A Full-Plate Global Reconstruction of the Neoproterozoic

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ABSTRACT

Neoproterozoic tectonic geography was dominated by the formation of the supercontinent Rodinia, its break-up and the subsequent amalgamation of Gondwana. The Neoproterozoic was a tumultuous time of Earth history, with large climatic variations, the emergence of complex life and a series of continent-building orogenies of a scale not repeated until the Cenozoic. Here we synthesise available geological and palaeomagnetic data and build the first full-plate, topological model of the Neoproterozoic that maps the evolution of the tectonic plate configurations during this time. Topological models trace evolving plate boundaries and facilitate the evaluation of "plate tectonic rules" such as subduction zone migration through time when building plate models. There is a rich history of subduction zone proxies preserved in the Neoproterozoic geological record, providing good evidence for the existence of continent-margin and intra-oceanic subduction zones through time. These are preserved either as volcanic arc protoliths accreted in continent-continent, or continent-arc collisions, or as the detritus of these volcanic arcs preserved in successor basins. Despite this, we find that the model presented here still predicts less subduction (ca. 90%) than on the modern earth, suggesting that we have produced a conservative model and are likely underestimating the amount of subduction, either due to a simplification of tectonically complex areas, or because of the absence of preservation in the geological record (e.g. ocean-ocean convergence). Furthermore, the reconstruction of plate boundary geometries provides constraints for global-scale earth system parameters, such as the role of volcanism or ridge production on the planet's icehouse climatic excursion during the Cryogenian. Besides modelling plate boundaries, our model presents some notable departures from previous Rodinia models. We omit India and South China from Rodinia completely, due to long-lived subduction preserved on margins of India and conflicting palaeomagnetic data for the Cryogenian, such that these two cratons act as 'lonely wanderers' for much of the Neoproterozoic. We also introduce a Tonian-Cryogenian aged rotation of the Congo-São Francisco Craton relative to Rodinia to better fit palaeomagnetic data and account for thick passive margin sediments along its southern margin during the Tonian. The *GPlates* files of the model are released to the public and it is our expectation that this model can act as a foundation for future model refinements, the testing of alternative models, as well as providing constraints for both geodynamic and palaeoclimate models.

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1. Introduction

Since its conception during last century, the theory of plate tectonics is intricately linked to advancements in geology. The theory provides a global framework, within which the microscopic textures and fabric of rocks can be reconciled through space and time with both the broad scale expression of geological features evident on the surface of our planet, and the dynamic evolution of the mantle underneath. Following Wegener's theory of continental drift (Wegener, 1912) and the adoption of plate tectonics during the 1960's by geoscientists, continents have been 'stitched' back together in order to understand long term trends and variations in mantle dynamics (e.g. Tackley, 2000), faunal diversity and evolutionary patterns (e.g. Cocks and Torsvik, 2002; Halverson et al., 2009; Meert and Lieberman, 2004; Valentine and Moores, 1972), distribution of ore deposits (e.g. Barley and Groves, 1992; Bierlein et al., 2009; Butterworth et al., 2016; Meyer, 1988; Pehrsson et al., 2016), seawater chemistry (Halverson et al., 2007; Hardie, 1996), palaeogeography and climate (Hoffman et al., 1998a; Kirschvink, 1992).

Reconstructing the interaction of plates along their boundaries – as opposed to reconstructing positions of continental blocks only – allows for testing and optimising reconstructions using plate tectonic and geodynamic principles, and assimilating the geological evidence preserved, especially along ancient subduction boundaries. Just as the predecessor of plate tectonic theory was continental drift, plate tectonic reconstructions are evolving from the relative (and absolute) motions of continental blocks to 'topological' plate models with closed plate boundaries that dynamically evolve through time (Gurnis et al., 2012). The importance of this paradigm shift of plate models is attributed to the significance that plate boundaries play in creating and destroying key geological environments, such as passive margins, island arcs and orogenic mountain belts, in better understanding hazards facing human society (e.g. volcanoes and earthquakes), and as an interface between mantle and crustal processes. Additionally, complete plate models create a foundation within which a broad range of hypotheses pertaining to topics such as magmatism, spreading rates, ocean basin size, mantle evolution and long-term palaeoclimatic variation can be tested more thoroughly than is possible through a model depicting only continental motions.

Plate models for the Cenozoic and Mesozoic (Müller et al., 2016; Seton et al., 2012), the late Palaeozoic (Domeier and Torsvik, 2014) and a bridging model between the two (Matthews et al., 2016) are available and global models for the Early Palaeozoic are partially complete (Domeier, 2016). However, as the amount of palaeomagnetic and, particularly, geological data that are available to constrain plate motions and boundaries increases, and software better able to compute topological plate models is developed (e.g. GPlates2.0 – gplates.org), the opportunity has arisen to construct such models further back in time in the Neoproterozoic. This is important as the transition from the Neoproterozoic to Phanerozoic is marked by the Ediacaran biotic revolution (Droser and Gehling, 2015), followed by the Cambrian explosion and the evolution of most present-day phyla and complex life (Meert and Lieberman, 2008). The Neoproterozoic was also a time of extreme climatic variation with (near?) global ice coverage over possibly as much as 60 million years at a time (Rooney et al., 2015). Many suggestions of how the planet descended into this ice-house world involve questions that can be addressed using a global topological plate model. For example, it has long been suggested that distributing continents around the tropics increases global albedo assisting with cooling the planet (Hoffman et al., 1998a; Kirschvink, 1992; Worsley and Kidder, 1991). However, more recently, volcanism has been identified as a possible cause, either by sulphuric acid aerosol production (Stern et al., 2008), weathering of terrestrial basalts (Goddéris et al., 2003; Cox et al., 2016) or by submarine hydration of volcanic glass produced along extensive ridge systems (Gernon et al., 2016). These suggested causal mechanisms, which relate to the tectonic geography of the times, and, in particular, the possible contribution of changing mid-ocean ridge lengths, can only be evaluated using a global topological plate model.

We present the first topological plate model for the Neoproterozoic, encompassing the evolution and breakup of Rodinia and subsequent amalgamation of Gondwana. The approach here is similar to that of both Domeier and Torsvik (2014) and Domeier (2016), with the model constrained by palaeomagnetic and geological data, as well as basic principles of plate kinematics (e.g. rates of movement). We stress that, considering the relative sparsity and uncertainty of data available to constrain plate kinematics and plate boundary configurations in the period covered by our study, there are likely to be many disagreements about the positions and movements of plates during this time. Furthermore, the configuration of boundaries within the oceanic domains of our model is almost entirely unconstrained by observations away from the continents. In such areas, we use simple plate tectonic rules to construct plausible plate boundary configurations that are consistent with our model for continental kinematics. To facilitate improvements and development of alternative models we make our plate model and all related files available to the public.

1.1. Previous models of Neoproterozoic supercontinents

Although the existence of a late Palaeozoic-Mesozoic supercontinent had been proposed in the early twentieth century, the possibility of earlier, Precambrian, supercontinents, is a much more recent suggestion. Piper et al. (1976) proposed a long-lived Pangaea ('Palaeopangaea') existing through the entire Proterozoic. This was based exclusively on palaeomagnetic data, and as data accumulated through the 1970's and 1980's, McMenamin and McMenamin (1990) suggested that a more plausible reconstruction involved two Neoproterozoic continents (East Gondwana and West Gondwana) that were derived from the breakup of a late Mesoproterozoic/early Neoproterozoic supercontinent, which they called Rodinia (from the Russian 'rodit', meaning 'to beget' or 'give birth'). Studies published by Dalziel (1991), Hoffman (1991) and Moores (1991) were the first papers to put forward a model for a Precambrian supercontinent, by integrating geological evidence with palaeomagnetic data. Principally, these reconstructions matched the east coast of Australia-Antarctica with the west coast of Laurentia (the cratonic part of North America) in the so-called SWEAT fit – **S**outh**w**est U.S. – **E**ast **A**ntarctica, and attached Amazonia to the east coast of Laurentia, whereas previously the only well-established connection was between Laurentia and Baltica (the cratonic part of Europe). The Laurentia-Australia connection was based on the similarly aged, and broadly congruent, sedimentary successions of eastern Australia-Antarctica and the Laurentian Cordillera, and an extension of the Grenville Orogen into the Shackleton Ranges of Antarctica (Dalziel, 1991; Moores, 1991). Hoffman (1991) also suggested a mechanism for the transition from Rodinia to Gondwana, with a 'fan-like' collapse of continents on the east and west side of Laurentia (India-Australia-Antarctica and Amazonia-West Africa respectively), around the Gondwanan nucleus of the Congo-São Francisco (C-SF) Craton (which rifted from the southern Laurentian margin) (Fig. 1a). Dalziel (1992) proposed a reconstructed Neoproterozoic supercontinent (Fig. 1b) that Torsvik et al. (1996) altered into a fit where Rodinia was more strongly constrained by the palaeomagnetic data available at the time, with a new configuration for Laurentia-Baltica-Amazonia (Fig. 3c). Both models followed Hoffman's (1991) suggestion of a 'fan-like' transition to Gondwana. These models were broadly supported by available palaeomagnetic data of McMenamin and McMenamin's East Gondwana, consisting of Australia, India and East Antarctica, which, at this time, was thought to act congruently, supported by the geological correlation of late Mesoproterozoic to early Neoproterozoic orogenies across India, Australia and Antarctica (e.g. Katz, 1989), and Laurentia proposed by Powell et al. (1993). Powell et al. (1993) suggested that Rodinia breakup occurred at ca. 700 Ma and final Gondwana amalgamation was at ca. 520 Ma, with East Gondwana remaining in low latitudes and Laurentia (including Baltica, Amazonia and West Africa) moving to polar latitudes. Finally, Dalziel (1997) capped nearly a decade of study since the proposal of Rodinia with a model spanning 725 to 422 Ma that summarised the consensus of previously published geological, palaeomagnetic and palaeontological data (Fig. 1c). Australia-Antarctica was placed against Laurentia in a SWEAT configuration, Baltica was placed against northern Greenland, with Amazonia placed slightly further down the northeast coast of Laurentia, connecting the Grenville Orogen in Laurentia with the Sunsas Orogen in Amazonia (Dalziel, 1992). The main deviation from the earlier, Dalziel (1992), model was that C-SF was tucked in closer to Amazonia along the eastern margin of Laurentia, rather than outboard of Kalahari, off the southern margin of Laurentia. This suggested that the western Gondwanan cratons were only separated by small ocean basins between Rodinia and Gondwana, and that they followed similar motion paths during the amalgamation of Gondwana. The model also considered the existence of Pannotia (e.g. Powell et al., 1995), a hypothesized, transient supercontinent occurring ca. 600 Ma, after amalgamation of Gondwana, but before the opening of the Iapetus Ocean (i.e. Laurentia, Baltica and Gondwana).

The suggestion that East Gondwana did not act congruently through the Neoproterozoic was originally proposed by Meert et al. (1995) based on geochronological data that suggested discrete phases of orogeny, between ca. 800 and 650 Ma and at ca. 550 Ma (Stern, 1994). They proposed that this represented a two-stage collision, between the Arabian-Nubian Shield, India, Madagascar, Kalahari and a small slither of Antarctica (a terrane they called IMSLEK), with Congo, as the earlier stage, and then the collision between this amalgamated continent with Australia-Antarctica as the later stage (Meert et al., 1995). Meert and Van der Voo (1997) extended this idea, describing the amalgamation of Gondwana as a series of three orogenic events. The Brasiliano Orogeny (between Congo and Amazonia-West Africa-Rio de la Plata), the East African Orogeny (between Congo and India-Madagascar) and the Kunngan Orogeny (between the rest of Gondwana and Australia-Antarctica). A key implication arising from their multi-stage amalgamation model was that irrespective of their positions in the early Neoproterozoic, India and Australia-Antarctica moved independently from one another during the transition from Rodinia to Gondwana. This was further supported by geochronological data from Antarctica,

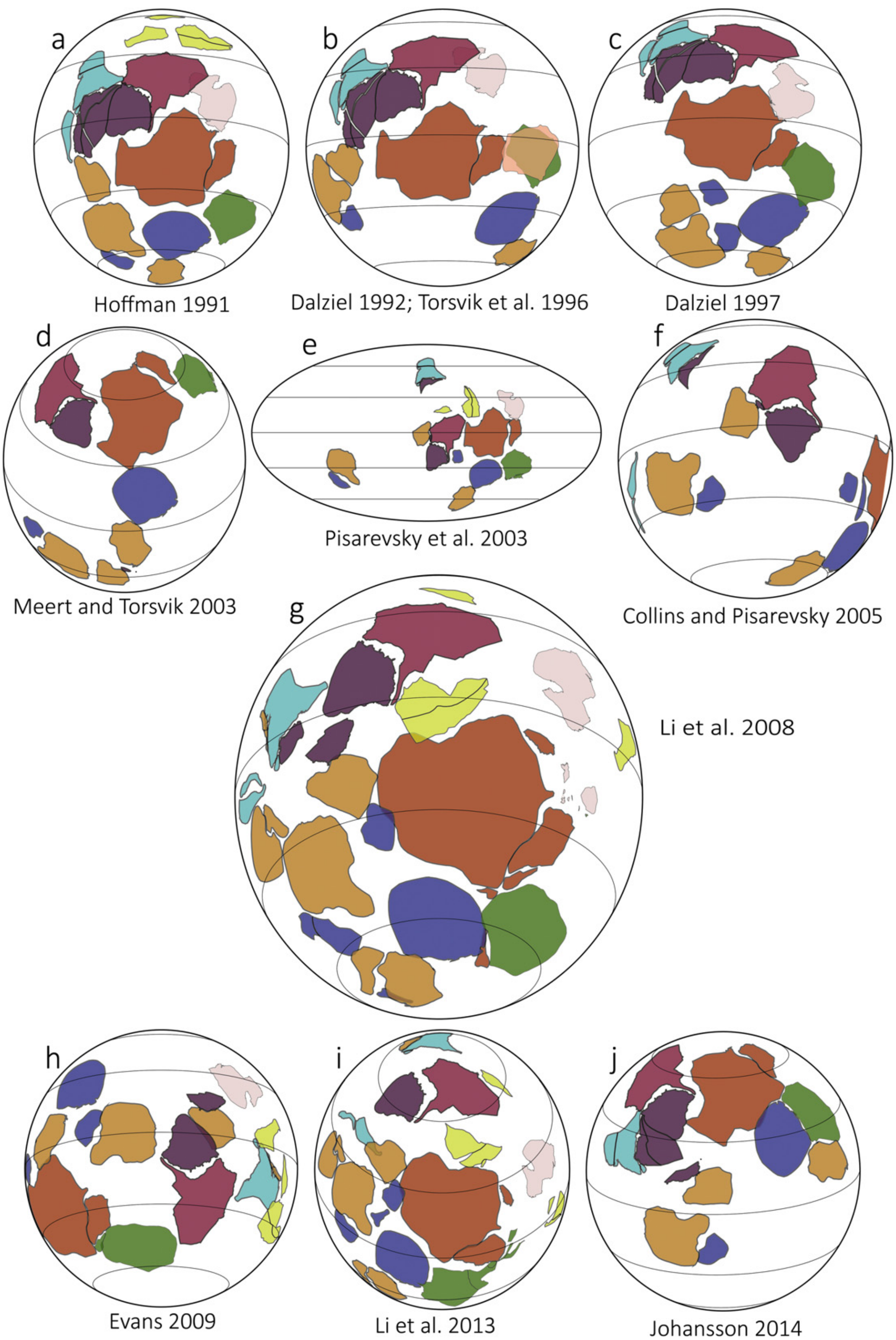

Fig. 1. Previous models of Rodinia reconstructed in *Gplates* using published Euler Pole rotations or reverse-engineered, if poles were not available. (a) Hoffman (1991) at 800 Ma; (b) Dalziel (1992) and Torsvik et al. (1996) at 750 Ma. The green Baltica belongs to Torsvik et al. (1996), and the pink belongs to Dalziel (1992). (c) Dalziel (1997) at 725 Ma; (d) Meert and Torsvik (2003) at 800 Ma; (e) Pisarevsky et al. (2003) at 800 Ma; (f) Collins and Pisarevsky (2005) at 750 Ma; (g) Li et al. (2008) at 900 Ma; (h) Evans (2009) at 900 Ma; (i) Li et al. (2013) at 800 Ma, and; (j) Johansson (2014) at 900 Ma. Cratonic blocks are colour coded by present day geography; North America, red; South America, dark blue; Baltica, green; Siberia, grey; India and the Middle East, light blue; China, yellow; Africa, orange; Australia, crimson; Antarctica, purple. (a)–(c), (e)–(i) in absolute framework; (d), (j) in Laurentian co-ordinates.

which indicated that the late Mesoproterozoic to early Neoproterozoic orogenies that were originally thought to tie East Gondwana together were separate events (Boger et al., 2001; Fitzsimons, 2000a,b), and by new high quality palaeomagnetic data from India and Australia at ca. 750 Ma that indicated at least a 30° latitudinal separation between them (Torsvik et al., 2001a,b; Wingate and Giddings, 2000). This led to a re-interpretation of the Darling Fault and the Leeuwin Block of south-west Western Australia (Collins, 2003), from being interpreted as a product of intracratonic deformation (e.g. Harris and Beeson, 1993) to representing a sinistral strike-slip tectonic boundary between two cratons moving independently (e.g. Fitzsimons, 2003), and suggested a staggered amalgamation of eastern Gondwana.

Palaeomagnetically constrained models of Rodinia without a unified East Gondwana were proposed by Meert and Torsvik (2003) (Fig. 1d) and Pisarevsky et al. (2003) (Fig. 1e). In addition to separating the constituents of East Gondwana during Rodinia breakup, both studies moved C-SF, Kalahari and India significantly from the 'traditional' model of Dalziel (1992, 1997). Meert and Torsvik (2003) analysed the possible positions of the major Precambrian cratons using palaeomagnetic data and moved C-SF, along with the Kalahari craton, to a distal position, outboard of Amazonia and West Africa. They also removed India completely from Rodinia. Comparably, Pisarevsky et al. (2003) removed both C-SF and India from Rodinia, and attached the Kalahari craton to the 'traditional' position of India, along the west coast of Australia-Antarctica, while leaving the other cratons in similar positions to Dalziel (1997). In their model, a ~7000 km ocean basin separated C-SF (and presumably the Saharan Metacraton (SM) and Hoggar) from Rodinia. Collins and Pisarevsky (2005) modelled the breakup of Rodinia to the amalgamation of Gondwana, with particular focus on the eastern Gondwana cratons (Fig. 1f). They adopted similar cratonic positions at 750 Ma as Pisarevsky et al. (2003) but focussed more closely on the closure of the Mozambique Ocean, and the suture between India and Congo. They drew attention to portions of Archaean-Palaeoproterozoic crust (Collins and Windley, 2002) preserved in modern day Madagascar, suggesting the existence of a small continent between Congo and India. They referred to this intra-East African Orogen continent as 'Azania' and proposed a two-stage collision of India-Azania-Congo, with initial closure of an ocean separating Azania and Congo (termed the Neomozambique ocean by Fitzsimons and Hulscher, 2005), followed by the collision of India with the then combined Congo and Azania, and the closure of the Mozambique Ocean sensu stricto (Collins and Pisarevsky, 2005).

The most holistic model of Neoproterozoic plate motions was by Li et al. (2008) as a contribution to IGCP 440, and that work underpins large parts of the model presented here (Fig. 1g). Using intersecting lines of evidence (e.g. palaeomagnetic data, geology, plate kinematics) a full reconstruction from 1100 to 520 Ma was built that depicted the assembly and breakup of Rodinia, and the transition from Rodinia to the amalgamation of Gondwana. Their interpretation of the temporal and spatial constraints of subduction zones, coupled with locations of dyke swarms, lead to the inferred presence of a large plume underneath Rodinia (e.g. Li et al., 2004). This model (re-) attached both India and C-SF to Rodinia on geological grounds. Coeval bimodal magmatism in India, South China and Australia was interpreted to indicate proximity between these cratons above a mantle plume (Li et al., 2003). Subsequently, India was placed in its 'traditional' configuration outside eastern Antarctica, with the accretion to Rodinia occurring between the early Neoproterozoic Eastern Ghats Belt of India with the Rayner Province of Antarctica. C-SF was attached to Rodinia based on U-Pb geochronological data indicating tectonism and magmatism between ca. 1050 and 950 Ma in the Irumide belt (de Waele et al., 2003), which was interpreted as an extension of the Grenvillian Orogeny. Inertial Interchange True Polar Wander (IITPW, Evans, 1998) was incorporated into the model between 820 and 750 Ma in order to account for a spread of palaeomagnetic data across South China (e.g. Li et al., 2004). The possible driver for IITPW was the hypothesized presence of the large mantle plume underneath Rodinia while it lay at a high latitude (a position suggested palaeomagnetically by the ca. 802 Ma Xiaofeng Dykes, Li et al., 2004). While acknowledging the variety of Australia-Laurentia configurations that are viable, they adopted the 'Missing-Link' configuration of Li et al. (1995), positioning South China in between Australia and Laurentia. Due to the large collaborative effort, as well as the broad range of evidence used, this model of Rodinia acts as a foundation for other global reconstructions, and also regional studies that seek to fit geological data into a larger Neoproterozoic picture.

In a marked departure from the 'traditional' configurations of Rodinia, Evans (2009) built an alternate model ('Radical Rodinia'), adhering strictly to palaeomagnetic data to define cratonic configurations (Fig. 1h). Laurentia still occupied a central position and Baltica was attached in a similar location to its 'traditional' position. Australia-Antarctica and India were placed outside of Baltica away from Laurentia, with Siberia and the three Chinese cratons placed outside of India, further away. C-SF was fitted against the northern coast of Laurentia and West Africa attached to the west coast of Laurentia, with Amazonia acting as a (large) promontory of Rodinia. A small intracontinental sea lay between Australia-Congo-Greenland. Although more recent palaeomagnetic data have disallowed parts of this configuration (e.g. Evans et al., 2016a, and references therein), a salient point of Evans (2009) is that any model presented will be a non-unique solution of the data available to constrain it and habitual familiarity with previous models can hinder us from approaching a 'true' reconstruction.

