intrusions and migmatites (van Schmuss et al., 2011; Ganade de Araujo et al., 2014a). If compared with the paradigmatic Saharan and LATEA metacratons (Abdesalam et al., 2002; Liégeois et al., 2003, 2013), it is thus clear that South American counterparts show a very similar scenario of metacratonization during the Brasiliano–Pan-African Orogeny (Fig. 5).

On the other hand, the second isotopic excursion towards less positive ε_{Hf} values is defined by zircons from the Arabian–Nubian Shield (Fig. 6a). The Arabian–Nubian Shield comprises dominantly juvenile Late Tonian–Cryogenian continental crust (ca. 880–700 Ma), which is well-recorded by juvenile Hf signatures ($\varepsilon_{Hf} > +6$; Fig. 6a) and resulted from accretion of island arcs (e.g., Liégeois and Stern, 2010; Stern et al., 2010; Morag et al., 2011; Fritz et al., 2013). Zircon xenocrysts and Lu–Hf data from Neoproterozoic zircons indicate limited contribution of pre-Neoproterozoic continental crust as well, which might result from the incorporation of subducted sediments derived from a proximal continental source (Stern et al., 2010; Morag et al., 2011; Ali et al., 2013). Alternatively,

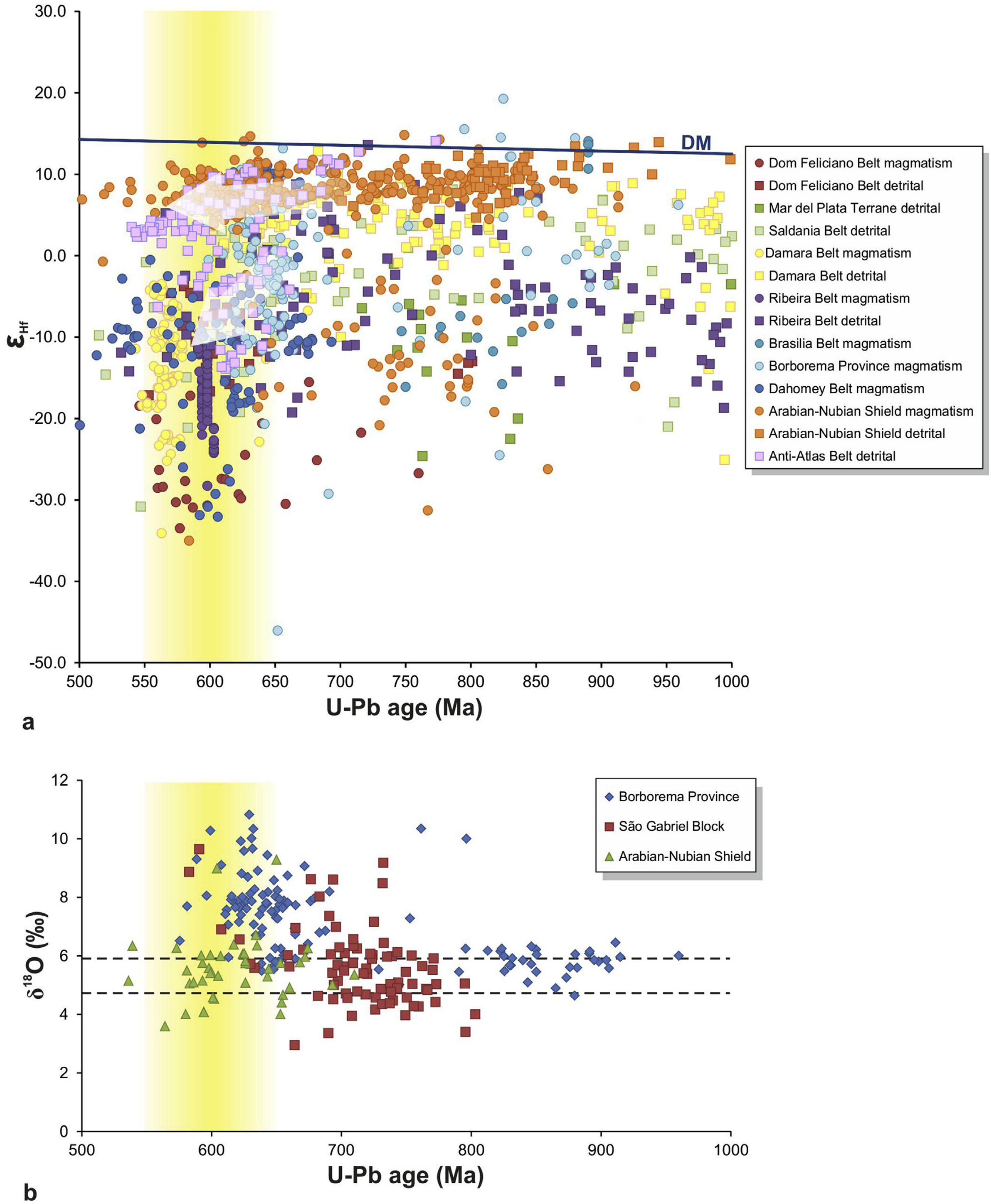

Figure 6. Synthesis of isotopic data from Neoproterozoic zircons of Brasiliano–Pan-African belts (analytical data in Appendix 1). Arrows indicate trends of increasing continental crust reworking. (a) ε_{Hf} vs. U–Pb zircon data (n = 1495) recalculated after Be'eri-Shlevin et al. (2010), Matteini et al. (2010), Morag et al. (2011), Rapela et al. (2011), Abati et al. (2012), Ali et al. (2012, 2013, 2016), Frimmel et al. (2013), Ganade de Araujo et al. (2014a), Fernandes et al. (2015), Foster et al. (2015), Milani et al. (2015), Pertille et al. (2015), Ganade et al. (2016), Janasi et al. (2016) and Oriolo et al. (2016c). The timing of Gondwana amalgamation is indicated in yellow. Data were recalculated considering a constant decay $\lambda\ ^{176}Lu = 1.867 \times 10^{-11}$ year^{-1} (Söderlund et al., 2004) and CHUR values of $^{176}Hf/^{177}Hf = 0.282772$ and $^{176}Lu/^{177}Hf = 0.0332$ (Blichert-Toft and Albarède, 1997). (b) U-Pb vs. $\delta^{18}O$ zircon data ($n = 241$) after Ganade de Araujo et al. (2014a), Fortes de Lena et al. (2014) and Ali et al. (2016). The timing of Gondwana amalgamation is indicated in yellow and mantle values are shown between dashed lines.