A new iteration of the Li et al. (2008) model that focussed predominantly on the analysis of sedimentary rocks deposited during the Cryogenian and Ediacaran and their relationship with both global glaciations and broad-scale geodynamics of supercontinent breakup was proposed by Li et al. (2013) (Fig. 1i). They updated palaeogeographic maps for the transition between Rodinia and Gondwana and included recent developments in Neoproterozoic tectonics, such as the intraplate rotation of Australia (Li and Evans, 2011), and tightened palaeomagnetic constraints. In order to account for the prevalent, late Tonian global evaporite deposits, a second, earlier pulse of IITPW was suggested to occur just prior to the IITPW event described in Li et al. (2008). This IITPW event would have dragged Rodinia from low-latitudes (where the evaporates were deposited) to higher latitudes at ca. 825 Ma, before the later event related to a high-latitude superplume in the mantle pulled Rodinia back to lower latitudes between 800 and 780 Ma (Li et al., 2013).

A derivative of the model proposed by Li et al. (2008) is the SAMBA (**S**outh **AM**erica-**BA**ltica) model of Johansson (2009, 2014), which is based upon a long-lived Proterozoic connection between Baltica and Amazonia that is preserved through both Nuna (the Mesoproterozoic supercontinent, e.g. Pisarevsky et al., 2014a and references therein) and Rodinia until the opening of the Iapetus Ocean from ca. 615 Ma (Johansson, 2009). This model of Rodinia depicts the positions of Australia, Siberia and Kalahari in similar positions to Li et al. (2008); Australia-Antarctica is depicted in a SWEAT configuration, although Johansson (2014) outlines that a Missing-Link configuration would also be compatible (Fig. 1j). Amazonia is rotated anti-clockwise from its position in Li et al. (2008), and is fitted into the (present day) southern margin of Baltica so that similar aged orogenic belts are aligned across the two cratons (from an Archaean core in the east, younging towards the west, and terminating with the ca. 1.1–1.0 Ga Sveconorwegian and Sunsas Orogens, which are linked to the Grenville Orogeny and Rodinia's amalgamation). West Africa is rotated 80° clockwise from its present-day position, and is placed outboard against both Amazonia and Baltica, in a tight fit that closes most of south and south-eastern Baltica off from an open ocean. However, the direction of tectonic growth of Palaeoproterozoic Baltica indicated by Bogdanova et al. (2015) would be perpindicular to the direction of coeval growth in Amazonia (in the Ventauri-Tapajos province), if they were in the SAMBA position, suggesting that this configuration may not be reliable for parts of the Proterozoic. The transition described by Johansson

(2014) from Rodinia to Gondwana would fit an 'orthoversion' type model of the supercontinent cycle (Mitchell et al., 2012), whereby the transition to the next supercontinent (i.e. Gondwana) occurs through ocean closure along the great circle of the girdle of subduction around the recently rifted apart supercontinent (i.e. Rodinia), similar in concept to the 'fan-like' collapse of Hoffman (1991).

An underlying assumption of previous models, and the model we present here, is that plate tectonics in the Neoproterozoic operated as they did in the Phanerozoic, and a key objective of this work is to provide the foundation for a geodynamic study that fully encompasses the transition from one supercontinent to another. The supercontinent cycle (e.g. Nance and Murphy, 2013; Worsley et al., 1984) presupposes cyclicity through plate divergence and convergence, necessitating, in some capacity, a mechanism for the motion of plates. Whether this motion during Precambrian times was through the coupling of the mantle with the crust similar to today, or through a different mechanism, such as 'lid tectonics' is uncertain (e.g. Roberts, 2013). The similarity between key cratonic configurations in Nuna and Rodinia (e.g. Meert, 2014a), despite both palaeomagnetic and geological data indicating relative significant motions of cratons between the two supercontinents (e.g. Pisarevsky et al., 2014) is suggested as by some workers evidence for a variation of plate tectonics, as the motions were either minor re-adjustments (e.g. Cawood et al., 2010, Baltica's rotation into Laurentia) or failed rifting events that reclosed along similar suture lines (e.g. Payne et al., 2009, Proto-SWEAT in Nuna). Nonetheless, Bradley (2011) outlined that most secular trends in the geological record were established either by or during the Neoproterozoic implying that modern day plate tectonics was probably recognisable during the Rodinia-Gondwana transition (Nance et al., 2014). An interesting note is that for all the new data collected since the original propositions of Rodinia (e.g. Dalziel, 1992; Hoffman, 1991), spatially, very little has changed in the overarching positions of most of the major Precambrian cratons; Laurentia still sits at the heart of Rodinia, Baltica and Amazonia on the east coast, Australia-East Antarctica on the west coast, C-SF and Kalahari on the south coast and Siberia off the north coast (Fig. 1).

2. Methodology

2.1. Foundations and starting points

The foundation of the plate tectonic model presented here is the Neoproterozoic reconstruction proposed by Li et al. (2008), which was principally based on geological observation and palaeomagnetic data from the late Mesoproterozoic to the Cambrian (ca. 1100–500 Ma). We supplement this with additional recent palaeomagnetic and geological data, and tectonic amendments (e.g. Evans et al., 2016b; Li and Evans, 2011; Pisarevsky et al., 2013) to create a new, topological plate model for the Neoproterozoic that encompasses the breakup of Rodinia and amalgamation of Gondwana. A description of both geological and palaeomagnetic data used to constrain cratonic configurations, positions and motions is provided in Sections 3, 4 and 5. We use cratonised blocks and terranes whose evolution can be tightly linked to the motions of larger cratons to create the model (Fig. 2a and b), with the expectation that other smaller, more poorly constrained terranes can be included in future iterations. The rationale for using the cratonised blocks is because during the late Neoproterozoic there was a major episode of continental crust building, involving the formation and accretion of many terranes to Gondwana along long-lived subduction zones (e.g. Arabian-Nubian Shield, Central Asian Terranes and Avalonian Terranes), and, due to the limited data available from the time, it is difficult to properly constrain both their absolute and relative positions.

Reliable palaeomagnetic data are the most useful data for deep time reconstructions, as they constrain continental palaeolatitudes at a specific time. However, major uncertainties include determining the age of magnetisation and accounting for any post-magnetisation deformation events. Consequently, palaeomagnetic data are often sparse for long periods of time, to the extent that all terranes and some cratons lack any reliable Neoproterozoic poles (e.g. Kalahari, West Africa, Azania, Amazonia, North China), while other blocks have no reliable poles for large time intervals (e.g. Australia between 1000 and 800 Ma, India between 1000 and 770 Ma and 750–550 Ma). Geological evidence is used to complement palaeomagnetic data, where age-dated tectonic geographic phenomena such as arc and rift magmatism, orogenic belt development, ophiolite formation and obduction, stratigraphic columns, metamorphism and shear zones help constrain both the motion of plates relative to one another, and the timing of key events (i.e. rifting and collision). In some cases, where there are no palaeomagnetic data available, the relative motions of a given craton are determined through geological observation and inference. An additional line of evidence used (in part as a 'sanity check' on plate motions) are plate kinematic rules, relating to how, geometrically, rigid plates move over a semi-rigid mantle. Using the evolution of the modern-day ocean basins as a guide, there are limits on the rates of motion and angular rotation that we expect plates to achieve. Such limits can help constrain a range of possible motions or help test competing plate configurations. As the majority of this reconstruction is before the evolution of complex life forms, palaeontological data is only useful for modelling the youngest time intervals (e.g. Meert and Lieberman, 2008).

2.2. Constructing the model

The plate reconstruction was built using the *GPlates* software (www.gplates.org). Gurnis et al. (2012) developed a topological functionality for *GPlates*, whereby plates can be defined by a finite list of boundaries (each boundary possessing its own Euler Pole) that dynamically evolve through time allowing for a 'continuously closed plate' (e.g. Fig. 2a). Plate boundaries in the Neoproterozoic are most tightly constrained around the margins of continents and/or continental crust; (i) where geological evidence for a subduction zone is preserved (e.g. ophiolites, subduction related magmatism, high pressure metamorphic rocks etc.), (ii) where there is evidence for continental rifting and breakup, so that two continents break apart and a spreading ridge forms (e.g. sedimentation, half grabens etc.), and, (iii) where there is transform motion between continental blocks preserved in the rock record (e.g. shearing, wrench tectonics). Away from continental crust, plate boundaries are more difficult to determine, so they must be inferred, either from connecting more tightly constrained boundaries, or from observing the motion of continental crust (such as from palaeomagnetic data) and interpreting the appropriate boundary requirement. Changes in the regional arrangement of plate boundary configurations (i.e. plate reorganisations) are linked, where possible, to broader global tectonic changes, such as initiation of rifting or subduction, since we have analogous evidence for plate reorganisation occurring as a consequence of these events in the Mesozoic-Cenozoic. For example, subduction cessation due to arc or terrane collision can cause reorganisations (e.g. Austermann et al., 2011; Patriat and Achache, 1984), as can subduction initiation (e.g. Faccenna et al., 2012; Whittaker et al., 2007), subduction of ridges or young buoyant oceanic crust (e.g. Matthews et al., 2012; Seton et al., 2015), and subduction of thickened crust (Knesel et al., 2008). Similarly, the arrival of plumes and superplumes is also thought to contribute to plate reorganisations (e.g. Cande and Stegman, 2011; van Hinsbergen et al., 2011), making times of supercontinent breakup critical for understanding and modelling plate motions and plate boundary changes. The process of constructing the model was completed iteratively to ensure that each boundary is both self-consistent with the paradigm of motion (i.e. convergence leads to subduction) as well as consistent with other boundaries forwards and backwards in time to ensure continuity.

The motions of the plates are described using a plate hierarchy similar to that of Seton et al. (2012) and Müller et al. (2016), such that a series of Euler poles describes each plate moving relative to another

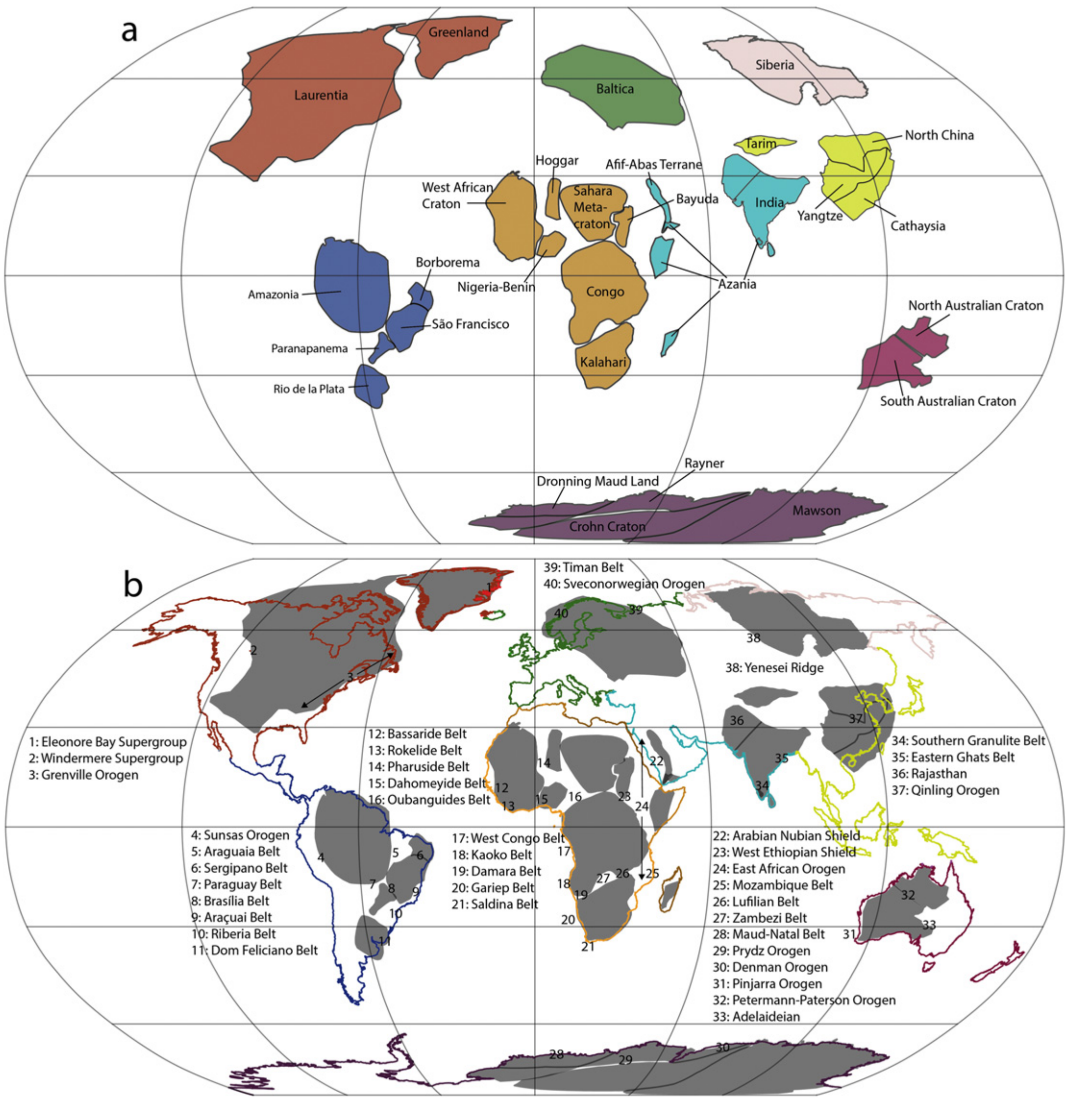

Fig. 2. (a) Map of Precambrian cratonic crust used in the reconstruction in their present-day locations. (b) Present day geographical map of the world with Precambrian cratonic crust used in the reconstruction in grey. Some key geological areas referred to sections 3, 4 and 5 are highlighted in (b). In both (a) and (b) cratonic crust and modern day continents are colour coded by their present-day position (consistent for all reconstruction figures).

(Fig. 3b). For the early part of the model (ca. 1000–700 Ma) the motions of plates are described relative to Laurentia, as it occupied the central position in Rodinia. From ca. 700 to 520 Ma, the motions of all plates except Laurentia, Baltica and Siberia are described relative to the Congo Craton, due to its central position in Gondwana. As there are no preserved ocean basins from this time, in some cases oceanic plates without a terrane or craton are constructed as a necessary requirement of plate tectonic theory. These plates have an arbitrary plateID and an artificial rotation to demonstrate that oceanic plates are implied to be present with new crust forming at spreading ridges, but there are no geological or palaeomagnetic constraints available to quantify the extent or velocity of these plates, rather their existence is inferred from the preservation of ocean-continent subduction zones in the geological record. The motion of synthetic oceanic plates is modelled to follow the general rule that slab pull dominates plate motions (Conrad and Lithgow-Bertelloni, 2002). By definition, convergence of oceanic plates towards active margins necessitates the modelling of divergent plate boundaries within the ocean basins. In these divergent settings, the precise geometry of these ridge and transform segments is synthetic and is drawn such that the orientation of the spreading segments follow great circles passing through the Euler Pole that describes the relative movement of one of the two plates, while transform segments offsetting this

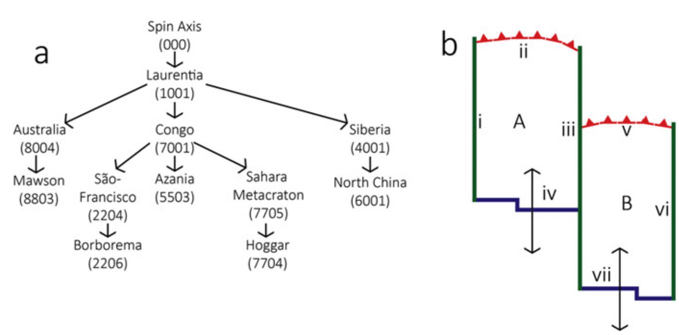

Fig. 3. Schematic depiction of (a) plate hierarchy (with some continents and their plateIDs), and (b) topological boundaries. In the plate hierarchy, all plates move relative to the plate above it (i.e. Hoggar moves relative to the Sahara Metacraton, which moves relative to Congo, which moves relative to Laurentia, which moves relative to the spin axis. The advantage of this is that it preserves relative plate motion between portions of crust that may not have any palaeomagnetic data. For example, the relative motion between Azania and Congo might involve some form of convergence inferred from geological data; there are no palaeomagnetic data to constrain Azania, but there are data to constrain Congo. Using a hierarchy, moving Congo into a palaeomagnetically defined position will preserve the relative rotations built on the geological basis between Azania and Congo, while allowing Congo to now better fit a pole. In (b), Plate A is defined by the boundaries consisting of (i), (ii), (iii) and (iv), Plate B is defined by the boundaries consisting of (iii), (v), (vi) and (vii).

spreading system follow small circles drawn around the same Euler Pole. We specifically point out that the large number and length of transform boundaries in the model are due to this simplification of boundaries and not an inherent indication that transform motions dominated Neoproterozoic tectonics.

3. Tonian evolution of Rodinia (1000–750 Ma)

3.1. 'North' Rodinia (Laurentia-Siberia ± North China)

Although a geological and palaeomagnetic connection between Siberia and Laurentia is generally accepted for the Mesoproterozoic and early Neoproterozoic (Ernst et al., 2016; Gladkochub et al., 2010a,b; Pisarevsky et al., 2014) there are large variations in possible configurations for Siberia as part of Rodinia (see Pisarevsky et al., 2008a for a review, e.g. Condie and Rosen, 1994; Frost et al., 1998; Hoffman, 1991; Rainbird et al., 1998, along the northern margin of Laurentia, or Sears and Price, 2000, along the western margin of Laurentia). We have attached Siberia to northern Laurentia following Pisarevsky and Natapov (2003), as it was adopted by Pisarevsky et al. (2008a, 2013) and Li et al. (2008). This configuration leaves a 30° gap between the northern margin of Laurentia and the southern margin of Siberia (possibly to be filled by minor North American terranes, e.g. 'Arctida' of Zonenshain et al., 1993). In the early Neoproterozoic Siberia was bordered by passive margins on its eastern, northern, western and maybe southern margins (Metelkin et al., 2012; Pisarevsky and Natapov, 2003). A subduction zone lay outboard of western margin of Siberia by ca. 880–860 Ma, which is represented by the Central Angara Terrane (Vernikovsky et al., 2016) that migrated to the continental margin by ca. 750 Ma (Vernikovsky et al., 2004; Metelkin et al., 2012). This subduction system has been proposed to be a continuation of the Valhalla Orogeny of Cawood et al. (2010) (e.g. Likhanov et al., 2014, 2015). Five terranes are identified, divided by the ENE sinistral Angara Fault. The oldest terrane (the Angara-Kan Terrane) is located on the southern side of the fault, and consists of Palaeoproterozoic, upper amphibolite-granulite metamorphosed volcanic and sedimentary rocks that overthrust the younger, Neoproterozoic aged island-arc assemblages of the Predivinsk Terrane (Nozhkin et al., 1999; Popov, 2001; Vernikovsky et al., 2004). The other three terranes are located on the northern side of the fault, and are Mesoproterozoic-Neoproterozoic aged marine sedimentary successions, consisting of carbonates, siliciclastics and ophiolites, which are generally metamorphosed to greenschist facies. The northern margin of Siberia transformed from a passive to active regime by ca. 750 Ma (Vernikovsky et al., 2004), with the accretion of the Central Taimyr terrane, which consists of volcanic and sedimentary units, during the late Neoproterozoic, at ca. 600 Ma. Here we match the east north-eastern margin of Siberia with the southern margin of North China, creating a nearly closed sea accounting for the thick sedimentation on the eastern margin of the Siberian Craton (Vernikovsky et al., 2004). The south-western margin of North China is outboard of Rodinia and here interpreted as a further extension of the Valhalla Orogen (and the early subduction outboard of Siberia), allowing for long-lived subduction and the collision of the North Qinling Terrane with the North China Block from ca. 900 Ma (Dong and Santosh, 2016; Wang et al., 2011). Considering the paucity of palaeomagnetic data for North China in the Neoproterozoic (Section 6.2), a position outboard of Australia could be an alternative to its position adjacent to Siberia. This is suggested by the similarity of their Mesoproterozoic positions (e.g. Pisarevsky et al., 2014; Zhang et al., 2012a) and their Palaeozoic positions (e.g. Metcalfe, 2006).

A relative motion of Siberia and North China to Laurentia along a dextral transform fault between ca. 800 and 720 Ma, which closes the gap between the two continents, is adopted here (Pisarevsky et al., 2013). This is based primarily on palaeomagnetic data (see Section 6). However, it does spatially associate the ca. 720 Ma mafic rocks of southern Siberia (e.g. the ca. 725 Ma Irkutsk large-igneous-province (LIP), Ernst et al., 2013, 2016) with the Franklin magmatic event in Laurentia, and suggests rifting from Rodinia during the mid Cryogenian from a position outboard of Greenland.

3.2. 'East' Rodinia (Laurentia-Baltica-Amazonia ± West Africa)

The formation of Rodinia in the late Mesoproterozoic–early Neoproterozoic necessitates a large (~95°) clockwise rotation of Baltica relative to Laurentia during the late Mesoproterozoic (Cawood et al., 2010), forming the Sveconorwegian Orogeny in Baltica and the Grenvillian Orogeny in Laurentia (Fig. 2b). Bingen et al. (2008) suggested that the primary phase of the Sveconorwegian Orogeny, consisting of granulite facies metamorphism, occurred between ca. 1050 and 980 Ma, and that orogenesis was over by ca. 920 Ma. After the Grenvillian-Sveconorwegian Orogeny, the Valhalla Orogeny initiated on the outboard margin of Greenland (Cawood et al., 2010) and possibly off Siberia and North China (see above). While the early stages of orogenesis here overlap with the final stages of the Grenvillian Orogeny, the two events are tectonically distinct, with the Valhalla Orogeny constituting an exterior rather than an interior orogeny due to the dominance of calc-alkaline magmatism (Cawood et al., 2010). The Valhalla Orogeny consists of two stages, the earlier Renlandian Orogeny, preserved in the Hebridean foreland of Scotland, from ca. 980 to 910 Ma, is characterised by peak metamorphism of upper amphibolite facies between 950 and 930 Ma (Cutts et al., 2009). This was followed by the Knoydartian Orogeny, from 830 to 710 Ma, preserved in the Moine succession in Scotland (Cawood et al., 2010). The Timan margin of Baltica was a passive margin throughout the Tonian and Cryogenian, with thick (>9000 m), predominantly terrigenous, sedimentary successions that were only mildly deformed during the Timan Orogeny (e.g. Cawood et al., 2007, 2016; Olovyanishnikov et al., 2000; Siedlecka et al., 2004). The eastern margin of Greenland and adjacent terranes also record thick sedimentary sequences during this time, preserved in the Krummedal and Eleonore Bay Supergroups in Greenland (e.g. Cawood et al., 2007; Strachan et al., 1995), in the Pearya Terrane of Canada (e.g. Malone et al., 2014) and in the Murchisonfjorden Supergroup of Svalbard (Gee and Teben'kov, 1996; Halverson et al., 2004). Malone et al. (2014) extends the Valhalla orogeny of Cawood et al. (2010) outboard of these passive margin sequences, suggesting that subduction continued (?) ~2000–3000 km offshore.

Similar to the Baltica-Laurentia connection, the Amazonia-Laurentia connection is also well established for the Neoproterozoic. The late Mesoproterozoic Ji-Paraná shear zone suggests a sinistral strike-slip movement of Amazonia from southern Laurentia up towards Baltica and northern Laurentia (e.g. Tohver et al., 2005a,b), and results in the 1300–1000 Ma Sunsas orogenic belt of southwest Amazonia (Litherland and Power, 1989; Santos et al., 2000, 2008a) being paired with the Grenville Orogen on the eastern margin of Laurentia (e.g. Hoffman, 1991; Li et al., 2008; Pisarevsky and Natapov, 2003; Sadowski and Bettencourt, 1996), just south of Baltica (Fig. 2b). Evans (2013) provides an alternative model for the collision, with a clockwise rotation of Amazonia into Laurentia between 1200 and 1000 Ma using new palaeomagnetic data. Evidence for a Baltica-Amazonia Grenvillian-aged collision is lacking, suggesting a small intracratonic sea separated the two, although some models propose paired movement of Baltica and Amazonia, suggested by an extension of the Mesoproterozoic belts from Amazonia into Baltica (e.g. SAMBA model, Johansson, 2009). Other models suggest that smaller late Mesoproterozoic terranes occupy this space (and hence the deformation), such as Oaxaquia (e.g. Cardona et al., 2010; Keppie and Ortega-Gutiérrez, 2010). We follow the 'traditional' configuration, with Amazonia linked to Laurentia along its western margin, matching the Sunsas Belt to the Grenville Orogen, proximal to Baltica after Hoffman (1991) and Li et al. (2008), but leaving space in between Amazonia and Baltica for minor terranes.

The West African Craton (WAC) remains enigmatic throughout the Neoproterozoic, with little information constraining its position for this time. By convention (and for ease of reconstruction) the WAC is attached to Amazonia in a configuration similar to its Gondwanan configuration (e.g. Hoffman, 1991; Li et al., 2008; Meert and Torsvik, 2003). The poorly dated early Neoproterozoic passive-margin sequences on the east and north coast of the WAC (e.g. Álvaro et al., 2014; Bouougri and Saquaque, 2004; Thomas et al., 2002, 2004), the Ougarta Aulacogen (Ennih and Liégeois, 2001) on the north-eastern margin and the Gourma Aulacogen in Mali to the southeast of the craton (e.g.; Ennih and Liégeois, 2001; Gasquet et al., 2005; Moussine-Pouchkine and Bertrand-Sarfati, 1978) suggest rifting between the WAC and an unknown craton by 800–750 Ma at the latest. The northern passive margin is 4–5 km thick and consists predominantly of volcano-sedimentary rocks, that grade from shallow marine facies of carbonates and quartzite, cross-cut by tholeiitic dolerite dykes (Álvaro et al., 2014; Thomas et al., 2002, 2004) to overlying deeper-water turbidites (Ennih and Liégeois, 2001; Fekkak et al., 1999). Ophiolites are also preserved further outboard from the margin (Samson et al., 2004), and the protoliths of tonalitic gneisses and biotitic schists were dated to ca. 750 Ma (Thomas et al., 2002), suggesting that the northern-north eastern extent of the WAC had transitioned to an island arc setting, with the WAC as the downgoing plate. Based on the position of C-SF during the early Neoproterozoic as constrained by palaeomagnetic data (Section 6.5), we infer that this unknown craton is the western margin of C-SF, and that this rifting propelled C-SF northwards along a dextral transform fault against the Rodinian margin towards the Kalahari Craton and Australia-Antarctica. This rifting event also resulted in minor separation between WAC and Amazonia (e.g. Paixão et al., 2008).

The opening of the Iapetus Ocean through the breaking up of Laurentia-Baltica-Amazonia was one of the final events in Rodinia's dispersal. Early extension along the eastern margin of Laurentia is recorded in rhyolitic lava flows in the central Appalachians between ca. 760 and 700 Ma (Aleinikoff et al., 1995) and the subsequent deposition of the Pinelog Formation (Li and Tull, 1998). However, a hiatus in deposition during the late Cryogenian (Li and Tull, 1998) suggests two phases of rifting, and that this earlier rifting was either aborted (e.g. Novak and Rankin, 2004) or related to rifting on the western margin of Laurentia (e.g. Cawood et al., 2001). The later phase of extension consists of a multi-stage rifting event (Cawood et al., 2001; Pisarevsky et al., 2008b) above a mantle plume (Puffer, 2002). The ca. 615 Ma Long Range and Egersund dyke swarms (in Laurentia and Baltica respectively, Bingen et al., 1998) demonstrate the first stage and suggests opening of the northern Iapetus Ocean (Baltica-Laurentia) and the Tornquist Sea (Baltica-Amazonia) (Bingen et al., 1998; Pisarevsky et al., 2008b; Puffer, 2002), although the deposition of deep water carbonates and development of the passive margin in north-eastern Laurentia is thought to have occurred later, in the Early Cambrian, from both stratigraphic controls and faunal distributions (Allen et al., 2010; Cawood et al., 2001; Lavoie et al., 2003). The second stage, the opening of the southern Iapetus Ocean, between Laurentia and Amazonia occurred later and was slightly more prolonged, with rift-related magmatism evident in a series of three pulses; the tholeiitic, ca. 590 Ma Grenville swarm (Kamo et al., 1995; Seymour and Kumarapeli, 1995) and ca. 580 Ma Blair River metagabbro dykes (Miller and Barr, 2004); the 565–560 Ma Sept Isle intrusion and Catoctin volcanics (Aleinikoff et al., 1995; Higgins and van Breemen, 1998); and the ocean island basalt (OIB)-like 550 Ma Skinner Cove volcanics (McCausland et al., 1997; Puffer, 2002), and development of a passive margin by ca. 540 Ma (Cawood et al., 2001). To reconcile robust palaeomagnetic data that indicate a large ocean basin between Laurentia, Baltica and Gondwana by 550 Ma (e.g. Cawood et al., 2001; Lubnina et al., 2014; Pisarevsky et al., 2008b), with a rift-drift transition by 540 Ma along the Laurentian margin, Cawood et al. (2001) proposed a more complex breakup model, whereby a rift-drift transition occurred at 570 Ma, followed by another rifting event at ca. 540 Ma, where a smaller microcontinent was rifted from the Appalachian margin. This second rifting event led to the development of the thick passive margin preserved today, and the microcontinent is typically thought to be the Argentine Precordillera (Allen et al., 2010; Cawood et al., 2001). However, since coeval dyke swarms or magmatism are yet discovered in Amazonia, a proper temporal constraint on the opening of the southern Iapetus remains enigmatic. Here our model depicts the earlier, 570 Ma, rift-drift transition, and is compatible with later rifting of the Argentine Cordillera.

3.3. 'West' Rodinia (Laurentia-Australia-Antarctica ± Tarim)

The pairing of the west coast of Laurentia with Australia-East Antarctica was one of the original motivators for suggesting Rodinia due to the thick sedimentary sequences (e.g. Rainbird et al., 1996; Young, 1981; Young et al., 1979), dyke swarms (Ernst et al., 2008) and rifted margins on their respective west and east margins. The original pairing of the two continents linked southwest Laurentia with East-Antarctica (SWEAT, Dalziel, 1991; Hoffman, 1991; Moores, 1991). Three other specific configurations were proposed based on palaeomagnetic and geological grounds. **AUS**tralia-**MEX**ico (AUSMEX), which pairs northern Queensland (Australia) with the southwest tip of Mexico, was proposed by Wingate et al. (2002) due to mismatches in the late Mesoproterozoic palaeomagnetic record of Laurentia-Australia. In this configuration, the conjugate margin to Laurentia is usually considered to be South China or Tarim (e.g. Pisarevsky et al., 2003). **AUS**tralia-**W**estern **U**nited **S**tates (AUSWUS), pushes Australia further south down the Laurentian coast so that Australia (and not Antarctica as in SWEAT) matches south west Laurentia. This configuration was originally proposed using Neoproterozoic-aged transform faults (as extensions of transform offsets of the spreading centre) to assist in matching stratigraphy between Australia and Laurentia (Brookfield, 1993). It was extended by Karlstrom et al. (1999, 2001) by suggesting that the Grenville Orogen, which is truncated in southern Laurentia, extended through into Australia and is expressed as the Albany-Fraser and Musgrave Orogens, on the basis of broadly similar tectonic histories and by Burrett and Berry (2000, 2002) who matched key geological provinces (e.g. Broken Hill-Olary block with Mojave terrane) that allowed a tighter fit between Australia and Laurentia. The Missing-Link model, originally proposed by Li et al. (1995), locates Antarctica in a similar position to SWEAT, although Australia is offset further away from northern Laurentia, with the South China craton sitting in between Laurentia and Australia (as the 'Missing-Link'). This model corrects some of the stratigraphic mismatches between Laurentia-Australia as well as connecting the temporally similar Sibao and Gardiner dyke swarms (Li et al., 1999, 2008). It also permits later breakup to occur as it more easily fits the ca. 750 Ma Mundine Dyke Swarm palaeomagnetic pole (Wingate and Giddings, 2000).

Rifting, and the transition from rift-to-drift, between Australia and Laurentia is poorly constrained to between 825 and 700 Ma due to difficulties in dating sedimentary sequences of the Adelaidian complex (e.g. Preiss, 2000). The 827 Ma Gardiner Dyke Swarm and flood basalts in the Adelaidian fold belt are typically seen as the first stage of rifting, although there are no matching swarms in Laurentia until 780 Ma (Park et al., 1995; Wingate et al., 1998) (the Missing Link model has coeval plumes in South China at ca. 827 Ma, Li et al., 1999). Comparatively, sedimentary sequences suggest later rifting, with deep-water sediments being more prevalent after 750 Ma. Palaeomagnetic data, which originally constrained breakup to ca. 750 Ma at the latest (Wingate and Giddings, 2000), now permit rifting until ca. 720–700 Ma with an intraplate rotation of the Northern Australian Craton (NAC) relative to the South Australian Craton (SAC) for some configurations (Li and Evans, 2011). Unfortunately, neither geological or palaeomagnetic data clearly discriminate a rifting time. We adopt here the earlier rifting at ca. 800 Ma with an AUSWUS configuration, to minimise the relative spreading velocity and angular rotation of Australia-Antarctica, but we note that later rifting events before ca. 770 Ma are kinematically feasible, and that the model here is broadly compatible with an AUSMEX configuration.

Tarim is typically attached to northern or north-western Australia during the Neoproterozoic (e.g. Huang et al., 2005; Li et al., 1996, 2008; Wen et al., 2013), although a tectonic model that fully integrates geological and palaeomagnetic data and its relationship with other blocks has only been recently proposed (Ge et al., 2014). U-Pb dating of zircons from granitic gneisses in the North Qaidam belt of (southern) Tarim and long lived subduction in the Quruqtag region along its northern margin during the Neoproterozoic are generally interpreted as evidence of it being included as part of Rodinia in a peripheral position, with its northern margin facing an open ocean (Ge et al., 2014, 2016; Liu et al., 2015; Shu et al., 2011; Song et al., 2012; Yu et al., 2013). A transition between ca. 830 and 780 Ma to retreating accretion on the northern margin, opening of the South Tianshan Ocean and deposition of the volcanic rock bearing, clastic Beiyixi Formation (Ge et al., 2014; Xu et al., 2005), 820–795 Ma mafic-ultramafic intrusives (Zhang et al., 2007) and a series of ca. 775 to 750 Ma mafic dyke swarms (Zhang et al., 2009) are suggested to be related to the break-up of Rodinia and motion of Australia-Antarctica and Tarim (on a single plate) away from Laurentia. Younger, 650–630 Ma dyke swarms (Zhu et al., 2008), ca. 615 Ma basalts, the geochemistry of which suggest an intracratonic rift environment (Xu et al., 2013), and sediment deposition forming the Sugetbrak Formation are interpreted here as indicators of the rifting away of Tarim from northern Australia during the intraplate rotation of the North Australia Craton into the Southern Australian Craton (Li and Evans, 2011). While we find the geological evidence convincing for having Tarim fixed off the northern margin of Australia in Rodinia, this position does not fit three reasonably reliable Tonian-Cryogenian aged palaeomagnetic poles (815–795 Ma Aksu dykes, Chen et al., 2004; 760–720 Ma Qiaoenbrak Formation, Wen et al., 2013, and; 770–717 Baiyisi Volcanics, Huang et al., 2005), although it does fit a younger three (also reasonably reliable) palaeomagnetic poles (635–550 Ma Sugetbrak Formation, Zhan et al., 2007; ca. 635 Ma Tereeken Cap Carbonate, Zhao et al., 2014; 621–609 Ma Zhamoketi Andesite, Zhao et al., 2014). To fit all poles, Tarim would have to be located in equatorial-low latitudes between ca. 800 and 700 Ma, and then transition and rotate ~180° to be in mid-high latitudes during the Cambrian. Palaeontological data from both the Cambrian and Ordovician suggest it was in close proximity to Australia, South China and North China at that time (e.g. Chen and Rong, 1992; Metcalfe, 2006, 2013; Rong et al., 1995). We find this motion difficult to account for given the global context of other portions of cratonic crust and inferred spreading systems; consequently, we omit Tarim from this model prior to 700 Ma, and expect that future work on palaeomagnetic data can help resolve this problem, although we do note recent proposals where Tarim acts as an extension of South China in a 'Missing-Link' position (Wen et al., 2017).

Reconstructions of Phanerozoic Gondwana and the cyclical opening and closing of the Tethyian oceans place all southeast Asian terranes and cratons outboard and proximal to both Australia and India (e.g. Burrett et al., 2014; Metcalfe, 2006, 2011, 2013). The larger cratons (South China, North China, Tarim) occupy the most distal positions from Gondwana, with the smaller terranes (e.g. Indochina, Lhasa, Qiantang, Sibumasu) located between them and the north Gondwanan margin, suggesting that their formation must predate the Palaeozoic Gondwanan configuration. Detrital zircon and palaeontology are used to infer affinity of these terranes with either Australia or India (or both). Typically, from west (India-affinity) to east (Australia-affinity), Indochina and Qiantang are placed outboard of northwest India, due to a peak of detrital zircon ages at ca. 950 Ma, suggesting input from the subduction outboard of NW or N India (e.g. Usuki et al., 2013). Indochina also contains a peak of zircons between ca. 700 and 500 Ma, likely from the East African Orogen, so it is inferred to be placed further west than Qiantang (Burrett et al., 2014). Comparably, the Lhasa terrane contains slightly older detrital zircon peaks of ca. 1170 Ma, which are typically associated with the Albany-Fraser Orogen of Australia (e.g. Zhu et al., 2011). The Sibumasu Terrane is inferred to share affinity with Australia due to similar faunal assemblages during the Palaeozoic (e.g. Fortey and Cocks, 1998; Metcalfe, 2011). Sparsely exposed basement in the North Lhasa terrane of southern Tibet consists of Neoproterozoic aged granulites (Zhang et al., 2012b, 2014). U-Pb dating of zircon cores and in-situ zircons from the granulites gives a protolith age of ca. 900 Ma, with metamorphism occurring at ca. 650 Ma (respectively), inferred to have occurred as a response to Gondwana amalgamation. Whole rock geochemistry suggests that the terrane protolith was oceanic crust (Zhang et al., 2012a,b). On the northern border of the Lhasa Terrane, the Amdo basement preserves an orthogneiss with a protolith age of 900–820 Ma, with a similar metamorphic overprint age as in the North Lhasa terrane (Guynn et al., 2006, 2012). Comparatively, the Sibumasu Terrane, placed outboard of Lhasa in Gondwana reconstructions, does not preserve any Neoproterozoic signatures. Younger protolith ages as young as ca. 740 Ma derived from U-Pb dating of inherited zircon cores have also been reported from the Nyainqnetanglha Group in southern Lhasa (Dong et al., 2011). Assuming a close link between western Australia and Lhasa they could be related in some capacity to 750 Ma granitoids of the Leeuwin Complex (Collins, 2003). We (tentatively) propose that subduction outboard of Australia was occurring for most of the Neoproterozoic, during the Tonian as an extension of a circum-Rodinia subduction zone, and during the Cryogenian as a response to Australia-Antarctica rifting from Laurentia, and is preserved in Lhasa and other southeast Asian terranes.

3.4. 'South' Rodinia (Laurentia-Kalahari-Rio de la Plata)

Southern Rodinia is probably the least constrained of all the Rodinian margins. We have followed the convention of previous models in attaching Kalahari to southern Laurentia, and fitting the other sizable South American craton, Rio de la Plata (RDLP), into the gap between the southeast Laurentian margin and Kalahari (e.g. Hoffman, 1991; Jacobs et al., 2008; Li et al., 2008). The Kalahari craton itself consists of an Archaean-Palaeoproterozoic nucleus that underwent substantial growth during the Mesoproterozoic (e.g. Hanson, 2003; Hanson et al., 2006; Jacobs et al., 2008). The Namaqua-Natal-Maud belt on its southern and southeastern margins is associated with the Rodinia amalgamation, and, depending on its orientation, was suggested as an extension of the Grenville Orogen, with the collision of Kalahari with the Coats Land Block (Loewy et al., 2011). Rifting during the breakup of Rodinia is preserved on the northwest-west-southwest-south margins of Kalahari (Jacobs et al., 2008) while any rifting on the eastern margin has been overprinted due to the high grade metamorphic events occurring with the amalgamation of eastern Gondwana. The Kalahari-Laurentia configuration is updated to that of Loewy et al. (2011) where the south and south-west margins have collided with RDLP and Laurentia, allowing the Maud-Natal Belt (Fig. 2b) to act an extension of the Grenville Orogen. This has two additional benefits; firstly, it creates a larger space between Australia-Antarctica and Kalahari, allowing more space for an AUSWUS configuration, or even for an AUSMEX-like configuration if desired. Secondly, it removes the need for the development of a thick passive margin on the eastern margin of Kalahari, as its relationship with Australia post Rodinia breakup is along a transform boundary.

The RDLP craton is more difficult to constrain than the Kalahari craton due to it being almost completely overprinted and reworked during Gondwana amalgamation. There are only minor suggestions of a Mesoproterozoic aged tectono-thermal event based on K-Ar muscovite ages (Basei et al., 2000) and detrital zircon (Gaucher et al., 2008). Previous studies have proposed a central positioning of the RDLP craton to Rodinia (e.g. Fuck et al., 2008; Gaucher et al., 2011), with it being positioned at the locus of Amazonia, Kalahari and Laurentia. This is based on the abundance of late Mesoproterozoic detrital zircons present in Ediacaran-aged sandstones, suggesting that the RDLP was ringed by Mesoproterozoic orogenic belts (Gaucher et al., 2011), although recent isotope geochemical studies have suggested that the eastern margin faced an open ocean from 800 Ma (e.g. Koester et al., 2016; Lenz et al., 2013).

We adopt a central position of RDLP on the southeastern margin of Laurentia between Kalahari and Amazonia as a small cratonic fragment recording part of the Grenvillian Orogeny (Gaucher et al., 2011). This puts RDLP into a strong position for its preferred model of amalgamation into Gondwana, through a series of dextral transform faults and oblique subduction (Rapela et al., 2007). There are little data to provide an accurate model of rifting, as such we suggest rifting from Laurentia at 590 Ma on the basis that RDLP forms the southern extent of the opening of the Iapetus Ocean, although we note that earlier rifting is permissible (provided it occurs after Kalahari has rifted away from Rodinia).