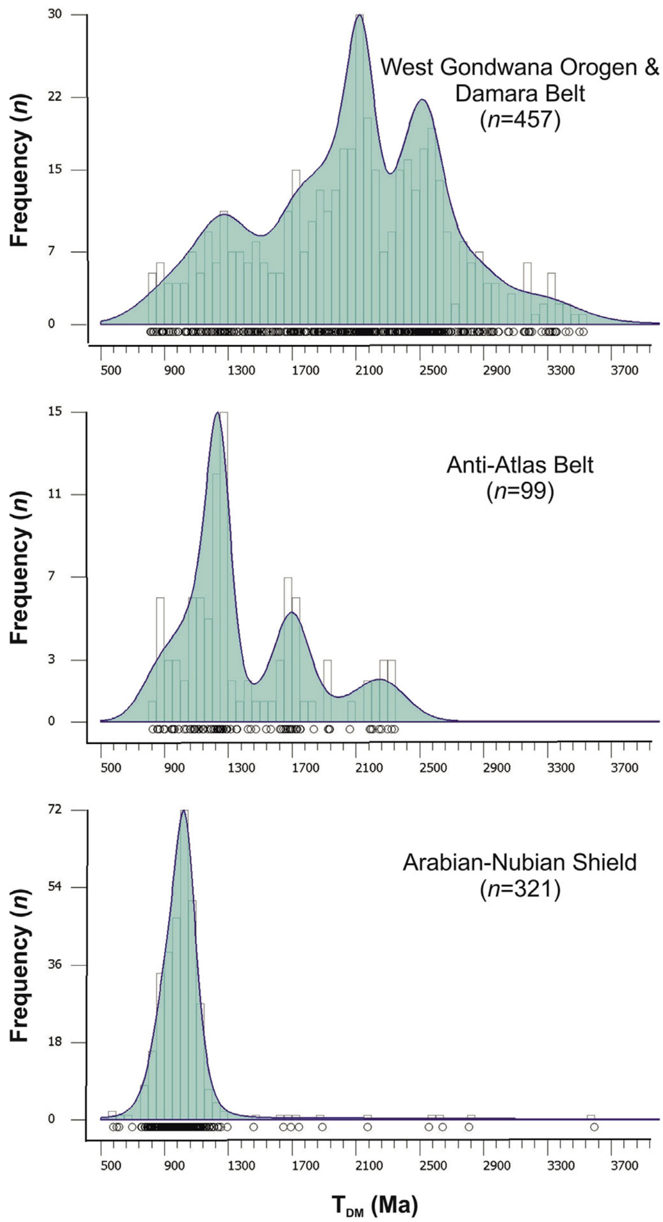

Figure 7. Kernel density estimation curve and histogram of two-stage Hf model ages plotted using Density Plotter (Vermeesch, 2012). Only zircons from Brasiliano–Pan-African belts yielding U–Pb crystallization ages younger than 650 Ma are included. Data source as for Fig. 6 (analytical data in Appendix 1). Bin width: 50 Ma. Data were recalculated considering a constant decay λ ^{176}Lu = 1.867 × 10^{-11} year^{-1} (Söderlund et al., 2004), CHUR values of ^{176}Hf/^{177}Hf = 0.282772 and ^{176}Lu/^{177}Hf = 0.0332 (Blichert-Toft and Albarède, 1997), depleted mantle (DM) values of ^{176}Hf/^{177}Hf = 0.283225 and ^{176}Lu/^{177}Hf = 0.038512 (Vervoort and Blichert-Toft, 1999) and ^{176}Lu/^{177}Hf = 0.015 for bulk Earth (Goodge and Vervoort, 2006).

Be'eri-Shlevin et al. (2010) argued for a major crustal growth event at ca. 1.2–1.1 Ga. On the other hand, whole-rock Sm–Nd and zircon Lu–Hf data from post-collisional magmatism recorded after ca. 650 Ma indicate reworking of older Neoproterozoic crust with addition of juvenile material to some extent (Be'eri-Shlevin et al., 2010; Liégeois and Stern, 2010; Morag et al., 2011; Ali et al., 2012, 2013, 2016). δ^{18}O combined with Lu–Hf data also point to mixing of juvenile Ediacaran crust and slightly older Neoproterozoic supracrustal material (Fig. 6b; Ali et al., 2016).

In the case of the Anti-Atlas Belt, zircon Lu–Hf and U–Pb data reveals a bimodal distribution that fits both trends recognized for the West Gondwana Orogen and the Arabian–Nubian Shield (Figs. 6a and 7). The excursion towards negative ε_{Hf} values is further supported by detrital zircons from Late Neoproterozoic sequences that indicate contributions from Archean and Paleoproterozoic crustal blocks and Paleoproterozoic Lu–Hf model ages (Fig. 7; Abati et al., 2012). The second trend, in turn, might result from the evolution of an intraoceanic arc between ca. 760 and 700 Ma (e.g., Bousquet et al., 2008; Triantafyllou et al., 2016), being thus comparable with the Arabian–Nubian Shield. Juvenile crust contribution was also reported for Pan-African detrital zircons and was interpreted as the result of arc magmatism along the northern Gondwana margin during the Late Neoproterozoic–Early Paleozoic (Abati et al., 2012, and references therein). Hence, this subduction-related magmatism might be a possible explanation for the juvenile Pan-African crust of the Arabian–Nubian Shield.

Despite being evident for the northern African margin, addition of Tonian–Cryogenian juvenile continental crust along the West Gondwana Orogen is not clearly reflected by data (Fig. 6a). Nevertheless, subduction with associated back-arc extension and development of island arc complexes at ca. 850–750 Ma accounts for addition of juvenile material in the Dahomey Belt (Ganade et al., 2016). For the same period, a similar setting was indicated for the Borborema Province (Ganade de Araujo et al., 2014a) and intraoceanic subduction was reported in the São Gabriel Block (Fig. 5, Fortes de Lena et al., 2014, and references therein). δ^{18}O isotopic data from these regions show mantle-like values for zircons yielding U–Pb ages of ca. 900–700 Ma and δ^{18}O > 6 for younger zircons (Fig. 6b; Fortes de Lena et al., 2014; Ganade de Araujo et al., 2014a), also indicating juvenile crust addition and subsequent reworking of supracrustal material during the assembly of Gondwana, respectively. Tonian–Cryogenian island arc development and subduction with back-arc extension recorded in several Western Gondwana regions (Fig. 8) thus represented a period of relative significant crustal growth for Western Gondwana and show a major contrast with coeval rifting and subsequent development of major oceanic basins between Rodinian blocks.