3.5. 'Extra South' Rodinia (Congo, Azania, Arabian Nubian Shield ± Sahara Metacraton)

The Congo-São Francisco craton (C-SF), a Palaeoproterozoic amalgam of Archaean cratons, has a disputed history during the Neoproterozoic due to the absence of large scale, continental-continental collision and related high pressure metamorphism as is evident in other Rodinian cratons (de Waele et al., 2008). Consequently, there are a number of models wherein C-SF is not part of Rodinia (e.g. Collins and Pisarevsky, 2005; Meert, 2003; Pisarevsky et al., 2003). Comparably, the C-SF forms the nucleus of Gondwana, because the craton is surrounded by a series of Ediacaran-Cambrian orogenies (Section 5), including the Brasiliano Orogen that traces the suture between Amazonia and C-SF, and the East African Orogen, consisting of the Kuungan Orogen (sometimes referred to as Pinjarra, or Pinjarra-Denman-Prydz Orogen) that stitches Antarctica together, the Malagasy Orogen, that connects India with C-SF through Madagascar and Azania, and the Damara Orogen, between the Congo and Kalahari cratons (Collins and Pisarevsky, 2005; Fitzsimons, 2003; Meert, 2003).

The present-day southern margin of the African Congo Craton preserves evidence for quite diverse Tonian tectonic environments along its length. The basement of the central Damara Belt of Namibia preserves evidence for Stenian orogenesis that is correlated with similar-aged volcanic-arc plutonism and volcanism in the Rehoboth inlier (Becker and Schalk, 2008). This reflects collision between the Kalahari craton and the C-SF at the end of the Mesoproterozoic (Becker et al., 2005). Evidence for early and mid-Tonian tectonism exists as rift-basin formation along the southern C-SF margin before ca. 760 Ma (McGee et al., 2012a; Miller, 2013). Extensive rifting along this margin, however, does not occur until ca. 750 Ma as shown by voluminous volcanic rocks and intrusions occur (Hoffman et al., 1996). Further east along this margin, in the Lufilian Arc of Zambia and the Democratic Republic of Congo, deposition of the syn-rift lower Roan Group, the basal group of the Katanga Supergroup, is constrained to after ca. 877 Ma, the age of the nonconformably underlying Nchanga Granite (Armstrong et al., 2005; Miller, 2013). Carbon isotope stratigraphy coupled with a sequence stratigraphy suggests that Tonian rifting ended, but then rejuvenated prior to the deposition of the basal Cryogenian Grand Conglomérat Formation of the Nguba Group, which overlies the Roan Group (Bull et al., 2011). In contrast to Namibia, considerable earlier Tonian magmatism occurs within the basement of the Katanga Supergroup. As well as the aforementioned Nchanga Granite, magmatism in the Lusaka area is dated at 820 ± 15 Ma (Johnson et al., 2007), which terminates rift basin formation in the Lusaka-Zambezi region that began with ca. 880 Ma felsic extrusive rocks (Johnson et al., 2007). We interpret the Tonian extension that these igneous rocks represent, reflects the increasing proximity of the Congo Craton to an active plate margin.

The Southern Irumide Belt lies to the south-east of the Lufilian Arc and the Zambezi supracrustal rocks. Here, latest Mesoproterozoic gneisses are interpreted to represent a Stenian volcanic arc terrane that accreted to the C-SF during the Irumide orogeny (Johnson et al., 2005, 2006; Karmakar and Schenk, 2016). Further east, in north-east Mozambique, high-grade metamorphism is early Tonian (ca. 950 Ma; Bingen et al., 2009), suggesting progressive accretion on this active margin along the south-east apex of the C-SF craton (Johnson et al., 2005; de Waele et al., 2008). Younger Tonian volcanic arc systems are preserved within the East African Orogen, east of the C-SF in Gondwana, and outboard of the Mozambique Tonian accretionary system. The Dabolava Arc of west Madagascar formed over a similar period to the Southern Irumide and Mozambique arc terranes (Tucker et al., 2007, 2011, 2014). It, however, formed over a west dipping subduction zone before colliding with central Madagascar (Azania) at ca. 950 Ma (D.B. Archibald, unpublished data). These arc-continent collisions preceded the intrusion of ca. 930–900 Ma igneous rocks in the Malagasy Vohibory volcanic arc terrane (Collins et al., 2012; Jöns and Schenk, 2008). This event in Madagascar may correlate with the Galana Arc of south-east Kenya (Hauzenberger et al., 2007) and the so-called TOAST terrane of East Antarctica (Jacobs et al., 2015). Finally, the major ca. 850–750 Ma continental margin arc that resulted in the emplacement of the Imorona-Itsindro Suite (Archibald et al., 2016; Archibald et al., in press; Boger et al., 2014; Handke et al., 1999) developed within Azania. Geochemical data are equivocal for determining subduction polarity (Archibald et al., in press), therefore subduction polarity was either from eastward-directed subduction of the ocean separating Azania from the C-SF continent, or, (as is portrayed here), from continued west-dipping subduction, but with a trench east of Azania subducting the Mozambique Ocean between central Madagascar and India.

The northern margin of C-SF is marked by the poorly known Oubanguides Belt (Fig. 2b) in Africa (Bouyo et al., 2013; de Wit et al., 2008; Poidevin, 1985; Toteu et al., 1994, 2001, 2006; Van Schmus et al., 2008) and the Sergipano Belt in Brazil. The Oubanguides consists of pre-Neoproterozoic and juvenile 1100 to 625 Ma thrust sheets emplaced over the Lindian Supergroup, a Neoproterozoic passive margin succession on the north margin of the C-SF (de Wit and Linol, 2015; Toteu et al., 2006). Deformation and metamorphism of these belts occurred in the Cryogenian to Ediacaran and are discussed below.

The Sahara Metacraton (SM) is typically placed to the north of the C-SF. However, whether it was fixed to the same plate as the C-SF or had some relative movement (and if so, how much?) is unknown, principally because palaeomagnetic data are unavailable. The term 'metacraton' refers to a quasi-stabilised section of crust that was remobilised to some degree and exhibits deformation throughout its entire area, not just its margins, while still maintaining geological coherency (i.e. never fully rifted apart) (Abdelsalam et al., 2002; Liégeois et al., 2013). The SM is referred to as a 'metacraton' by Abdelsalam et al. (2002) because during the Neoproterozoic it neither appears to have acted rigidly and congruently as cratons do, nor does it appear to be dominated by juvenile accreted terranes and be fully tectonically mobile as is the Arabian-Nubian Shield (ANS), for example. The SM is dominated by medium to high grade gneisses, migmatites and the intrusion of post 750 Ma granitoids (Abdelsalam et al., 2002), and is tectonically and geologically distinct from both the other African cratons that are surrounded by thick continental orogenies, and the lower grade, greenschist facies, volcano-sedimentary metamorphic rocks of the ANS (e.g. Johnson et al., 2011). The southern parts of the SM were recently shown to be made of composed late Mesoproterozoic crust (de Wit and Linol, 2015). The western SM boundary has been taken to be the Raghane shear zone, as it juxtaposes the Neoproterozoic aged Assodé-Issalane terrane, which exhibits amphibolite facies metamorphism during Gondwana amalgamation, from the older Aouzegueur terrane that exhibits only greenschist facies metamorphism (Henry et al., 2009; Liégeois et al., 1994). The primary movement of this shear zone is thought to have occurred no later than 590–580 Ma (e.g. Abdallah et al., 2007; Henry et al., 2009) due to the emplacement of undeformed granitoids that exhibit distinct magnetic foliations aligned with the Raghane shear zone (Henry et al., 2009). The eastern margin is typically defined by the sinistral, Keraf Suture, preserved in northern Sudan, which juxtaposed the younger, juvenile arc terranes of the ANS against Mesoproterozoic-aged rocks of the eastern SM (e.g. Abdelsalam et al., 1998) between 640 and 600 Ma (Johnson et al., 2011; Stern et al., 1989). ^{40}Ar-^{39}Ar dating on biotite and hornblende from a deformed granite suggest that shear zone deformation ended by ca. 580 Ma (Abdelsalam et al., 1998). The Keraf Suture was inferred to be a response to a sinistral, transpressive tectonic regime related to the closing of the Mozambique Ocean and amalgamation of Gondwana (Abdelsalam et al., 1998) (Section 5.1).

Pulses of 920–900 Ma magmatism and amphibolite facies metamorphism are recorded in the Bayuda Block in North Sudan, south of the Keraf Suture (Evuk et al., 2014; Karmakar and Schenk, 2015; Küster et al., 2008), with slightly younger 850–760 Ma oceanic arc magmatism recorded in the Western Ethiopian Shield (WES) (Ayalew et al., 1990; Blades et al., 2015; Johnson et al., 2011), and outboard the Bayuda Block at ca. 805 Ma, where geochemical trends and volcaniclastic sedimentary successions suggest the existence of a back-arc basin (Küster and Liégeois, 2001). Finally, collisions of arc terranes with the Bayuda Block at ca. 700 Ma are used to support the existence of an ocean basin between the ANS and SM during the Tonian (Evuk et al., 2014; Küster and Liégeois, 2001), although the main phase of compression and deformation occurred later during Gondwana amalgamation.

The ANS is a series of juvenile Neoproterozoic island arc terranes that accreted to the Sahara Metacraton during the East African Orogeny. The tectonic history of the ANS is complicated, perhaps similar to the island arcs of present day southeast Asia, where, if they were to be accreted to Asia, would preserve a complex melange of terranes, accreted arcs, extension and back-arc basins, subduction polarity reversals and changes in major stress regimes. Recent syntheses (e.g. Fritz et al., 2013; Johnson et al., 2011) have summarised the geological and tectonic history of the area, but, without palaeomagnetic data, it is difficult to constrain its position, relative to both the SM and C-SF. Early Neoproterozoic subduction follows on from late Mesoproterozoic subduction in west Sudan/Chad (de Wit and Linol, 2015) and the Sinai (Be'eri-Shlevin et al., 2012; Eyal et al., 2014). These are followed by a distinct period of quiescence during the mid Tonian (ca. 930–880 Ma) followed by voluminous volcanic arc development between ca. 870 and 830 Ma (Robinson et al., 2014, 2015a) and accretion of these terranes to the SM during the late Cryogenian to Ediacaran (ca. 650–580 Ma, Johnson et al., 2011, Section 5.1). The Afif terrane contains a small Palaeoproterozoic sub-terrane (the Khida terrane - Stacey and Hedge, 1984; Stoeser et al., 2001; Whitehouse et al., 2001), and potentially Archaean-aged basement rocks based on Nd model ages (Abas terrane - Whitehouse et al., 1998, 2001; Windley et al., 1996), suggesting that they could represent a northern extension of Azania, acting as a semi-continuous palaeogeographic, but not necessarily tectonically congruent, archipelago.

Fritz et al. (2013) identified four distinct, but overlapping stages of accretion. Broadly, the first two stages are pertinent to Rodinia, while the latter two relate to Gondwana amalgamation and are a response to the closure of the Mozambique Ocean and East African Orogeny (Section 5.1). In the first stage, arc age decreases from the Tokar/Barka Terrane in the south (870–840 Ma), towards the Hijaz Terrane in the north (780–710 Ma), with suturing between them occurring from 830 to 710 Ma (Johnson et al., 2011; Johnson and Woldehaimanot, 2003). The second stage involved the formation of the 810–710 Ma Midyan-Eastern Desert terrane further north, against the interior of the older Sa'al Terrane, and its suturing with the earlier, southern terranes between ca. 760 and 730 Ma along the E-W Yanbu-Onib-Sol Hamed-Gerf-Allaqi-Heiani suture, forming the western arc or oceanic terranes of the ANS (Stoeser and Frost, 2006). The older, further east, Afif-Jiddah-Abas terranes amalgamated with the oceanic terranes close to the Cryogenian-Ediacaran boundary (ca. 650 Ma). Subduction is inferred to be east dipping under the Afif-Jiddah-Abas terranes, which conflicts with the slightly earlier, supposed west dipping subduction in the more westerly WES and Bayuda Block, perhaps suggesting a subduction polarity reversal during the late Tonian, or a spreading system with subduction occurring on each side. The final stages relate specifically to the suturing of the youngest arcs (post ca. 680 Ma) on the eastern margin of the Afif Terrane and the closure of the Mozambique Ocean, and young to the east, away from the ANS (Cox et al., 2012).

The western margin of the Saharan Metacraton consists of a series of (primarily) dextral strike-slip faults and a regime of transpressive tectonics (e.g. Fezaa et al., 2010). The evolution of Hoggar and the west Saharan Metacraton represents a series of accreted arcs, with a transition to a dextral-transpressive tectonic environment during the late Neoproterozoic as the WAC collided with it (Fezaa et al., 2010; Liégeois et al., 2003). The suture of the Hoggar Shield and Saharan Metacraton is the Raghane shear zone (Henry et al., 2009; Liégeois et al., 1994), and was tectonically active from ca. 730 to 580 Ma, at the conclusion of which most of western Gondwana was amalgamated. Early suturing was of the easternmost terrane of the Hoggar shield (the Aïr Massif) was with the Sahara Metacraton between 700 and 670 Ma (Liégeois et al., 1994). Tectonic models for the evolution of the central and western parts of the Hoggar Shield are more convoluted. Proposed tectonic models suggest some separation between the western, central and eastern (which are attached to the Sahara Metacraton) terranes allowing for a complex history of subduction to occur amongst 23 identified terranes (Black et al., 1994; Caby, 2003). The central terranes, principally the LATEA (**L**aouni, **A**zrou-n-Fad, **T**efedest, **E**géré-**A**leksod) terranes (Liégeois et al., 2003), acted as the nucleus for the majority of the Hoggar Shield, as they had a series of smaller terranes accreted onto them between 900 and 580 Ma (Caby and Monié, 2003; Liégeois et al., 2003). Initially, subduction was away from the central terranes, with the Iskel Island Arc accreting onto the western margin of LATEA by ca. 850 Ma (Liégeois et al., 2003). Further subduction to the west resulted in a series of small terranes (such as the ca. 690–650 Ma Pharusian terrane, Caby, 2003) being accreted onto LATEA by 620 Ma (Caby, 2003). Subduction was east dipping under Hoggar for this collision (Berger et al., 2014).

Further south from the Hoggar Shield are the Benin-Nigeria and Borborema provinces of present day Africa and South America respectively. Their congruency during the Neoproterozoic is interpreted from their broadly similar lithologies and tectonic history. These consist of an Archaean nucleus with rift-related magmatism occurring during the Palaeoproterozoic (ca. 1.85–1.73 Ga), similar detrital zircon spectra and a range of metasedimentary and metavolcanic rocks with Sm-Nd model ages between 1.6 and 1.0 Ga (Arthaud et al., 2008; Brito Neves et al., 2000; Kalsbeek et al., 2012). Both provinces sit on the eastern margin of the cryptic Transbrasiliano Lineament (that divides C-SF and the SM from Amazonia and the WAC), suggesting affinity with C-SF and/or the SM prior to Gondwana amalgamation. The Borborema Province, in particular, was strongly deformed and reworked during Gondwana amalgamation, as it was located at the locus of convergence between the Amazonia, West Africa, Congo, São Francisco and Sahara cratons (e.g. Brito Neves et al., 2000; Ganade de Araujo et al., 2014a,b, Section 5.8). An earlier magmatic event, the Cariris Velhos orogeny, is preserved towards the southern margin in the Transversal Zone in the Borborema (e.g. Brito Neves et al., 1995; Santos et al., 2008b). Here, 1000 to 920 Ma gneisses, migmatites and volcanoclastic sequences suggest a continental-arc environment, perhaps with the development of a back-arc basin, but without considerable crustal thickening occurring at its conclusion (Caxito et al., 2014; Santos et al., 2010; Van Schmus et al., 2008). The Tamboril-Santa Quitéria Complex of the Ceará Central Domain in western Borborema records long-lived subduction from the mid-Tonian until Gondwana accretion. U-Pb dating of zircon constrain the timing of earliest subduction to be from ca. 900 Ma, although the development of a juvenile arcs is suggested to be from 880 to 800 Ma due to the emplacement of tonalite and granodiorite with positive $\varepsilon_{Hf}(t)$ and $\varepsilon_{Nd}(t)$ and a low $^{87}Sr/^{86}Sr$ ratio (Ganade de Araujo et al., 2012, 2014a). Juvenile arc signatures persisted until the late Cryogenian-early Ediacaran (ca. 660–650 Ma), when subduction matured due to the subduction of the approaching WAC (Section 5.9) (Ganade de Araujo et al., 2014b).

Owing to the uncertainty of Borborema's position in Rodinia (e.g. Fuck et al., 2008) and the slightly younger ages of the Cariris Velhos orogeny, it is difficult to determine how Borborema relates to wider Rodinian tectonic geology. We leave Borborema fixed to the SM, and infer this subduction to be part of the closure of an ocean basin between the wider SM Plate and the C-SF Plate, although the (present-day) western margin of Borborema faced an open ocean.

We include C-SF (and by extension, SM) as a distal part of Rodinia, although with some relative movement between it and the Rodinian core of Laurentia-Baltica-Amazonia-Kalahari-Australia. The subduction along the eastern margin (i.e. Azania, ANS etc.) acts as a leading edge of the continent as it moves along a dextral fault, relative to Rodinia, from outboard of WAC to outboard of Kalahari.

4. Tonian evolution of India and South China

Both India and South China are enigmatic continents throughout the Neoproterozoic, as data are available from both cratons without providing a strong discriminatory argument for both their position and motion. Inevitably, this has led to the proposition of a number of positions of both continents. For India, suggestions have included omitting it from Rodinia altogether (e.g. Pisarevsky et al., 2003; Collins and Pisarevsky, 2005); if it was in Rodinia, it has been placed outboard of East Antarctica (e.g. Li et al., 2008), or outboard of Australia (e.g. Meert and Torsvik, 2003). South China has been put as the heart of Rodinia (Li et al., 1995, 2008), or attached to Western Australia (e.g. Cawood et al., 2013; Niu et al., 2016). Generally, the only well established (and agreed) upon positions and motions of India and South China are from the late Neoproterozoic, where their palaeomagnetic (e.g. Zhang et al., 2013) and geological (e.g. Collins et al., 2014) data suggest a southward motion due to subduction of the Mozambique Ocean under the Gondwana nucleus, with South China located north of India such that it can rift off uninhibited during the Palaeozoic (see Section 5.1).

The core of India consists of a series of Archaean-aged cratons overlain by a series of Proterozoic basins, separated by Proterozoic-aged orogenic belts. These cratons make up two main groups that formed separate Mesoproterozoic continents. A northern Bhundelkund craton was separated from a southern cratonic group, consisting of the Dharwar, Bastar and Singbhum cratons, by the Central Indian Tectonic Zone (see Meert et al., 2010 for a recent synthesis). The timing of amalgamation between these continents has been controversial (Acharyya, 2003; Mishra et al., 2000; Rekha et al., 2011; Roy and Prasad, 2003; Roy et al., 2006; Stein et al., 2006), but recent work on the age of collisional metamorphism in the Sausar Orogeny that sutured the continents, demonstrated that peak orogenesis occurred at ca. 1.06 Ga (Bhowmik et al., 2012). Sedimentological studies of late Mesoproterozoic rocks in the Chhattisgarh Basin where an oceanic connection through the CITZ is inferred from the basin architecture and the documentation of tidal bedforms support this hypothesis (Patranabis-Deb, 2004; Saha et al., 2016). Prior to this latest Mesoproterozoic orogenesis, India was at least two separate continents, with the beginning of the Neoproterozoic marking the amalgamation of the kernel of Peninsular India.

The northwest boundary between this latest Mesoproterozoic core of India and the Stenian-Neoproterozoic crust that extends west from the Delhi-Aravalli Orogen, is a major lithospheric boundary visible on deep seismic reflection surveys where it is interpreted as reflecting an oceanic suture pre-dated by east-dipping subduction beneath the Bhundelkund Craton (Rao and Krishna, 2013) (Fig. 2b). We suggest that this boundary is the ultimate 'east' margin of the northern East African Orogen (see below).

Stenian-Tonian times saw the progressive accretion of volcanic arc rocks to this NW margin of Neoproterozoic India, this accretion extends through the basement of Pakistan; into eastern Arabia, where Tonian-aged arc rocks occur as inliers in Oman (Alessio et al., in press; Allen, 2007; Bowring et al., 2007; Whitehouse et al. 2016). Afghanistan lies between the Indian Subcontinent and Oman in Gondwana reconstructions. Archaean rocks were reported from the Kabul Block (Collett et al., 2015; Faryad et al., 2016) that show ca. 2.8–2.5 Ga protoliths overprinted by a ca. 1.85–1.80 Ga metamorphic event, with a further, lower temperature, tectonothermal event in the Tonian. We suggest that this terrane forms a microcontinent accreted to the active NW Indian margin in the Tonian. Faryad et al. (2016) noted the similarity in tectonothermal events recorded in the Kabul Block to those in South China, but dismissed this possible link due to palaeogeographic considerations. However, we suggest that the accretionary history of western South China (i.e. the Yangtze Block), as a series of terranes assembled to an older eastern continental terrane (Cawood et al., 2013), is remarkably similar to that of NW India as detailed above. Therefore, we support the observation of Faryad et al. (2016) and suggest that the Kabul Block is a similar micro-continental terrane to that proposed for the pre-Neoproterozoic of the Yangtze Block of South China.

The Seychelles microcontinent shares similar geology to west Rajasthan, with granitoids intruded between ca. 810 and 750 Ma (Stephens et al., 1997; Torsvik et al., 2001b; Tucker et al., 2001). Similar-aged (ca. 750–730 Ma) juvenile arc granitoids are also found in the northern Bemarivo Belt in NE Madagascar (Collins, 2006; Collins and Windley, 2002; Thomas et al., 2009; Tucker et al., 1999; Tucker et al., 2014). We propose that these presently disparate regions, which were Gondwana neighbours, also form part of the Tonian-aged accretionary complex northwest of India and west of South China.

The Eastern Ghats Orogen (on the eastern margin of India), and the Krishna Orogen to its south, were active subduction zones for most of the Mesoproterozoic and early Neoproterozoic (Dobmeier and Raith, 2003; Henderson et al., 2014) (Fig. 2b). The orogens are tectonically complex, and were divided into discrete terranes based on lithological differences (Ramakrishnan et al., 1998), isotope provinces (Rickers et al., 2001) or crustal provinces based on distinct geological histories between different areas (Dobmeier and Raith, 2003). Recent geochronological and petrological studies confirmed the distinction between the ca. 1.6 Ga Krishna Orogeny in the southern Eastern Ghats region (Henderson et al., 2014) and the ca. 1.0–0.9 Ga Eastern Ghats Orogeny sensu stricto in the north (Korhonen et al., 2013). The Eastern Ghats Orogen is linked with the similar-aged Rayner province in Antarctica, due to a similar tectonic and metamorphic history, as well as similar Nd model ages for gneisses in both areas, suggesting that the combined Rayner/Eastern Ghats complex grew as a consequence of the long-lived subduction on the eastern margin of India (Dobmeier and Raith, 2003; Mikhalsky et al., 2013; Rickers et al., 2001). The outboard margin of the Rayner/Eastern Ghats complex is marked by the Archaean Ruker terrane, the southernmost rocks exposed in East Antarctica, at ca. 960 Ma, followed by full closure by ca. 900 Ma (Boger et al., 2008; Corvino et al., 2008; Dasgupta et al., 2013; Mezger and Cosca, 1999). This prolonged magmatism/metamorphism during the early Tonian served as evidence of India's amalgamation with Rodinia (e.g. Dasgupta et al., 2013; Li et al., 2008), however, here we follow the suggestion that India was a separate continent and that this subduction represents the final accretion of Indian-Antarctica with India (e.g. Liu et al., 2013). Later Tonian and Cryogenian tectonothermal events rework the Eastern Ghats orogen between 900 and 650 Ma (Dobmeier and Simmat, 2002; Gupta, 2012; Nanda et al., 2008), which, although 'intra-continental', suggests that a plate margin outboard of the exposed Rayner/Eastern Ghats Complex still periodically transmitted compressive stress inland.

We adopt the position of Cawood et al. (2013) for a connection between South China and India in the late Mesoproterozoic, and preserve this for the entire Neoproterozoic (i.e. no relative motion between South China and India). South China is also reduced to its Cathaysian core at the beginning of the Neoproterozoic, with the Yangtze 'craton' here interpreted as a Tonian-aged accretionary complex in a manner following Cawood et al. (2013). The Jiangnan Orogen in central South

China preserves evidence for a Tonian (ca. 870–860 Ma) accretionary complex that post-dates earlier subduction/accretion complexes (Wang et al., 2015) and pre-dates final continent formation in this region (Yao et al., 2016). Subduction polarity was suggested to dip to the west (present-day coordinates), suggesting that the western Archaean-Palaeoproterozoic core of the Yangtze Block may have formed a microcontinental kernel for Tonian arc-accretion that collided with the Cathaysia block at ca. 860 Ma (Yao et al., 2016). As discussed above, we correlate this Yangtze Block core with the Kabul Block and suggest that the accretionary record of South China is coeval with the accretionary history of NW India (Bhowmik et al., 2010, 2012; Cawood et al., 2013; Meert et al., 2013; Roy and Prasad, 2003).

5. Rodinia to Gondwana: geology of the Gondwana-forming orogens

5.1. East African Orogen

Rocks deformed and metamorphosed in the East African Orogen (Stern, 1994) extend, in a reconstructed Gondwana, from the Eastern Mediterranean in the north (e.g. Candan et al., 2016), through Arabia (including NW India/Pakistan/Afghanistan), eastern Africa, Madagascar, southern India, Sri Lanka to East Antarctica where outcrop disappears beneath the ice south of Lützow-Holm Bay and Sør Rondane (Fig. 2b). The orogen is likely to follow the subglacial East Antarctica Mountain Range (An et al., 2015) to the Gambutschev suture (Ferraccioli et al., 2011), where, it meets the Kuunga Orogen (Meert et al., 1995; also known as the Pinjarra-Prydz-Denman Orogen; Fitzsimons, 2003). Together, these orogens delineate the west, south and east margins of Neoproterozoic India.

The northern East African Orogen is commonly called the Arabian-Nubian Shield (Johnson et al., 2011) and is characterised by voluminous juvenile Neoproterozoic crust that formed as a series of volcanic arcs, dominantly younging away from an older continental terrane (Robinson et al., 2014, 2015a,b). Less extensive continental terranes exist in the region, particularly in the Sinai (Be'eri-Shlevin et al., 2012; Eyal et al., 2014), in the Khida and Afif Terranes of Saudi Arabia (Stoeser et al., 2001; Whitehouse et al., 2001) and in Yemen (Whitehouse et al., 1998, 2001; Windley et al., 1996) with corollaries along the southern Gulf of Aden escarpment (Collins and Windley, 2002; Sassi et al., 1993; Whitehouse et al., 2001). The eastern margin of the East African Orogen in Arabia is often left at the margin of the exposed Neoproterozoic in Saudi Arabia, but similar magnetic anomalies to the eastern-most exposed Saudi terrane (the Ar-Rayn terrane, Cox et al., 2012; Doebrich et al., 2007) occur beneath the Ediacaran Rub Al-Khali Basin of Saudi Arabia (Johnson and Stewart, 1995) and where Precambrian basement is exposed in the east of the Arabian Peninsula—in Oman—it is again Neoproterozoic juvenile crust that formed in volcanic arc tectonic environments (Alessio et al., in press; Bowring et al., 2007; Whitehouse et al., 2016). This leads us to extend the East African Orogen to regions of Gondwana east of the Arabian Peninsula—regions that now form the basement of southern Afghanistan, Pakistan and NW India (see also Cozzi et al., 2012, and above).

The Mozambique Belt is the common name for the southern East African Orogen and here the Tonian and pre-Tonian terranes described earlier came together in two main orogenic events. The earlier one occurred at ca. 650–640 Ma and forms the time of peak metamorphism in Uganda, Kenya, Tanzania and in northern Mozambique (Appel and Schenk, 1998; Fritz et al., 2013; Hauzenberger et al., 2004, 2007; Tenczer et al., 2013). This orogenic event was interpreted as being due to intra-arc extension (Appel and Schenk, 1998), based on its anticlockwise path, but on a regional scale, it correlates with the amalgamation of the main Arabian-Nubian Shield along the Keraf Suture to the north, and is focussed along the suture of Azania with the C-SF. This is particularly apparent in Madagascar where ca. 650–640 Ma metamorphic ages occur in the west of the country (Jöns and Schenk, 2008), whereas the east of the country is dominated by younger metamorphic ages of ca. 570–540 Ma (Collins et al., 2003; Jöns and Schenk, 2008; Tucker et al., 1999, 2014). This younger, eastern, orogenesis correlates with the Ediacaran arc accretion recorded in the far east of the Saudi Arabian Shield that separates the exposed Saudi Shield from the basement of Oman. These observations led Collins and Pisarevsky (2005) to propose that the western ca. 650–640 Ma orogenesis was due to late Cryogenian collision of Azania with the C-SF continent (the East African Orogeny sensu stricto; e.g. Meert and Van der Voo, 1997; Stern, 1994), while the younger ca. 570–540 Ma orogenesis was due to the final collision of Neoproterozoic India with the then amalgamated Azania/C-SF continent, closing the Mozambique Ocean (and forming the Malagasy Orogeny; Collins and Pisarevsky, 2005). Studies from southern India support this hypothesis because orogenesis in the Southern Granulite Belt is focussed at 570–520 Ma and forms a part of the Malagasy Orogeny (Clark et al., 2015; Collins et al., 2007a,b, 2014; Johnson et al., 2015; Plavsa et al., 2012, 2014, 2015; Taylor et al., 2015; Kumar et al., 2016).

5.2. Oubanguides-Sergipano Orogen

The Oubanguides and Sergipano belts run roughly E-W across central Africa and South America, separating Congo and the SM in Africa, and between the São-Francisco Craton and the Borborema Block in South America (Fig. 2b). The Oubanguides are interpreted to continue between the two cratons towards Sudan in the east, though they crop out most prominently, and have only been studied, in the west; in Cameroon, Chad and the Central African Republic (e.g. Pin and Poidevin, 1987). Broadly, the Oubanguides consists of Palaeoproterozoic basement thrust over the Congo Craton, with intense metamorphism (granulite facies) following collision (e.g. Toteu et al., 2004). It is divided into three tectonic units separated by shear zones. These are, from south to north, the schistose and gneissic Yaoundé Domain, which contains extensive nappes thrust over the Congo Craton (e.g. Owona et al., 2011), the reworked Palaeoproterozoic Adamawa-Yadé Domain, intruded by numerous syn- to late tectonic ca. 630–570 Ma granitoids, and the volcano-sedimentary schistose and gneissic Western Cameroon Domain, which was intruded by calc-alkaline granites between 660 and 580 Ma (Toteu et al., 2001). Dating of metagabbroic rocks in the northern Yaoundé domain show the earliest intrusions to be 660 ± 22 Ma (Toteu et al., 2006), likely due to slight extension as a response to subduction further south. This facilitated deposition within the Yaoundé Basin (Toteu et al., 2006). Deformation and metamorphism are constrained to ca. 660–600 Ma (Nkoumbou et al., 2014; Owona et al., 2011; Rolin, 1995; Toteu et al., 2006). Dating of granitoids from shear zones within the Western Cameroon belt (northernmost unit) constrain a period of sinistral shear motion to ca. 620 Ma, and a younger, more dominant, period of dextral shear at 580 Ma, likely due to collision of Amazonia and WAC with the Gondwana nucleus (Kwékam et al., 2010).

This correlates with the Sergipano Belt of NE Brazil where metamorphism and deformation is constrained to ca. 630–570 Ma (Neves et al., 2016; Oliveira et al., 2015). A continental margin arc is suggested here at ca. 630 Ma, overprinted by syn- continent-continent collisional granitoids emplaced between 590 and 570 Ma (Oliveira et al., 2015). The Sergipano Belt consists of five domains sutured by shear zones; from south to north these are the Estância, Vaza Barris, Macururé, Poço Redondo-Marancó and Canindé Domains (Davison and Dos Santos, 1989; Oliveira et al., 2010). The three southerly domains consist predominantly of metasedimentary and sedimentary rocks. From south to north, the domains are interpreted to represent the foreland basin (Estância domain), grading towards a fold thrust belt consisting of slightly deformed sedimentary rocks of a shallow platform (Vaza Barris domain) with provenance from the south (i.e. São Francisco), and deformed metasedimentary rocks (Macururé domain) derived from the north (i.e. Borborema) (Oliveira et al., 2010). The only age constraints available for these domains are from the Macururé Domain, where 620–570 Ma granitoids intrude (Bueno et al., 2009; Silva Filho et al., 1997). Here, the ca. 620 Ma granites are inferred to be emplaced in a continental arc environment (Bueno et al., 2009). The two more northerly domains are more varied in their composition; the Poço Redondo-Marancó contains metasedimentary rocks deformed up to amphibolite facies, with calc-alkaline andesitic-dacitic rocks likely derived from a volcanic or continental arc, and large granitoid intrusions. Ages determined from the volcanic arc rocks (Carvalho et al., 2005), and granitic intrusions (Oliveira et al., 2015) suggest magmatism spanned from ca. 630 to 600 Ma. The Canindé Domain contains a broad range of rock units, leading to multiple interpretations of its tectonic environment, including ophiolitic remnants, island arc accretions and intracontinental magmatism (Jardim de Sá et al., 1986; Oliveira and Tarney, 1990; Silva Filho, 1976). Recent U-Pb dating of some of the igneous bodies give ages ranging from ca. 720 Ma to 620 Ma (Nascimento et al., 2005; Oliveira et al., 2010), and, coupled with geochemical trends, suggest that the Canindé Domain represents a Wilson Cycle of ocean basin formation and then destruction (Oliveira et al., 2010).

Oliveira et al. (2015) dated granitic bodies from the Canindé, Poço Redondo-Marancó and Macururé Domains. They found two distinct age clusters, an early cluster between 630 and 618 Ma, preserved in all three domains, wherein the granitic bodies contain prominent mafic enclaves, and a later cluster emplaced between 590 and 570 Ma, present only on in the metasedimentary successions of the Macururé Domain. The early cluster constrains the latest onset of subduction to ca. 618 Ma and was interpreted to represent magma mingling from the melt of the underthrusted C-SF plate, with slab break off allowing for upwelling in the asthenosphere and partial melting of the slab, generating mafic fluids. The later cluster is thought to represent continent-continent collision between Borborema and São Francisco (Bueno et al., 2009; Oliveira et al., 2015), and is related to strike-slip granites generated further into the Borborema Block away from active margins (Bueno et al., 2009), suggesting that relative motion was finalised by ca. 570 Ma. Brito Neves et al. (2016) focussed on the northern margin of the Sergipano Belt and suggested that in this area between ca. 673 and 647 Ma, extension and basin formation ended with contractional orogenesis by ca. 630–600 Ma.

5.3. Zambezi-Lufilian-Damara Orogen

The Zambezi-Lufilian-Damara orogeny marks the closure of the Khomas Ocean and the collision of the Kalahari craton against C-SF. The Zambezi Belt passes east into the East African Orogen in Malawi and Mozambique (Fig. 2b). To the west, the belt is displaced by the large sinistral ca. 550–530 Ma Mwembeshi Shear Zone (Naydenov et al., 2014). West of this shear zone, the boundary between the C-SF and Kalahari continents is delineated by the broad curve of the Lufilian Arc, snaking through Zambia and the Democratic Republic of Congo, before linking up with the Central Damara Belt in Botswana and Namibia. Here, it meets a triple junction of Neoproterozoic oblique subduction, with the Gariep-Saldania Belt to the south along the coast of Namibia and South Africa, and the Ribeira-Kaoko Belt to the north, marking the coasts of Brazil and Namibia and tracing the RDLP-C-SF suture.

The Zambezi Belt preserves evidence of the formation of a significant ocean between the Kalahari and C-SF continents that was being subducted from at least ca. 670 to 600 Ma (John et al., 2003, 2004a). A gabbro-metagrabbro-eclogite assemblage preserves incompatible elemental patterns similar to that of MORB, and are, therefore, interpreted to represent subduction of oceanic crust (John et al., 2003). John et al. (2003) also calculated a low thermal gradient (~8 °C/km), and, coupled with the subduction depth, concluded that the oceanic crust would have to be at least 30 Myr old, suggesting that the oceanic domain between C-SF and Kalahari was larger than 1000 km. Collision between the Kalahari and the C-SF is constrained to between 545 and 525 Ma (Goscombe et al., 2000; John et al., 2004b) by more regional metamorphism through the Zambezi Belt. This is roughly coeval with the greenschist to lower amphibolite facies metamorphism evident in the Lufilian Belt (e.g. Rainaud et al., 2005), although peak orogeny in the Lufilian Belt is inferred to be between 560 and 530 Ma based on U-Pb from monazite grains (John et al., 2004b), magmatic intensity (Hanson, 2003) and crustal thickening exhibited by extensive folding, thrusting, and nappe development, peaking with the formation of whiteschists at ca. 520 Ma (John et al., 2004b; Rainaud et al., 2005). The Lufilian Belt hosts the late Tonian-Cryogenian Katanga Supergroup that itself hosts world-class Cu-Co deposits (El Desouky et al., 2010). Evidence for early continental deformation here is constrained to ca. 590 Ma preserved in metamorphic monazite from the Roan and Nguba Groups of the Chambishi Basin (Rainaud et al., 2005). Termination of convergence is estimated to be ca. 530–520 Ma due to successive ages from 510 Ma interpreted to reflect post orogenic cooling (John et al., 2004b; Rainaud et al., 2005). The tectonic evolution of the sinistral Mwembeshi Shear Zone (MSZ), separating the Zambezi and Lufilian Belts, is poorly known. The displacement was explained by variation in shortening between the Zambezi Belt, which is narrower and experienced a higher grade of metamorphism, compared to the Lufilian Belt, which is wider and records predominantly greenschist metamorphism, with the MSZ representing a reactivated fault during Gondwana amalgamation (Porada and Berhorst, 2000). Recent dating of the Hook Batholith by Naydenov et al. (2014), located just north of the MSZ in the Lufilian Belt, shows emplacement occurred between ca. 550 and 530 Ma based on U-Pb dating. The ages are syntectonic with an early (ca. 550 Ma) deformation event that resulted in E-W shortening associated with closure of the Mozambique Ocean. A slightly younger (ca. 520 Ma) N-S shortening event associated with Kalahari-C-SF collision is preserved in the Katanga Supergroup and the Hook Batholith, suggesting that the MSZ responded to accommodate oblique convergence (Naydenov et al., 2014).

Further west, the Damara Belt preserves a similar age range of deformation and convergence, although it is structurally more complex, and, geologically, considerably larger than the Lufilian and Zambezi Belts (see Frimmel et al., 2011; Gray et al., 2008 for overviews). Frimmel et al. (2011) described the sedimentation in the Damara Belt as occurring between the Kalahari Craton and the smaller Angola Craton (present-day southern Congo Craton) that they interpreted to have rifted off Kalahari earlier, but which acted separately from the larger C-SF craton further northwards until Gondwana amalgamation. Thus, in their model, while the Damara Belt records the convergence and closure of the ocean basin between Kalahari and Angola (and, by extension, Kalahari and C-SF), the extension and sedimentation that it preserves is not necessarily a record of C-SF rifting from Kalahari, but records a related, later rifting event. The Damara Belt is divided into zones of distinct tectonostratigraphy that are typically bounded by faults and/or lineaments, and, broadly, preserve the development of a Cryogenian-Ediacaran passive margin and its deformation into a fold belt (e.g. Frimmel et al., 2011; Hoffmann et al., 2004; Miller and Becker, 2008). Fold interference from the Gariep and Kaoko belts is prevalent throughout the northern units of the Damara Belt (e.g. Lehmann et al., 2016), and three main phases of deformation were identified; D1, early (ca. 620 Ma) N-S shortening from closure of the Khomas Ocean; D2, (ca. 580 Ma) E-W shortening due to closure of the Adamastor Ocean, and; D3, late (ca. 530 Ma) N-S shortening due to Kalahari-C-SF collision (Lehmann et al., 2016). Thick (>6000 m), relatively flat-lying strata preserved in grabens in the northern units, are comprised of dominant feldspathic sandstones with minor evaporates and some volcanic detritus. These are overlain by diamictites of the Chuos Formation and followed by thick platform carbonates until ca. 635 Ma, when units of the Ghuab glaciation (Hoffmann et al., 2004) were deposited, suggestive of a failed rift (rifting moved further south) and shallow sea, which preserved little N-S deformation from convergence (e.g. Frimmel et al., 2011). Further to the south, passed the Okahandja Lineament, these older sediments are not preserved, rather basal MORBs are observed (Matchless Amphibolite Group, Schmidt and Wedepohl, 1983) and are inferred to be younger than 635 Ma (no evidence of glaciation). The turbiditic and schistose units of the Kuiseb Formation overlie them, and the southern zones are interpreted to preserve short-lived sea floor spreading. Closure of the basin occurred from 600 Ma, somewhat synchronous with the Lufilian and Zambezi belts, although the primary phase of collision was between 560 and 530 Ma as demonstrated by prominent magmatism and granitoid emplacement. Higher metamorphic grades of deformation due to N-S convergence are primarily preserved in the central and southern units (as opposed to northern units), with higher metamorphic grades and fold interference more common towards the west (Hartmann et al., 1983; Lehmann et al., 2016). There are few temporal constraints on early convergence related deformation. Miller (1983) proposed that early deformation began just prior to 650 Ma based on Rb-Sr whole rock age dating on granitoids that intruded D1 folds. The southern zones, which are reminiscent of an accretionary wedge, underwent medium pressure-temperature metamorphism (Kasch, 1983), while 580 to 520 Ma magmatism coupled with crustal thickening leading to low-granulite facies metamorphism occurred in the central zones (Jung and Mezger, 2003; Jung et al., 2000; Longridge, 2012; Longridge et al., 2014; Ward et al., 2008). Similar to the Lufilian and Zambezi belts, the Damara Belt was intruded by post-collisional granitoids from 500 to 480 Ma (e.g. Jung et al., 2000).

5.4. Pinjarra-Prydz-Denman (Kuunga) Orogen

The Pinjarra-Prydz-Denman orogeny is the final major amalgamation of continental crust into Gondwana, with the suturing of Australia-East Antarctica against India and Kalahari. The Pinjarra orogeny typically refers to the entire orogen, although here we separate the three to discretely talk about varying tectonic events. The Pinjarra orogeny preserves the suture along the west coast of Australia. Further south, in Antarctica, snow and ice cover limit most exposure, although the suture is preserved in outcrops in the Denman glacier area, and, further south, in the Prince Charles Mountains-Prydz Bay area, where India and the Rayner province collided with the main crustal part of Antarctica (e.g. Boger, 2011) (Fig. 2b).