When compared with other supercontinents, the Gondwana assembly shows the most evolved Hf fingerprint, thus implying reworking of a great amount of old crustal material (Spencer et al., 2013; Gardiner et al., 2016). This has been attributed to the presence of single-sided subduction zones (Spencer et al., 2013) or, alternatively, to enhanced subduction-erosion due to steeper subduction angles (Gardiner et al., 2016). However, most of the Brasiliano–Pan-African magmatism was related to metacratonization processes, i.e., collisional to post-collisional magmatism. Metacratonization results from the lack of a thick lithospheric mantle, which leads to craton remobilization during an orogenic event, and is more likely to occur along the former active margin due to subduction (Abdesalam et al., 2002; Liégeois et al., 2013). As previously described, most Gondwanan metacratonized areas comprised the pre-collisional active margin and recorded back-arc extension as well (Fig. 8; Fritz et al., 2013; Ganade de Araujo et al., 2014c; Oriolo et al., 2016a). Likewise, back-arc development further promotes the removal of lithospheric mantle as a result of asthenospheric upwelling and, consequently, back arc zones are favorable zones for strain localization during subsequent continental collision (Hyndman et al., 2005). Though single-sided subduction (Spencer et al., 2013) and enhanced subduction-erosion (Gardiner et al., 2016)

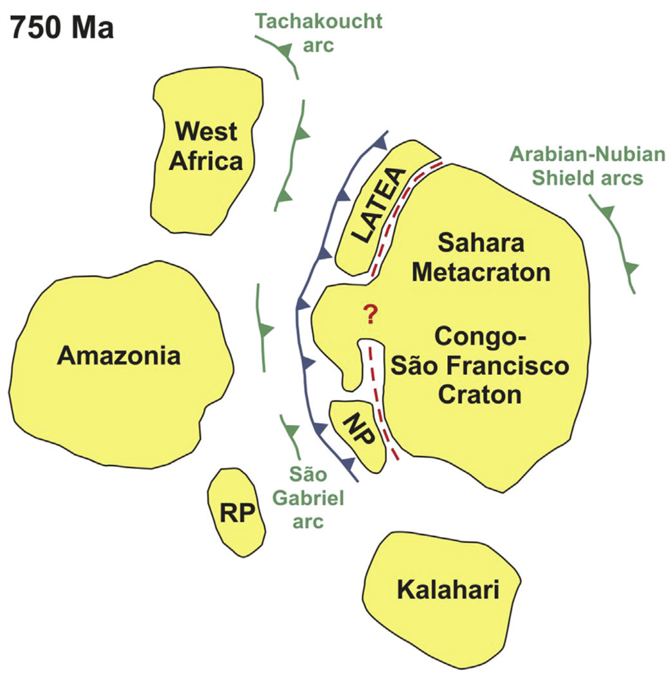

Figure 8. Sketch showing the geodynamic scenario for Western Gondwana blocks at ca. 750 Ma (modified after Fritz et al., 2013; Fortes de Lena et al., 2014; Ganade et al., 2016; Triantafyllou et al., 2016). Subduction (blue) with associated back-arc extension (red line), development of island arcs (green) and subordinated oceanic crust generation in internal oceans dominated in most Western Gondwana domains, thus contrasting with coeval rifting and major oceanic basin development recorded by Rodinian blocks (not shown). RP: Río de la Plata Craton, NP: Nico Pérez Terrane.

might however contribute, the Gondwana assembly Hf fingerprint is thus most likely associated with metacratonization processes, which in turn were influenced by the pre-collisional configuration of active margins and back-arc zones.

5. Implications for the supercontinent cycle

Together with the development of intraoceanic arcs (Section 4), ophiolite remnants between Western Gondwanan blocks record post-Rodinia oceanic crust formation during the Tonian–Cryogenian (Fig. 5). Nevertheless, relicts of older oceanic crust are present as well, such as the ca. 1.4 Ga Chewore ophiolite of the Zambezi Belt (Oliver et al., 1998).

In the São Gabriel Block of southeastern Brazil, Tonian–Cryogenian oceanic crust is recorded by ophiolitic sequences that comprise metabasalts, amphibolites, magnesian schists, serpentinites, harzburgites and albitites (Hartmann and Chemale, 2003; Arena et al., 2016). Zircons yield U–Pb SHRIMP concordant ages of 923 ± 3 and 829.4 ± 2.8 Ma for albitites of the Cerro Mantiqueiras and Ibaré ophiolites, respectively, thus constraining the timing of the magmatism (Arena et al., 2016). Likewise, zircon trace element data and ε_{Hf} values between +8 and +13 point to juvenile mantle-derived magmas (Arena et al., 2016).

Further north (present coordinates), slices of ophiolitic rocks are also present in the Araguaia Belt. The Quatipuru ophiolite is constituted by serpentinized peridotites intruded by mafic to ultramafic dykes (Paixão et al., 2008). Sm–Nd data of the dykes provide a whole-rock isochrone age of 757 ± 49 Ma, whereas ε_{Nd} values between +6.4 and +6.9 indicate a juvenile mantle source (Paixão et al., 2008).

Cryogenian oceanic crust is recorded in the southern Borborema Province of Brazil as well. The Monte Orebe ophiolite comprises basic metavolcanites, metacherts, garnet–mica schists and minor lenses of amphibolites and metaultramafic rocks (Caxito et al., 2014). Based on geochemical and Sm–Nd data, a juvenile depleted mantle source can be inferred for the metabasalts, which also present a whole-rock Sm–Nd isochrone age of 819 ± 120 Ma (Caxito et al., 2014).

In a similar way, Cryogenian ophiolites are present in the Anti-Atlas Belt, Morocco. The Bou-Azzer ophiolite comprises serpentinites, metagabbros, metabasalts and minor metasedimentary rocks, which are intruded by arc-related granodiorites, diorites and tonalites (El Hadi et al., 2010, and references therein). Geochemical data reveal a MORB signature for the metabasalts, although a second group with island arc affinity was recognized as well (Naidoo et al., 1991). A U–Pb SHRIMP zircon age of 697 ± 8 Ma obtained for a gabbro constrains the age of the ophiolite (El Hadi et al., 2010), which is further supported by ages of ca. 655–640 Ma of subsequent subduction-related intrusions (Inglis et al., 2005). On the other hand, the Tasriwine ophiolitic complex is mostly made up of metaultramafic rocks, metagabbros, leucogranites and mafic dykes (Samson et al., 2004, and references therein). Major elements and REE geochemical data reveal that leucogranites represent plagiogranites, which also present whole-rock ε_{Nd} and zircon ε_{Hf} values of ca. +6.0 and +14, respectively, thus indicating a mantle derivation (Samson et al., 2004). The age of the plagiogranites, in turn, is constrained at 762 ± 2 Ma by U–Pb TIMS zircon data (Samson et al., 2004).

In the Arabian–Nubian Shield, the YOSHGAH ophiolite belt records two Cryogenian events of oceanic crust formation at ca. 810–780 and 750–730 Ma (Ali et al., 2010, and references therein). The YOSHGAH belt comprises serpentinites, isotropic and layered metagabbros, pillow metabasalts and diabase dykes (Zimmer et al., 1995; Gahlan and Arai, 2009; Ali et al., 2010). In the Gerf nappe, Zimmer et al. (1995) reported a N-MORB signature for pillow basaltic lavas and sheeted dykes based on major and trace element geochemical data. These rocks also show ε_{Nd} between +4.5 and +8.8, further supporting a juvenile signature (Zimmer et al., 1995). Sm–Nd whole-rock isochrone ages of 720 ± 9 and 758 ± 34 Ma were obtained for gabbros and basalts, whereas one gabbro yielded a Sm–Nd mineral isochrone age of 771 ± 52 Ma (Zimmer et al., 1995). Kröner et al. (1992) reported Pb–Pb zircon evaporation ages of 741 ± 21 and 808 ± 14 Ma for a gabbro of the Gerf ophiolite and a plagiogranite of the Onib ophiolite, respectively. In turn, zircons from a layered gabbro from the Allaqi ophiolite present a weighted mean $^{206}Pb/^{238}U$ age of 730 ± 6 Ma (U–Pb LA-ICP-MS; Ali et al., 2010). Mineral chemistry and whole-rock geochemical data indicate that peridotites of the Allaqi ophiolite show compositions similar to fore-arc peridotites (Azer et al., 2013).