Exposure of the Pinjarra orogeny in Australia is constrained to small inliers along the western coastline of the continent, such as the Leeuwin, Northampton and Mullingara Complexes. The Leeuwin Complex in the southwest best preserves the orogeny (e.g. Collins, 2003; Collins and Fitzsimons, 2001). Here pink granitic gneisses had their protoliths emplaced at ca. 750 Ma and exhibit upper amphibolite-granulite metamorphism dated to ca. 522 Ma (Collins, 2003). This is broadly coeval with the end of tectonism, as ca. 520 Ma dykes that intrude the Leeuwin Complex exhibit no deformation (e.g. Fitzsimons, 2003). The tectonic environment of emplacement was inferred to be a rift, related to Rodinia breakup (Collins, 2003), since at the time it was postulated that Kalahari was attached to this margin of Australia (e.g. Powell and Pisarevsky, 2002). Sinistral shearing is preserved in the Northampton Complex (Embleton and Schmidt, 1985), and alkali granitoids in the Leeuwin Complex originally inferred to be rift related, are now thought to have been emplaced in a sinistral transpressive environment (e.g. Fitzsimons, 2003; Harris, 1994). Further evidence of this motion is preserved in the Darling Fault along the coastline of Western Australia, where structures from the Albany Fraser Orogen were sinistrally rotated nearly 90° (Beeson et al., 1995; Fitzsimons, 2003). Dating of dykes affected by the shear suggest that the motion was occurring at least between 600 and 550 Ma, although an earlier initiation is inferred by reconstructions (e.g. Fitzsimons, 2003; Powell and Pisarevsky, 2002). Further south, the Denman Glacier area is on the coast of Antarctica fitting tightly with the Leeuwin Complex in a reconstructed Gondwana. Here, U-Pb dating of zircons from syenite give an age of ca. 516 Ma, and orthogneisses with a protolith age of ca. 3 Ga record a metamorphic overprint age between 550 and 520 Ma (e.g. Halpin et al., 2008). Some data show substantial lead loss between 600 and 520 Ma (Black et al., 1992), indicating a similar Ediacaran history to rocks further north in Australia. The suture between India-Antarctica and Australia-Antarctica is typically traced from this area, south, towards Prydz Bay and the Prince Charles Mountains (e.g. Boger et al., 2001).

The Prydz Bay area is found further south in Antarctica and preserves the primary suture between India-Antarctica and Australia-Antarctica, since it is found slightly further away from a reconstructed India than the late Mesoproterozoic-early Neoproterozoic Rayner Province. Due to the similarity of protoliths, neodymium modelling ages and metamorphic events, the Prydz Bay area is inferred to be part of the Indo-Antarctica plate (e.g. Boger, 2011; Kelsey et al., 2007; Liu et al., 2009; Wang et al., 2008; Zhao et al., 1995). Here too, late Ediacaran-early Cambrian metamorphism up to granulite facies (Liu et al., 2003) is evident, with 540–500 Ma charnockite and granite plutons intruding gneisses (Liu et al., 2006, 2009; Mikhalsky and Sheraton, 2011). Further inland from Prydz Bay (Antarctica), zircons from gneiss within the Grove Mountains suggest magmatic emplacement at ca. 900 Ma, with a high-grade metamorphic overprint between 530 and 520 Ma (Liu et al., 2003; Zhao et al., 2000). Intruding ca. 500 Ma granitic dykes exhibit no metamorphism suggesting deformation had finished by this time (Zhao et al., 2000).

5.5. Gariep-Dom Feliciano-Kaoko Orogen

The Gariep, Dom Feliciano and Kaoko belts preserve the suture between C-SF, RDLP and Kalahari, and form two arms of the triple junction between the cratons, with the Damara Belt, between C-SF and Kalahari, being the third arm (Fig. 2b). The Kaoko belt of the western Congo Craton traces the suture through northwest Namibia towards the junction with the Damara Belt in central Namibia, it is extended further to the south along the southwestern Namibia border and western South Africa as the Gariep Belt. The suture is preserved on the South American side of Gondwana through the Dom Feliciano Belt, which extends from southern Brazil in the north, through Uruguay to the south, along the eastern margin of the RDLP craton, where it is juxtaposed against by the Neoproterozoic-aged, sinistral Sarandí del Yi megashear.

The Kaoko Belt is divided into three distinct tectonostratigraphic units (Miller, 1983); the Eastern Kaoko Zone is a late Tonian(?) to Cryogenian carbonate platform, containing some deeper water argillite sequences (Miller, 2008), and is broadly coeval to the northern zone in the Damara Orogen (e.g. Hoffman et al., 1998b). The Central Kaoko Zone consists of deformed and metamorphosed Archaean and Palaeoproterozoic basement, with metamorphic grade increasing to the west, from greenschist to upper-amphibolite facies. The Western Kaoko Zone (WKZ) was subjected to an even higher grade of metamorphism, up to granulite facies, and consists predominantly of metasedimentary rocks with some Neoproterozoic granitoids. A high temperature metamorphic event from 650 to 630 Ma in the WKZ not preserved elsewhere in the Damara region, suggests that the WKZ contains an exotic terrane (the Coastal Terrane, Goscombe et al., 2003a), linked to arc accretion over an east dipping subduction zone until ca. 600 Ma (Goscombe and Gray, 2008; Goscombe et al., 2003a,b; Gray et al., 2008; Masberg et al., 2005). This is roughly synchronous with the collision of the Oriental Terrane further north in the Brazilian Ribeira belt (e.g. Heilbron and Machado, 2003). Younger metamorphic events, between 580 and 550 Ma, which are preserved in the entire belt and get progressively weaker towards the east, are interpreted to be related to sinistral, transpressional deformation from oblique collision between the South American cratons and the C-SF craton. Peak metamorphism was reached early in the main metamorphic cycle (ca. 580–570 Ma, Goscombe et al., 2003a), with granitoid intrusions and magmatism occurring until ca. 550 Ma. Minor N-S shortening occurred between 530 and 510 Ma, likely related to the arrival of either (or both) RDLP and Kalahari against the C-SF Craton (Goscombe et al., 2003b; Gray et al., 2008).

The Gariep Belt lacks the higher grades of metamorphism prevalent in the African belts further north and, consequently, it better preserves Neoproterozoic sedimentary sequences from the Adamastor Ocean.

Two distinct units are recognised, the folded and thrusted(?) volcano-sedimentary Port Nolloth Group in the east, and the younger, oceanic Marmora Terrane. The Marmora Terrane is comprised of oceanic crust with minor overlaying carbonates with an approximate minimum age of ca. 600 Ma based on U-Pb dating of zircon in a stromatolite (Frimmel and Fölling, 2004; Frimmel et al., 2002). The young age of oceanic crust indicates some seafloor spreading occurred until relatively late in the Adamastor Ocean. High-pressure metamorphism is preserved in the Marmora Terrane at ca. 575 Ma and peak metamorphism inferred to occur between ca. 550 and 540 Ma with continent-continent collision and the thrusting of the Marmora Terrane onto the Port Nolloth Group (Frimmel and Frank, 1998). Ages from muscovite in the Port Nolloth group at ca. 530–525 Ma suggest post-tectonic cooling and exhumation at this time.

The Dom Feliciano Belt (DFB) of Uruguay preserves an older record of subduction than either the Gariep or Kaoko belts and presents a more complicated story that has invited alternate interpretations that have profound implications on Neoproterozoic palaeogeography. The DFB runs from the southeast corner of Brazil to the south of Uruguay, parallel to the Atlantic coastline. Unlike the Gariep and Kaoko belts, which consist predominantly of folded sediments and oceanic terranes from basin inversion, the Dom Feliciano Belt preserves fragments of Palaeoproterozoic continental crust, Neoproterozoic arcs and sedimentary wedges, forced together under a transpressional tectonic regime (Basei et al., 2000). Three distinct domains (or belts) were identified (Basei et al., 2000), the Western Domain consists of ca. 750 Ma juvenile arc rocks, followed by ca. 730 Ma mafic rocks associated with ophiolitic assemblages (serpentine, peridotite) and volcanoclastic sedimentary successions (Leite et al., 1998). The Central Domain in Brazil contains Palaeoproterozoic gneisses overlain by lower grade Neoproterozoic supracrustal rocks, and is represented further south in Uruguay by a schistose volcanosedimentary fold/thrust belt prevalent in the Lavalleja Complex and Nico Perez Terrane (Sanchez Bettucci et al., 2010). Both the Western and Central Domains crop out mostly in the north (southern Brazil); towards the south they narrow between the RDLP craton and the rocks of the Eastern Domain. The Eastern Domain (or Granite belt) comprises a long orogenic belt of syn-post tectonic calc-alkaline granitic batholiths that were emplaced in the south between 650 and 550 Ma (Basei et al., 2000; Hartmann et al., 2002; Oyhantçabal et al., 2007, 2009). Towards the south in Uruguay, the Eastern Domain crops out between the older rocks of the RDLP craton and central/schist belt, with sedimentation here leading to the development of a foreland basin (e.g. Oyhantçabal et al., 2009). The foreland belt (acting as the southerly extent of the Central Domain in the north), is bounded by a reworked Archaean and Palaeoproterozoic terrane (Nico Perez Terrane), and overlying volcano-sedimentary units (Basei et al., 2000; Hartmann et al., 2001). Various terranes within the Eastern Domain are proposed to be linked with both the Gariep Belt (e.g. Basei et al., 2005), and the Kaoko Belt (e.g. Oyhantçabal et al., 2009).

The northern rocks of the DFB preserve older magmatic signatures, with continental arc magmatism preserved from the early Cryogenian (e.g. Koester et al., 2016; Lenz et al., 2013; Saalmann et al., 2006). In the south, magmatic age decreases from ca. 650 to 600 Ma with the emplacement of the granitoid bodies; although recent studies have found small areas of earlier magmatism do exist from 800 to 770 Ma, with a metamorphic overprint at ca. 650 Ma (Lenz et al., 2013). Similar to the Kaoko and Gariep belts, transpressional deformation is prevalent from ca. 580 Ma (e.g. Oyhantçabal et al., 2011). The DFB is thought to have not developed on the RDLP craton; firstly, because of the shear and thrust relationships between them, secondly, because of the allochthonous older terranes in the Central Domain, and, finally, because RDLP's position in Rodinia does not allow long lived subduction on its northern and eastern margins (e.g. Fuck et al., 2008; Gaucher et al., 2011). The model of Rapela et al. (2011) suggests that a ca. 800 Ma rifting event between the Kalahari, Nico Perez Terrane and the Angola Block (part of Congo) was responsible for the opening of the Adamastor Ocean and building of the DFB, and that sinistral shearing occurred when the RDLP collided with Kalahari, squeezing the fold belts.

5.6. Ribeira-Araçuaí-West Congo Orogen

The Araçuaí belt in South America (and its African equivalent, the West Congo belt) formed from the closure of the intra-cratonic ocean basin between the São Francisco and Congo Cratons, and is connected to the more southerly Damara-Dom Feliciano-Kaoko belts through the Ribeira Belt, which lies along the southern tip of the São Francisco craton (Pedrosa-Soares et al., 2001) (Fig. 2b). The Ribeira Belt contains a number of tectonostratigraphic units (e.g. Heilbron et al., 2008), preserving Archaean-Palaeoproterozoic basement (e.g. Occidental Terrane, Trouw et al., 2000), early Neoproterozoic rift histories (e.g. Andrelândia Basin, Paciullo et al., 2000; Valladares et al., 2004) and Cryogenian to Cambrian subduction and magmatism (e.g. Oriental Terrane, Heilbron and Machado, 2003; Heilbron et al., 2010). MORB-like magmatism in ca. 848 Ma intrusions in these terranes suggests that there was an ocean basin between them and the São Francisco Craton during the Tonian (Heilbron and Machado, 2003). The Oriental Terrane of the southern São Francisco Craton preserves the oldest evidence of subduction, with early arc complexes recording a ca. 790 Ma U-Pb age in zircon and monazite from tonalitic gneiss (Heilbron et al., 2003). The primary phrase of subduction in the Ribeira Belt is inferred to have occurred from 650 to 620 Ma, with the collision of the Rio Negro Arc and the Oriental Terrane (Heilbron and Machado, 2003; Heilbron et al., 2008; Tupinambá et al., 2012). This collision is roughly synchronous with the collision of the Coastal Terrane in the Kaoko Belt further south, suggesting a long-lived semi-continuous arc between São Francisco and the Congo (or Angola?) Craton. Closure of the ocean basin and continental collision occurred after 620 Ma, with collision of the Oriental Terrane and São Francisco Craton occurring between 580 and 550 Ma (Heilbron and Machado, 2003). Heilbron and Machado (2003) inferred a small landlocked sea between Congo and São Francisco in the south that did not close until the earliest Ordovician (Schmitt et al., 2008; Fernandes et al., 2015).

Further north, the Araçuaí Belt preserves the closure of an ocean basin that is surrounded on three margins by cratonic crust, leading to a tectonically complex kinematic evolution (e.g. Maurin, 1993; Pedrosa-Soares et al., 1992), where 'nutcracker' tectonics are inferred as one of the drivers of subduction (rather than slab pull) (Alkmim et al., 2006). Here, the earlier magmatism of the Rio Negro Arc is not preserved, and five generations of granite emplacement are recognised (Pedrosa-Soares and Wiedemann-Leonardos, 2000). The earliest generation of granites are calc-alkaline and dated between 630 and 590 Ma, indicating the start of subduction (da Silva et al., 2005; Pedrosa-Soares et al., 2001; Tupinamba, 1999). Closure and development of a thick orogen occurred between 590 and 570 Ma, demonstrated by a second generation of S-type granitoids, and peak metamorphic conditions and deformation to granulite facies (e.g. da Silva et al., 2003, 2005; Pedrosa-Soares et al., 2001). Post-collisional escape tectonics occurred to the south, towards the Kaoko and Dom Feliciano belts, and is coupled with the third generation of granite emplacement between 560 and 510 Ma (Alkmim et al., 2006). This closure is described as having occurred because the approaching Amazonia-WAC cratons pushed against the São Francisco, forcing convergence between itself and the Congo Craton, which could help explain sinistral displacement further south in the Kaoko and Dom Feliciano Belt between RDLP and Kalahari. The final two generations of granite intrusions relate to orogenic collapse and were emplaced between 510 and 480 Ma (Pedrosa-Soares and Wiedemann-Leonardos, 2000).

5.7. Brasília Orogen

The Brasília Belt forms the eastern part of the primary suture between Amazonia and C-SF, the two largest areas of Precambrian crust in South America and Africa, and is preserved on the western margin of the São Francisco Craton (Fig. 2b). It forms part of the Brasiliano Orogen, along with the Araguaia, Paraguay, Dahomeyide and Pharuside belts that stitch the WAC and Amazonia to the C-SF and the SM. The Brasília Belt contains a number of distinct zones, from east (SF) to west (Amazonia), including a foreland basin sitting unconformably above the Archaean-Palaeoproterozoic basement of the São Francisco craton and an external zone of metasedimentary rocks deposited in a shelf environment comprising both silicic and carbonate successions (Dardenne, 2000; Paciullo et al., 2000; Pimentel et al., 2011). Further west, an internal zone is preserved, with W-NW verging nappes and pelitic metasedimentary rocks, suggesting deeper shelf sedimentation, although further to the north along the zone, rocks derived from an oceanic environment become more prominent, perhaps suggesting an open ocean to the N-NW (Seer and Dardenne, 2000; Seer et al., 2001; Valeriano et al., 2008). Westerly subduction is inferred to have occurred under the Neoarchaean granite-greenstone Goiás Massif between 650 and 610 Ma (e.g. Campos Neto and Caby, 1999; Töpfner, 1996; Valeriano et al., 2004), acting here as a microcontinent (Queiroz et al., 2008; Valeriano et al., 2008) accreted to the Brasília Belt sometime during the Neoproterozoic. The southerly extension of the Goiás Massif is typically inferred to be the Paranapanema Block (e.g. Valeriano et al., 2008). The Goiás magmatic arc is preserved both north and south of the Goiás Massif, and records magmatism from the Tonian, with juvenile magmatism and intrusions of tonalite-granodiorite from 860 to 610 Ma, with the younger rocks becoming more evolved (Matteini et al., 2010; Pimentel et al., 2000, 2004, 2011). Peak metamorphic grade is upper amphibolite to lower granulite facies and is constrained to between 650 and 630 Ma (e.g. Moraes et al., 2002; Piuzana et al., 2003; Seer et al., 2005; Valeriano et al., 2004), with post tectonic cooling from 610 to 600 Ma, suggesting collision between the Goiás magmatic arc, Goiás Massif and São Francisco occurred prior to the opening of the Iapetus Ocean.

5.8. Paraguay-Araguaia-Rokelides-Bassarides Orogen

The Paraguay and Araguaia belts, south and north of the Brasília Belt, respectively, preserve the final suture of Amazonia and C-SF, although on the Amazonian side of the margin rather than the São Francisco side (Cordani et al., 2003; Moura et al., 2008) (Fig. 2b). The Rokelide and Bassaride belts lie further north, in present-day Africa, between Amazonia and the WAC in a reconstructed Gondwana.

The Paraguay belt consists of a northern and southern section separated by Phanerozoic cover. Traditionally the belt was interpreted as an inverted rift, as metasedimentary and sedimentary rocks dominate it with the only granitic bodies being from post-collisional tectonism (e.g. Alvarenga et al., 2009; McGee et al., 2012b). Broadly, the lower units in the northern Paraguay Belt are interpreted to represent passive margin sedimentation and are dominated by a thick diamictite layer, overlain by a foreland succession of carbonate and sandstone layers, with a thinner diamictite layer (Alvarenga et al., 2008, 2009; McGee et al., 2013). The lower diamictite has a carbonate cap dated to ca. 630 Ma, suggesting it is part of the Marinoan glacial event (Alvarenga et al., 2004), while the higher, thinner layer, is stratigraphically correlated to the 580 Ma Gaskiers glaciation (Alvarenga et al., 2007; Pu et al., 2016). U-Pb dated detrital zircon provides a maximum depositional age of ca. 706 Ma (Babinski et al., 2013; McGee et al., 2015a,b). Furthermore, Sm-Nd modelling ages of the sediments in the basin, U-Pb detrital zircons and ^{40}Ar-^{39}Ar detrital muscovite ages suggest that older sediments were sourced from the Amazonian Craton due to the predominance of Archaean and Palaeoproterozoic signatures, while the younger, higher sediments have a younger component (Dantas et al., 2009; McGee et al., 2015a,b). Dantas et al. (2009) and McGee et al. (2015a,b) suggested that the Goiás Magmatic Arc in the Brasília Belt was a likely source, with sediment derived from this belt as the intervening Clymene Ocean closed in the late Ediacaran/Cambrian (Tohver et al., 2010; Bandeira et al., 2012). In the Southern Paraguay Belt the glacial events are more poorly preserved, and the stratigraphy differs broadly, being comprised predominantly of siliciclastic rocks (Alvarenga et al., 2009). A transition from rift to drift is suggested in the Corumbá Group, although age controls are poor, with only one of the drift successions having an age constraint of ca. 540 Ma from zircon in a volcanic ash, suggesting that drifting happened sometime in the late Ediacaran (Alvarenga et al., 2009; Babinski et al., 2008). The connection between the southern and northern Paraguay belts is poorly understood due to overlaying Phanerozoic basins. No evidence of oceanic crust exists in the Paraguay Belt, and deformation is constrained to 550–520 Ma as one of the latest events in the Brasiliano Orogeny, with peak metamorphism only reaching greenschist facies (e.g. Pimentel et al., 1996; Trompette, 1994). A minimum age for deformation and metamorphism is ca. 520 Ma based on undeformed, post tectonic granite emplacement (McGee et al., 2012b).

The Araguaia Belt separates rocks deformed during Gondwana amalgamation to the east with the undeformed rocks of the Amazonian Craton to the west and is interpreted to represent the northerly extension of the Paraguay Belt, although the two have quite a different stratigraphy and geology. Rocks increase in metamorphic grade to the amphibolite facies (Abreu et al., 1994) in the east, and are comprised predominantly of sedimentary successions thrust westwards over Amazonia. Protoliths are mainly siliciclastic rocks, with some minor carbonates and mafic-ultramafic rocks, and are classified into the Baixa Supergroup (Alvarenga et al., 2000). The western portion of the belt is comprised mostly of schists with minor quartzite and phyllite, grading towards phyllite and slate, with minor quartzite (e.g. Alvarenga et al., 2000). Mafic-ultramafic rocks (serpentinised peridotites) are also evident here, and are described as metamorphosed ophiolites dated to ca. 757 Ma (Paixão et al., 2008). Comparably, the eastern portion contains schist, quartzite and meta-conglomerates, without any mafic-ultramafic bodies (Alvarenga et al., 2000). Sm-Nd model ages of zircon taken from quartzite units of the Baxio Supergroup do not support provenance from only Amazonia, as younger signatures are preserved as well as older Archaean-Palaeoproterozoic signatures (Moura et al., 2008). Instead, Moura et al. (2008) proposed that a source of detritus for the Araguaia basin was from the Goiás Magmatic Arc, Goiás Massif and Brasília Belt. Early deformation is possibly slightly older than in the Paraguay Belt, with zircon from granitoids dated at ca. 650 Ma (Moura and Gaudette, 1993), although these zircons may be inherited, as a range of zircon ages are also recorded (Teixeira et al., 2002). A recent age of ca. 550 Ma was found by Alves (2006) and is now thought to be a better representation of the timing of the collision.

The Araguaia Belt is bound to the more easterly Brasília Belt by the cryptic Transbrasiliano Lineament, which separates pre-Neoproterozoic western Gondwana crust into Congo-affinity (e.g. São Francisco, Paraná, Borborema, SM) or Amazonia-affinity (e.g. WAC). The Araguaia Belt is typically extended into northern Africa along the 4°50′ or Kandi fault (e.g. Caby, 1989; Cordani et al., 2013; de Wit et al., 2008; Ganade de Araujo et al., 2016). This lineament is prominent as a sub-vertical continental-scale magnetic discontinuity (e.g. Curto et al., 2014; Fairhead and Maus, 2003), and is interpreted as a dextral transform boundary, active sometime during the Ediacaran and Early Cambrian (e.g. Ganade de Araujo et al., 2014b; Ramos et al., 2010), although due to Palaeozoic basin cover, and the prominence of oblique subduction throughout the Brasiliano belts, it could represent a transpressive regime. Timing of the lineament is also poorly constrained; Ganade de Araujo et al. (2016) suggested ca. 590 Ma activation to match the Borborema-São Francisco collision along the Sergipano Orogen, whereas Ramos et al. (2010) preferred a late Ediacaran-Early Cambrian time to fit a Pampia-RDLP collision.

The Rokelide-Bassaride belts (from south to north) preserve a cryptic Neoproterozoic-Palaeozoic suture on the western and southwestern margin of the WAC, from Morocco in the north towards Liberia in the south (Fig. 2b). The belts are not synchronous with each other, but rather preserve separate events and indicate that Amazonia and the WAC did not act congruently throughout the Neoproterozoic, with the Bassaride Belt preserving an earlier ca. 650 Ma suture and the Rokelides a ca. 550 Ma suture (e.g. Villeneuve, 2008). The Bassaride Belt consists of two lower groups overlain by sedimentary sequences (Villeneuve, 1984). The basal lower group (Guinguan Group) consists of rocks metamorphosed to greenschist-amphibolite facies including amphibolites, schists, metamorphosed basalts and dolerites, and serpentinites. The second group (Niokolo Koba Group), thrusted over the Guinguan Group, consists of granitoid batholiths and lava flows, including a basaltic-andesitic series and rhyolitic series (Villeneuve, 1984). Rb-Sr dating of the batholiths gave an age of 683–660 Ma (Bassot and Vachette, 1983; Villeneuve, 2008), with metamorphism of the Guinguan group thought to be ca. 660 Ma based on the younger ages of these intrusions. High K_2O content suggests that the collision was continent-continent (Villeneuve et al., 1991). The lavas of the Niokolo Koba Group are inferred to be younger (Villeneuve, 2008), and the volcanosedimentary and sedimentary successions that overlie the Niokolo Koba Group contain some Cambrian microfossils (Culver et al., 1996), although an age date of ca. 600 Ma was suggested by Deynoux et al. (2006) based on stratigraphic correlation. Here alkali basalts that preserve MORB affinity are also present, suggesting that Amazonia and the WAC were not fellow travellers for all the Neoproterozoic (Villeneuve, 1984).

The Rokelide Belt, lying slightly further south than the Bassaride Belt, represents the Amazonia-WAC suture preserved in Gondwana through to the Mesozoic opening of the Atlantic Ocean. It is broadly similar to the Bassaride Belt but exhibits a higher metamorphic grade, with granulitic gneisses and schists prevalent within the thrust belt (Villeneuve, 1984). Three stages of metamorphism were identified, an early peak event at granulite facies, followed by a second amphibolite facies event and later retrogression, with post tectonic granitoids dated at ca. 530 Ma using Rb-Sr and ^{40}Ar-^{39}Ar (Dallmeyer et al., 1987). Further dating on syntectonic granitoids and gneisses suggest an age of 570–550 Ma for peak metamorphism and continent-continent collision (Delor et al., 2002), inferred to be between Amazonia and WAC.

5.9. Dahomeyide-Pharuside Orogen

The Dahomeyide and Pharuside belts are the northerly extension of the Brasiliano Belt and record the suture of the WAC to the Borborema Block in the south (Dahomeyides), the accretion of the westerly terranes and shields (e.g. Hoggar, LATEA, Tuareg Shield etc.) and the WAC to the SM in the north (Pharusides) (Fig. 2b). Few data are available for either belt due to lack of exposure and access. The Dahomeyides expose the suture through central western Africa, predominantly in Togo, Benin, Nigeria, but also in Ghana and Cameroon. Here, the passive margin of the WAC collided with the Nigerian Shield (inferred to act as the northern extension of the Borborema Block), creating a thick fold belt, with subduction to the east, under the Nigerian Shield, away from the WAC (Santos et al., 2008b). The belt consists of a western external zone (foreland of WAC?) consisting of units of undeformed sedimentary successions in the west and quartzite, schist and gneiss towards the east that crop out as nappes (Attoh et al., 1997; Castaing et al., 1993, 1994). The suture zone itself is more heavily metamorphosed, consisting of rocks deformed to granulite and eclogite facies (e.g. Attoh et al., 1997), and zircon from gneiss in the suture zone record a U-Pb age of ca. 610 Ma, interpreted to represent peak metamorphism (Attoh et al., 2013), with later ^{40}Ar-^{39}Ar muscovite data recording an exhumation age of ca. 580 Ma (Attoh et al., 1997).

Further north, the Pharuside Belt has a more extensive record preserved, as ongoing subduction and accretion within the Hoggar Shield, off the western margin of the SM, continued throughout most of the Neoproterozoic (e.g. Caby, 2003). Earlier accretion is attributed to the organising/accretion of the terranes preserved in the Tuareg and Hoggar Shield, with further accretion and more intense deformation more prevalent in the Ediacaran, as a response to the closing of the Pharusian Ocean and the incipient collision with the approaching WAC (e.g. Caby, 2003; Liégeois et al., 1994). Calc-alkaline granitoids preserve magmatic ages between 700 and 520 Ma (see Caby, 2003 for an overview). The 700 to 650 Ma ages, are typically associated with intrusions into the westernmost terranes of the Hoggar Shield, while the later magmatism (ca. 620 Ma) is associated with HP eclogite formation after rapid (~3 Myr) exhumation (e.g. Caby and Monié, 2003; Berger et al., 2014). A transition to a transpressive tectonic regime is inferred to occur after this event (Berger et al., 2014), and to last until at least 592 Ma, where U-Pb dated dykes were affected by the event (Hadj-Kaddour et al., 1998). Later dextral motion was also suggested to have occurred at ca. 530–520 Ma (Paquette et al., 1998).

5.10. Gondwanides

The Gondwanides (or Terra Australis Orogen after Cawood, 2005) are a series of accreted terranes along the southern margin of Gondwana that formed due to long-lived subduction after Gondwana amalgamated. They initiated in the Cambrian and at their maximum length, stretched from Australia (Delamerian orogeny; Foden et al., 2006) through Antarctica (Ross orogeny), across South Africa (Saldania belt, Cape Fold belt) and into South America (Sierra del Ventana, Pampian orogeny, Patagonia fold belt) (Cawood, 2005) (Fig. 2b). The timing of subduction initiation is thought to have occurred from 540 to 500 Ma based on arc magmatism assemblages preserved in eastern Australia (Johnson et al., 2016; Foden et al., 2006), granitoids preserved in Antarctica (Vogel et al., 2002) and ophiolites in South America (Rapela et al., 1998), with a more mature subduction system developing by the Ordovician. Their initiation is typically interpreted as a consequence of Gondwana amalgamation, with interior subduction zones closing and repositioning to the exterior of the supercontinent, forming a circum-Gondwana subduction system.

5.11. Avalonia and Cadomia

The peri-Gondwanan terranes are a series of oceanic and continental arcs accreted to the north-eastern margin of Gondwana during the Ediacaran and Cambrian that later rifted off, closing the Iapetus Ocean and opening the Rheic Ocean (e.g. Murphy et al., 2004; Nance et al., 1991). These terranes were accreted to Baltica and Laurentia, and are currently preserved in southern and western Europe and eastern North America (e.g. Domeier, 2016; Mallard and Rogers, 1997; Murphy et al., 2004; Nance et al., 1991, 2008). Of particular interest to the Neoproterozoic are the Avalonian terranes of Laurentia and northern Europe, and the Cadomian terranes of southern Europe. Detrital zircon and muscovite data suggest that the Avalonian terranes evolved outboard of Amazonia (e.g. Nance et al., 2008; Gutiérrez-Alonso et al., 2005) (Fig. 2b), although recent studies of hafnium isotopes suggest that Baltica may have also provided some detritus (e.g. Henderson et al., 2016). The Avalonian terranes evolved on juvenile, or Mesoproterozoic recycled crust, with the earliest development of primitive oceanic arcs suggested to be at ca. 1.2–1.0 Ga based on Sm-Nd model ages (Murphy et al., 2000). The oldest preserved basement rocks consist of recycled felsic orthogneiss and variably deformed calc-alkaline plutons, dated between 750 and 675 Ma that formed above an active subduction zone (Murphy et al., 2000). Arc accretion to Amazonia (± West Africa) by ca. 650 Ma is suggested by medium- to high-grade metamorphism preserved throughout the terranes (e.g. Strachan et al., 1996, 2007), and was followed by more mature, 640 to 540 Ma subduction, reflecting the dominant period of Avalonian magmatism (Murphy et al., 2013; Nance et al., 1991). They consist predominantly of magmatic volcanic rocks, volcano-sedimentary turbidites and sedimentary successions attributed to a variety of tectonic environments including back-arc and intra-arc basins, and are inferred to have occurred under a regime of oblique subduction (Murphy et al., 1999; Murphy and Nance, 1989). A transition from a convergent tectonic setting to an intracontinental wrench setting is suggested to occur during the late Ediacaran (ca. 590–550 Ma), based on the synchronous development of transtensional basins and bimodal magmatism (e.g. Nance et al., 2008). This is inferred to represent the rifting of Avalonia from Gondwana, which is also indicated by faunal distributions (Landing, 2005). Ridge subduction (or ridge-trench interaction) was suggested to have occured during the latest Neoproterozoic due as well to a diachronous termination of subduction along the margin (e.g. Keppie et al., 2000; Nance et al., 2002).

The Cadomian terranes have exposed Palaeoproterozoic (ca. 2.1 Ga) basement, indicating that early magmatism in the Neoproterozoic occurred along a slither of continental crust (e.g. Samson and D'Lemos, 1998). The similarity in age of this basement to the WAC, coupled with detrital zircon data, suggest that the Cadomian terranes developed further east than the Avalonian terranes, outboard of WAC instead of Amazonia or Baltica (Fig. 2b). Their development during the Neoproterozoic is similar to that of the Avalonian terranes, with early magmatism from ca. 750 Ma preserved as the protolith ages of orthogneisses (Egal et al., 1996), and collision indicated by metamorphism and deformation occurring (slightly) later than in Avalonia, at ca. 620 Ma (Inglis et al., 2005; Samson and D'Lemos, 1999). Volcanoclastic sedimentary successions including turbidites (Egal et al., 1996) and detritus input also from the SM (Garfunkel, 2015) during the late Ediacaran, suggest some extension, likely due to the formation of a back-arc basin (Linnemann et al., 2008). Later magmatism in a transpressional tectonic regime due to oblique subduction was suggested to have occurred from ca. 540 Ma (Linnemann et al., 2008; Strachan et al., 1989).

6. Palaeomagnetic constraints

A summary of palaeomagnetic data used in the reconstruction is presented in Table 1. The abbreviations in this table refer to orthogonal reconstructions shown in Figs. 5–14 in Section 7.

6.1. Laurentia and Baltica

Palaeomagnetic data from Laurentia and Baltica are important to defining the motion of Rodinia for much of the early Neoproterozoic due to the central position that Laurentia occupies in the supercontinent. Unfortunately, high quality Early Neoproterozoic palaeomagnetic data from Laurentia are unavailable (Table 1) due to the majority of data coming from rocks that underwent intense metamorphism during the Grenvillian orogeny (e.g. Weil et al., 2006). The Grenvillian loop (the name ascribed to the early Neoproterozoic Apparent Polar Wander Path (APWP) of Laurentia) is based on a collection of poles from three distinct time periods, at 1100–1020 Ma indicating an equatorial position (e.g. Jacobsville Sandstone, Roy and Robertson, 1978), at ca. 980 Ma indicating high latitudes (e.g. Haliburton Intrusions A, Buchan and Dunlop, 1976; although these were re-dated to ca. 1015 Ma, Warnock et al., 2000) and at ca. 800 Ma suggesting a motion back towards the Equator (e.g. Galeros Formation, Weil et al., 2004). However, only the late Mesoproterozoic poles are well dated, leading to uncertainty as to whether the loop is clockwise (Hyodo and Dunlop, 1993) or counterclockwise (e.g. Weil et al., 1998, 2006). Comparably, the Sveconorwegian APWP from Baltica consists of poorly age-constrained poles during the latest Mesoproterozoic (1100–1000 Ma, e.g. Elming et al., 2014 and references therein), but, the early Neoproterozoic poles (ca. 950–850 Ma, Table 1) are well dated. Using palaeomagnetic data from Baltica to supplement the Laurentian data suggests a clockwise motion, but also suggests a decoupling of the Grenvillian loop from the Sveconorwegian loop due to a ~150 Myr age difference in the poles constraining their highest latitude positions (Elming et al., 2014). This age disparity is based on the suggestion that the high latitude Baltican poles from 920 to 900 Ma were remagnetised (Walderhaug et al., 2007). In spite of this discrepancy in APWPs, a movement from low-to-high latitudes and back again, between ca. 940 Ma and 800 Ma, with the peak occurring at 870 Ma, is implemented following Elming et al. (2014).

There is a converse situation for the time interval between 800 and 700 Ma, where there are several reliable Laurentian poles, but no poles for Baltica. Harlan et al. (2008) combined several palaeomagnetic studies on the ca. 780 Ma Gunbarrel intrusions, and related rocks, into one reliable pole (Table 1). Weil et al. (2004, 2006) reported a well-dated pole from the Kwagunt Formation and the imprecisely dated, but similar pole from the Uinta Formation (Table 1, Fig. 10b, c). There are many palaeomagnetic studies on the Franklin magmatic province, with the most recent, highly reliable pole compiled by Denyszyn et al. (2009) (Table 1). All these poles constrain Laurentia's position to low latitudes during the mid-late Cryogenian. There is a plethora of Laurentian poles for the Ediacaran and Early Cambrian (e.g. Lubnina et al., 2014). Ediacaran poles from Laurentia have a convoluted story, with models proposing either a high and low latitude Laurentia between ca. 620 and 540 Ma based on palaeomagnetic data (e.g. Cawood and Pisarevsky, 2006; Collins and Pisarevsky, 2005; Li et al., 2008, 2013; Pisarevsky et al., 2008b). Well dated ca. 615 Ma (Long Range Dykes, Murthy et al., 1992) and ca. 550 Ma (Skinner Cove Formation, McCausland and Hodych, 1998) poles keep Laurentia at low latitudes, although a series of poles between 600 and 560 Ma place Laurentia at either a high, or low latitude position (e.g. the 575 Ma Callander Complex, McCausland et al., 2011; Symons and Chiasson, 1991). Unreasonably high rates of motion or Inertial Interchange True Polar Wander (IITPW) are some hypotheses used to account for the palaeomagnetic poles, although Hodych et al. (2004) notes that the 600 to 560 Ma poles simply permit Laurentia to be at a high latitude, rather than require a high latitude position. Halls et al. (2015) reported a new detailed study on the Laurentian Ediacaran mafic intrusions. They demonstrated that highly reliable palaeomagnetic data from two coeval well-dated dykes are very different, with one (steep remanence) supporting the high-latitude position of Laurentia, while the other (shallow remanence) supported a low-latitude position. Even IITPW cannot explain such high speeds of apparent plate motion. In addition, Halls et al. (2015) found that 90% of altered dykes carry the steep remanence direction, while 75% of petrographically fresh dykes do not. Halls et al. (2015) suggested that the “duality” in Laurentian Ediacaran poles were caused by a combination of remagnetisation and high reversal frequency at that time. Hence, here we have elected to keep Laurentia at low latitudes, omitting the majority of Ediacaran palaeomagnetic data from Laurentia. A somewhat similar shift from high to low latitudes is also observed in palaeomagnetic poles from Baltica (e.g. Lubnina et al., 2014; Klein et al., 2015), although an absence of 600 to 580 Ma Baltican poles coupled with geological evidence indicating that Baltica and Laurentia had separated by this point (e.g. Cawood et al., 2007), make it difficult to determine whether the same disparity is evident in Baltican poles.

The reliable ca. 615 Ma pole, from the Egersund dykes (Walderhaug et al., 2007, Table 1) constrains Baltica to a mid-latitude position (with Laurentia still equatorial). This pole overlaps with the ca. 615 Ma Long Range Dykes pole from Laurentia (Table 1), supporting a Neoproterozoic connection between the two, and constraining rifting to after 615 Ma. A series of 570 to 550 Ma late Ediacaran Baltic poles (Iglesia Llanos et al., 2005; Levashova et al., 2013; Lubnina et al., 2014; Popov et al., 2002, 2005; Table 1) suggest a mid-high latitude Baltica, rotating approximately ~90° counter-clockwise from its position against Laurentia to be in mid latitudes at 570 Ma and low-mid latitudes by ca. 550 Ma (Lubnina et al., 2014; Meert, 2014b). The large rotation can (perhaps in part) be accounted for by the presence of a triple junction between Amazonia-Baltica-Laurentia coupled with the onset of subduction in the Timan area (creating a ‘pulling’ force for Baltica), which would explain its movement towards the equator.

In lieu of the absence of reliable Cambrian poles from Baltica, we follow the reconstruction of Lubnina et al. (2014) for Baltica from 600 to 550 Ma, and then fit Baltica at 520 Ma in a position similar to that in the 2009 Atlas of Plate Reconstructions of Lowver et al. (http://www-

Table 1
List of palaeomagnetic data used in the reconstruction. Q-factor after Van der Voo (1990). Key is used for identification of poles in reconstruction Figs. 5–14.

Key	Rock unit	Age	Plat	Plong	A95	Author, year	Q-factor
Laurentia							
L1	Gunbarrel Intrusions combined	780–776	9.2	138.7	9	Harlan et al., 2008	1110111 (6)
L2	Uinta Formation	800–750	0.8	161.3	4.7	Weil et al., 2006	0110111 (5)
L3	Kwagunt Formation	748–736	18.2	166	7	Weil et al., 2004	1111111 (7)
L4	Franklin Dykes	727–712	8.4	163.8	2.8	Denyszyn et al., 2009	1111111 (7)
L5	Long Range Dykes	617–613	−19	175.3	17.4	Murthy et al., 1992; Hodych et al., 2004 Age: Kamo and Gower, 1994	1011111 (6)
L6	Skinner Cove Formation	554–548	15	157	9	McCausland and Hodych, 1998	1111001 (5)
Baltica							
B1	Southern Sweden Dykes	946–935	−0.9	240.7	6.7	Elming et al., 2014; Pisarevsky and Bylund, 2006	1111111 (7)
B2	Brantön-Älgö Anortthosite	927–905	5	249	3.9	Stearn and Piper, 1984; age: Schersten et al., 2000	1110101 (5)
B3	Rogaland Igneous Complex	883–855	−46	238	18.1	Walderhaug et al., 2007	1010101 (4)
B4	Hunnedalen Dykes	875–821	−41	222	10	Walderhaug et al., 1999	1110101 (5)
B5	Egersund Dykes	619–613	−31.4	224.1	15.6	Walderhaug et al., 2007	1111101 (6)
B6	Kurgashlya Formation	570–560	−51	135	4.9	Lubnina et al., 2014	1110111 (6)
B7	Bakeevo Formation	570–560	−42	119	5.3	Lubnina et al., 2014	1111111 (7)
B8	Winter coast sediments	558–552	−25.3	132.5	2.8	Popov et al., 2002; age: Martin et al., 2000	1110111 (6)
B9	Zolotitsa sediments I, Russia	560–550	−31.7	112.9	2.4	Popov et al., 2005	1111111 (7)
B10	Verkhotina sediments	560–550	−32.2	107.1	2	Popov et al., 2005	1110111 (6)
B11	Zolotitsa sediments II	560–550	−28.3	109.9	3.8	Iglesia Llanos et al., 2005	1110111 (6)
B12	Zigan Formation	552–544	−16	138	3.7	Levashova et al., 2013	1110111 (6)
Siberia							
S1	Ust-Kirba Formation	960–930	−8.1	182.6	10.4	Pavlov et al., 2002	0110101 (4)
S2	Kitoi Dykes	762–754	−0.4	201.8	6.1	Pisarevsky et al., 2013	1111101 (6)
S3	Kesyussa Formation	542–535	−37.6	165	5.2	Pisarevsky et al., 1997	1010111 (5)
W. Australia							
WA1	Browne Formation	830–730	44.5	141.7	6.8	Pisarevsky et al., 2007	0010101 (3)
WA2	Hussar Formation	800–730	62.2	85.8	10.3	Pisarevsky et al., 2007	0010111 (4)
WA3	Mundine Dykes	758–752	45.3	135.4	4.1	Wingate and Giddings, 2000	1111101 (6)
N. Australia							
NA1	Johnny's Creek Member	780–660	15.8	83	13.5	Swanson-Hysell et al., 2012	0111101 (5)
S. Australia							
SA1	Angepena Formation	660–640	47.1	176.6	5.3	Williams and Schmidt, 2015	0111111 (6)
SA2	Yaltipena Formation	650–635	44.2	172.7	8.2	Sohl et al., 1999	1111111 (7)
SA3	Elatina Formation, MEAN	645–635	49.9	164.4	13.5	Embleton and Williams, 1986 Schmidt et al., 1991 Schmidt and Williams, 1995 Sohl et al., 1999	1111111 (7)
SA4	Nuccaleena Formation	635–610	32.3	170.8	2.9	Schmidt et al., 2009	1111111 (7)
SA5	Brachina Formation	620–590	46	135.4	3.3	Schmidt and Williams, 2010	0111111 (6)
SA6	Bunyeroo Formation	590–570	18.1	196.3	8.8	Schmidt and Williams, 1996	1111111 (7)
SA7	Wonoka Formation	575–555	5.2	210.5	4.9	Schmidt and Williams, 2010	1011111 (6)
North China							
NC1	Huaibei Sills 890 Ma	913–876	−52.3	149.3	3.5	Fu et al., 2015	1011110 (5)
South China							
SC1	Yanbian Dykes A	830–818	45.1	130.4	19	Niu et al., 2016	1111011 (6)
SC2	Xiaofeng Dykes	812–792	13.5	91	10.9	Li et al., 2004	1110111 (6)
SC3	Yanbian Dykes B	814–798	14.1	32.5	20.4	Niu et al., 2016	1011001 (4)
SC4	Liantuo Formation	735–705	9.9	160.3	4.6	Jing et al., 2015 + Evans et al., 2000 combined	1111111 (7)
SC5	Nantuo Formation	641–631	7.5	161.6	5.9	Zhang et al., 2013; Zhang and Piper, 1997	1111111 (7)
SC6	Doushantuo Formation	614–590	25.9	185.5	6.7	Zhang et al., 2015	1110111 (6)
Congo							
C1	Luakela Volcanics A	770–757	−40.2	122	14.1	Wingate et al., 2010	1110111 (6)
C2	Mbozi Complex	773–713	−46	145	6.7	Meert et al., 1995	1110111 (6)
Sao Francisco							
SF1	Bahia Dykes (N + R)	928–912	7.3	106.4	6.2	Evans et al., 2016b	1111100 (5)
Tarim							
T1	Sugetbrak Formation	635–550	19.1	149.7	9.3	Zhan et al., 2007	0111111 (6)
T2	Tereeken Cap Carbonate	ca. 635	27.6	140.4	9.9	Zhao et al., 2014	0111101 (5)
T3	Zhamoketi Andesite	621–609	−4.9	146.7	3.9	Zhao et al., 2014	0110101 (4)
India							
I1	Malani Igneous Suite grand mean	770–734	69.4	75.7	6.5	Meert et al., 2013	1111111 (7)
I2	Bhander and Rewa formations	650–530	47.3	212.7	5.8	McElhinny et al., 1978	0110111 (5)
I3	Jodphur Group	570–520	1	164	6.7	Davis et al., 2014	0110111 (5)
Seyshelles							
SE1	Mahe Dykes	753–747	54.8	57.6	12.1	Torsvik et al., 2001a,b	1110001 (4)
Rio de La Plata							

Table 1 (*continued*)

Key	Rock unit	Age	Plat	Plong	A95	Author, year	Q-factor
RP1	Sierra de las Ánimas Complex	582–574	12.2	78.9	14.9	Rapalini et al., 2015	1110011 (5)
RP2	Sierra de los Barrientos Redbeds	600–500	15.1	72.6	12.4	Rapalini, 2006	0110111 (5)1

udc.ig.utexas.edu/external/plates/), as a connection on its motion towards the 500 Ma position in Domeier (2016).