Hence, data indicate the presence of Cryogenian oceanic lithosphere younger than Rodinia break-up, although remnants of contemporaneous or even older oceanic rocks are present as well. The presence of older oceanic lithosphere could be explained by geological evidence supporting that several cratons were not part of Rodinia (Section 2, Fig. 1a; Cordani et al., 2003; Kröner and Cordani, 2003). Likewise, most Neoproterozoic ophiolites were associated with convergent settings, i.e. supra-subduction zones or island arc sequences (Fig. 8; e.g., Samson et al., 2004; Ali et al., 2010; Azer et al., 2013; Fortes de Lena et al., 2014), thus suggesting that they might represent limited events of extension of internal oceans rather than the development of major Cryogenian oceanic basins (Cordani et al., 2003; Johansson, 2014). The amalgamation of Western Gondwana thus resulted from introversion, which is further supported by paleogeographic reconstructions indicating that Western Gondwana assembly took place in the southern hemisphere, where most cratons were positioned since Rodinia assembly (Fig. 1a; Meert and Torsvik, 2003; Li et al., 2008; Evans, 2009; Evans et al., 2016). Nevertheless, extroversion can still be considered valid for the amalgamation of Western and Eastern Gondwana based on paleogeographic reconstructions (Tohver et al., 2006; Li et al., 2008; Evans, 2009), as indicated by Murphy and Nance (2003). Consequently, the assembly of Gondwana resulted from a combination of introversion and extroversion.

6. Concluding remarks

After assembly during the Late Mesoproterozoic, Rodinia underwent rifting at ca. 800–700 Ma leading to the opening of a major ocean between Laurentia and Eastern Gondwana cratons. In contrast, most Western Gondwana cratons occupied a marginal position in the southern hemisphere and recorded a different evolution during the same period, including subduction with back-arc extension, island arc development and limited formation of oceanic crust in internal oceans. Hence, paleogeographic reconstructions for the Tonian–Cryogenian, which classically consider a geodynamic scenario related to Rodinia break-up, need to be reevaluated.

The first collisional event during Gondwana assembly is recorded at ca. 630 Ma between the Río de la Plata and Congo-São Francisco cratons, which was succeeded by the assembly of the Amazonian and West African cratons to this early Gondwana nucleus up to ca. 600 Ma along the West Gondwana Orogen. These events are coeval with the onset of the opening of the Iapetus Ocean at ca. 610–600 Ma, which gave rise to the separation of Baltica, Laurentia and Amazonas and resulted from the final Rodinia break-up. The East African/Antarctic Orogen records the subsequent amalgamation of Western and Eastern Gondwana after ca. 580 Ma, contemporaneously with the beginning of subduction in the Terra Australis Orogen along the southern margin of Gondwana. Finally, the Kalahari Craton was incorporated during the Late Ediacaran–Early Cambrian. The proposed Gondwana evolution rules out the existence of Pannotia, as the final Gondwana amalgamation postdates latest connections between Laurentia and Amazonia. Likewise, the contemporaneous record of final Rodinia break-up and Gondwana assembly has major implications for the supercontinent cycle, as supercontinent amalgamation and break-up do not necessarily represent alternating episodic processes but overlap in time.

On the other hand, ε_{Hf} vs. U–Pb zircon age data from different Brasiliano–Pan-African belts show a fanning isotopic array indicating increased continental loss towards the timing of Gondwana assembly as expected for collisional orogenies (Collins et al., 2011; Roberts, 2012). Reworking of mostly Archean and Paleoproterozoic crust is recorded in the Damara Belt and the West Gondwana Orogen, being closely related to metacratonization of several crustal fragments in the latter, such as the Nico Pérez Terrane, the São Francisco Craton, the Borborema Province and the LATEA Metacraton. Remobilization of much younger crust and subordinated addition of juvenile Late Neoproterozoic continental crust took place along the northern African Gondwana margin. The Hf fingerprint of the assembly of Gondwana is thus controlled by metacratonization processes that, in turn, were strongly influenced by the pre-collisional location of active margins and back-arc zones. In contrast to crustal reworking during Gondwana assembly, crustal growth resulting from addition of juvenile continental crust along convergent margins was dominant since the Late Tonian in several Western Gondwana regions.

Finally, Late Tonian–Cryogenian oceans between Western Gondwana blocks were closed during the assembly of Western Gondwana, thus pointing to introversion. Nevertheless, pre-Neoproterozoic remnants of oceanic crust are present as well and can be explained as the result of isolation of several Western Gondwana cratons during the amalgamation of Rodinia. As extroversion is recorded during Western and Eastern Gondwana amalgamation, an alternative model of combined introversion and extroversion is proposed for the assembly of Gondwana.

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Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.gsf.2017.01.009.

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