6.2. Siberia and North China

A connection between southern Siberia and northern Laurentia during Rodinia with an approximately 30° gap between them, is supported by palaeomagnetic data for the early Neoproterozoic as both share a common APWP from 1050 to 980 Ma (Pisarevsky and Natapov, 2003; Pisarevsky et al., 2003 but see alternative hypothesis – Evans et al., 2016a). An absence of Tonian and early Cryogenian poles from Siberia makes it difficult to test whether this configuration remains constant for the early Neoproterozoic, although a ca. 760 Ma pole from the Kitoi dykes (Table 1) suggests that there had been some relative movement between Siberia and Laurentia since the Mesoproterozoic (Pisarevsky et al., 2013). This pole suggests that Siberia moved closer to Laurentia along a dextral transform boundary, and it is inferred that the beginning of this motion is a response to early stages of Rodinia breakup. Given the spatial proximity of Siberia to Australia in Rodinia, we infer Siberia's motion to be the antithesis of the rifting of Australia off Laurentia, which we depict at ca. 800 Ma (about ~20 Myr earlier than Pisarevsky et al., 2013). Although we note that based on the distance that Siberia moved (~30° of latitude), that this movement beginning anytime between ca. 900 and 800 Ma would produce a spreading rate congruent with present day limits. High quality poles from the late Cryogenian and Ediacaran are also scarce for Siberia. A ca. 540 Ma pole from the Kessyusa Formation (Pisarevsky et al., 1997) necessitates a large (~180°) rotation of the Siberian craton from its 760 Ma position (e.g. Smethurst et al., 1998). Recent palaeomagnetic data from sedimentary successions of Siberia (e.g. Pavlov et al., 2015 and references therein) generally support this rotation, but because of uncertain ages, an unequivocal model for this rotation has yet to be made.

The exact configuration of North China around Laurentia in Rodinia is uncertain due to the disparity in geology between it and other cratons. The reconstruction of North China in Rodinia, close to Siberia (Li et al., 2008), was suggested on the basis of the Precambrian APWP (1300–510 Ma) for North China proposed by Zhang et al. (2006), from a palaeomagnetic study of sedimentary successions in Henan Province. However, recent geochronological data (Su et al., 2012) clearly indicate that the poles of Zhang et al. (2006) are significantly older, and that the proposed 1300 to 510 Ma APWP should be dismissed. On the other hand, the recently published ca. 890 Ma Huaibei Sills pole (Fu et al., 2015) suggests a mid-latitude for North China at this time, consistent with its position relative to Siberia and Laurentia shown in Li et al. (2008). Consequently, we still use this model, although we note that North China's position is very poorly constrained and future palaeomagnetic data may change its position. Palaeomagnetic data from the Dongija Formation dated at ca. 650 Ma (Zhang et al., 2000) and from the Wennan Area (Late Cambrian, Zhao et al., 1992) suggest North China remained in low latitudes between 700 and 520 Ma.

6.3. Australia

The Neoproterozoic palaeomagnetic record from Australia is fragmented; there are no reliable palaeomagnetic poles from the Tonian, although there are a number of poles from the Cryogenian and from Ediacaran-Cambrian sedimentary successions. The sparse record in the early Neoproterozoic is particularly troublesome as it makes it difficult to determine both Australia's fit against Laurentia into Rodinia and to isolate the time of rifting (Section 3.3). Two poorly dated poles, from the Browne and Hussar formations (Pisarevsky et al., 2007), with regional correlation ages between 830 and 800 Ma and 800 and 760 Ma respectively, and a better dated pole from the Johnny's Creek Member (ca. 770 Ma, Swanson-Hysell et al., 2012) (Table 1) suggest Australia was at low latitudes, with the Hussar formation favouring an AUSMEX configuration (Wingate et al., 2002) if rifting from Laurentia was late (post 775 Ma). However, the Browne and Hussar poles may be younger due to their age uncertainties and the closeness of the Browne pole to the Mundine Well pole. Schmidt (2014) recently reviewed the palaeomagnetic Precambrian record for Australia, and determined two 'Grand Poles' (similar results from a key pole from two independent laboratories), the Mundine Dyke Swarms (MDS) at ca. 750 Ma, and the Elatina Formation (EF) at ca. 635 Ma (Table 1). The MDS pole also suggests a low latitude position, and requires that SWEAT-type (including AUSWUS) configurations had already broken up (Missing-Link, or configurations where Australia is more distal from Laurentia can accommodate the MDS pole without earlier breakup). The 40° intraplate rotation of NAC to SAC proposed by Li and Evans (2011) can allow for later rifting of SWEAT-like configurations, and more importantly, reconciles a triad of pairs of poles from the Mesoproterozoic to Neoproterozoic. Significant to Rodinia reconstructions is the restoration of the Walsh Tillite Cap (WTC) (Li, 2000) and the Johnny's Creek Member to the MDS. The absence of high quality, pre-ca. 825 Ma palaeomagnetic data limit our ability to determine the configuration between Australia and Laurentia, because geological evidence for rifting between the two cratons does not conclusively discriminate rifting time more precisely than between 825 and 700 Ma. This means that the Cryogenian-aged poles discussed above are accommodated by most configurations simply by arguing for earlier rifting (an argument supported on kinematic grounds, see above). The series of Ediacaran poles come from 650 to 555 Ma sedimentary successions in South Australia and allow for the construction of an Ediacaran-Cambrian APWP for Australia (Schmidt, 2014). The poles indicate a low-mid latitude position of Australia, consistent with the Gondwana nucleus (i.e. Congo-Amazonia) being over the South Pole.

6.4. India, Seychelles and South China

Palaeomagnetic data from India are sparse, only three (semi-) reliable poles exist for the Neoproterozoic (Table 1); a 770 to 750 Ma pole from dykes in the Malani Igneous Suite (Torsvik et al., 2001a; Gregory et al., 2009; Meert et al., 2013), a ca. 750 Ma pole from the Seychelles (Torsvik et al., 2001b) and a pole from the Bhander and Rewa Series (McElhinny et al., 1978) that is commonly attributed to ca. 545 Ma, but is of questionable age. An earlier pole, the 815 Ma Harohalli dykes (Miller and Hargraves, 1994), suggests a polar location for India, although it has been redated to ca. 1200 Ma and so is discounted here (Pradhan et al., 2008). These poles suggest a movement from polar to equatorial latitudes during the Cryogenian and Ediacaran. The paucity of early Neoproterozoic data make India's position in Rodinia difficult to determine (or even if it was part of Rodinia at all). Traditionally India was placed against the northwestern margin of Australia (e.g. Boger et al., 2001; Powell and Pisarevsky, 2002) in a similar position to its location in Gondwana, in part due to the assumption that East Gondwana acted congruently since the Mesoproterozoic (e.g. McWilliams, 1981). The large mismatch between the 770–750 Ma poles from Malani and the Seychelles with the MDS pole from Australia necessitated early rifting of India from Rodinia (e.g. Li et al., 2008, 2013), although some reconstructions favoured removing India completely from Rodinia (e.g. Collins and Pisarevsky, 2005). We have followed this approach and have excluded India from Rodinia in the model presented here.

Palaeomagnetic data from South China are broadly similar to that from India, with early Cryogenian poles (830–790 Ma) in South China indicating a high latitude position (e.g. Xiaofeng Dykes, Li et al., 2004; Yanbian Dykes, Niu et al., 2016) and successive data from the later Cryogenian and Ediacaran suggesting movement towards lower latitudes (e.g. Liantuo Formation, Evans et al., 2000; Jing et al., 2015, and Nantuo Formation, Zhang et al., 2013; Zhang and Piper, 1997). There is a large discrepancy between the poles of the Xiaofeng Dykes (Li et al., 2004) and the Yangbian Dykes A and B (Niu et al., 2016) that is difficult to resolve without resorting to extra-ordinary explanations such as true polar wander (Table 1). We elect to follow the pole from the Yangbian Dykes A, which is more compatible with that of the Xiaofeng Dykes than is the pole of the Yangbian Dykes B. Assuming a fixed geological relationship between South China and India, a ~55° rotation of South China and India is needed to fit the Indian 770–750 Ma poles from the Mahe Dykes and Malani Igneous Suite, with the ca. 720 Ma Liantuo Formation from South China, which constrains South China to a mid-latitude position. Late Cryogenian and Ediacaran poles from South China suggest an equatorial position, consistent with it moving towards its inferred position in Gondwana (Zhang et al., 2013).

6.5. Congo-São Francisco

A São Francisco palaeomagnetic pole, which was previously dated as Late Mesoproterozoic, suggested a low-mid latitude position (D'Agrella-Filho et al., 1990; Renne et al., 1990), necessitating a gap between C-SF and Laurentia in the Late Mesoproterozoic (D'Agrella-Filho et al., 2004). However, the age of this pole has been recently reconsidered as Tonian (Evans et al., 2016b). During the Neoproterozoic, a 928–912 Ma pole from the Bahia Dykes in the São Francisco Craton constrains its position to low latitudes (Evans et al., 2016b). Originally, two poles in the Congo Craton, the ca. 795 Ma Gagwe Lava and 750 Ma Mbozi Complex (Meert et al., 1995) necessitated a rapid 90° counter-clockwise rotation at low latitudes between 800 and 750 Ma. Further work by Wingate et al. (2010) demonstrated that the ca. 795 Ma Gagwe remanence was probably not primary due to its similarity to the secondary reprint of the ca. 765 Ma (maximum age) Luakela volcanics (component B). However, the Mbozi pole coupled with another reliable 770–757 Ma pole (Luakela Volcanics A, Wingate et al., 2010, Table 1) suggests that the C-SF was relatively stable during the late Tonian, and that the 90° clockwise rotation occurred after 765 Ma (Wingate et al., 2010).

6.6. Rio de la Plata

The only well-dated pole from RDLP is from the Sierra de las Ánimas Complex, which is dated to between 582 and 574 Ma, constraining RDLP to a mid-high latitude (Rapalini et al., 2015). The reliable, but poorly dated Sierra de los Barrientos pole (Rapalini, 2006) is very close to the Sierra de las Ánimas pole (Table 1).

6.7. Tarim

Three Ediacaran poles from Tarim (Table 1) are not precisely dated, but are internally consistent and two of them are supported by positive fold tests. These poles suggest a low-mid latitude position, consistent with Tarim being situated as a northern extension of Australia. As discussed previously (Section 3.3) three earlier poles from the Tonian and Cryogenian are not compatible with Tarim in this position, and consequently we only incorporate Tarim from 700 Ma in this model.

7. Plate model

In this section, we present snapshots of the plate model at times pertinent to understanding the tectonic evolution of the Neoproterozoic. Although the plate model is based on the geological and palaeomagnetic constraints described above, we also present a discussion on the integration of these observations with plate tectonic theory at these time steps, comparing the final model with the current understanding of plate tectonics from better-constrained reconstructions of the Phanerozoic. We encourage readers to access either the animation or both the associated plate model files and Gplates (www.gplates.org) in the online version of the publication (Supplementary Material) to further investigate periods of interest. Where possible we have labelled oceans based on previous and standard nomenclature and, except were otherwise outlined, they refer to the entire ocean basin, as opposed to a specific plate. For example, the Adamastor Ocean refers to the non-cratonic portion of plates that were built from the spreading system between Kalahari and RDLP, and consists of both the Kalahari Plate and the RDLP plate. For the descriptions below, all orientations are expressed relative to the reconstructed model except where explicitly stated that they are relative to present day. To help follow the latitudinal movements of the cratons in this time we have plotted palaeolatitude vs. time for each major craton (Fig. 4a–e).

7.1. 1000–950 Ma, Rodinia

The core of Rodinia occupies equatorial-low latitudes for the early Neoproterozoic, with Australia sitting at ~30° N, and Baltica, Amazonia, West Africa sitting at ~30° S (Figs. 4d, e and 5). The opening of the Araçuaí Basin between the Congo and São Francisco cratons initiates at ca. 1000 Ma (Pedrosa-Soares et al., 2001). Based on the active subduction on the other side of the Congo Craton, preserved in the Southern Irumide arc in Zambia (e.g. Johnson et al., 2006) and the Dabolava arc (Tucker et al., 2007) outboard of Congo against Azania (D.B. Archibald, unpublished data), we move Congo northwards, with the Southern Irumide arc acting as the boundary between the Congo Plate and the larger Rodinian plate. This subduction zone is offset by a transform fault to the north, linking it with the Dabolava arc, and the extension of this subduction system away from Rodinia, along the (present day) northern margin of Congo is the early Neoproterozoic subduction between the SM-Borborema Plate and the C-SF Plate preserved in the Cariris Velhos orogeny (e.g. Caxito et al., 2014; Santos et al., 2010). A transform boundary connects the inferred spreading ridge outside of the SM with the subduction zone between the SM-Borborema Plate and C-SF Plate, although given the complexity of the Hoggar Shield area and cover of the Sahara Desert early small-scale subduction could be preserved at this time as well. This subduction zone forms the furthest extent of the circum-Rodinia subduction girdle, and is connected via an inferred subduction zone outboard of São Francisco (based on crustal displacement due to spreading in the Araçuaí Basin) to distal subduction outside West Africa-Amazonia-Baltica (Fig. 5).

The earliest evidence of the Avalonian and Cadomian terranes is late Mesoproterozoic to early Neoproterozoic (see Section 5.11, e.g. Murphy et al., 2000), yet there is no record of subduction or arc collision in Baltica and Amazonia until the late Neoproterozoic. We therefore extend the circum-Rodinian subduction zone outboard of Amazonia-WAC-Baltica, such that the Rodinian plate in this section contains a large portion of oceanic crust allowing for the development of thick sedimentary sequences. It is offset by a transform fault to account for the more proximal Valhalla orogeny against Greenland (Cawood et al., 2010), and then extended into North China and Siberia to account for Tonian subduction preserved in the North Qinling Terrane (Dong and Santosh, 2016) and the Yenesei Ridge (Volobuev, 1993; Vernikovsky et al., 2016) respectively.

There is no record of subduction in Western Australia during the Neoproterozoic, which is somewhat problematic for both leaving

Australia exposed on the margin of Rodinia, and for its rifting during the Cryogenian. Similar to the distal subduction zone of Baltica and Amazonia, we infer subduction at a distance from the margin, and note that alternate configurations of Rodinia that place either South China or Tarim here would be broadly compatible with subduction taking place. We instead propose a hypothesis that this Tonian subduction is preserved in the basement of the southeast Asian terranes, such as Sibumasu, Indochina and Lhasa, which record faint magmatic age and detrital zircon signatures from the late Mesoproterozoic to the Cambrian (e.g. Lan et al., 2003; Nagy et al., 2000; Qi et al., 2014; Zhang et al., 2014; Zhu et al., 2011, see Section 3.3).

Late Mesoproterozoic palaeomagnetic data from the Majhgawan Kimberlite Pipe in India suggest a low latitude position (e.g. Gregory et al., 2006), although this pole represents just one kimberlite pipe without properly averaging the geomagnetic secular variation. However, Pradhan et al. (2010) reported higher palaeomagnetic inclination from the 1027.2 ± 13 Ma Anantapur dyke swarm of the Dharwar craton, which indicates medium palaeolatitude. The (chronologically) next reliable pole being the 770–750 Ma Malani Igneous Suite and Takamaka dyke from the Seychelles, suggesting polar latitudes (Gregory et al., 2009; Torsvik et al., 2001a,b). As such we tentatively assume an equatorial-low latitude position for India and South China,

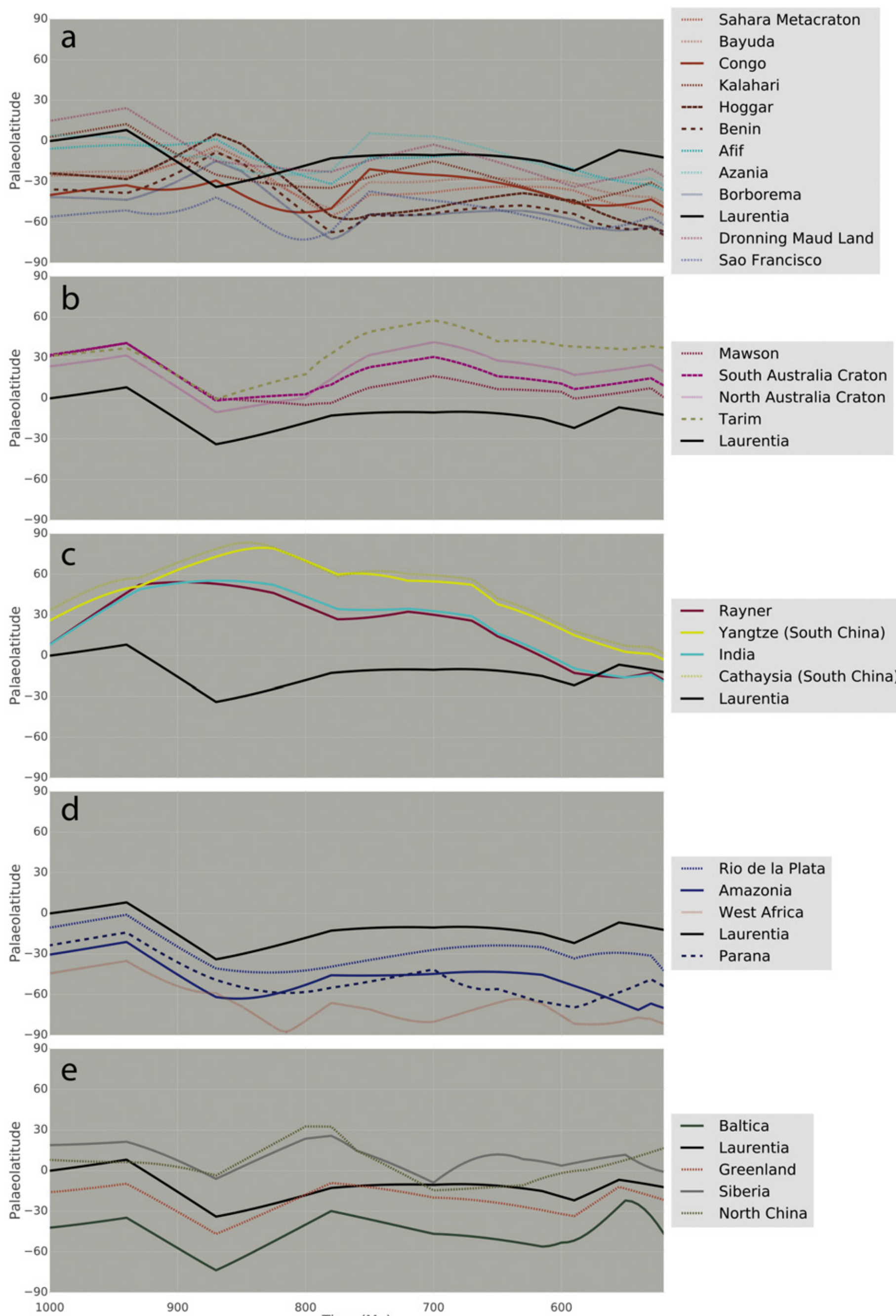

Fig. 4. Palaeolatitude of continental crust fragments in the Neoproterozoic. Each panel broadly reflects a common Neoproterozoic journey; (a) Congo-São Francisco and the Sahara Metacraton (extra south Rodinia); (b) Australia and Mawson (west Rodinia); (c) India and South China; (d) Amazonia (east Rodinia); (e) Laurentia and northern Rodinia. Colour is based on present day geographical position; red – North America; dark blue – South American; green, Europe; grey, Siberia; light blue, India, Madagascar and the Middle East; yellow, China; purple, Australia and Antarctica. Laurentia is plotted in black in each panel as it is considered the heart of Rodinia.

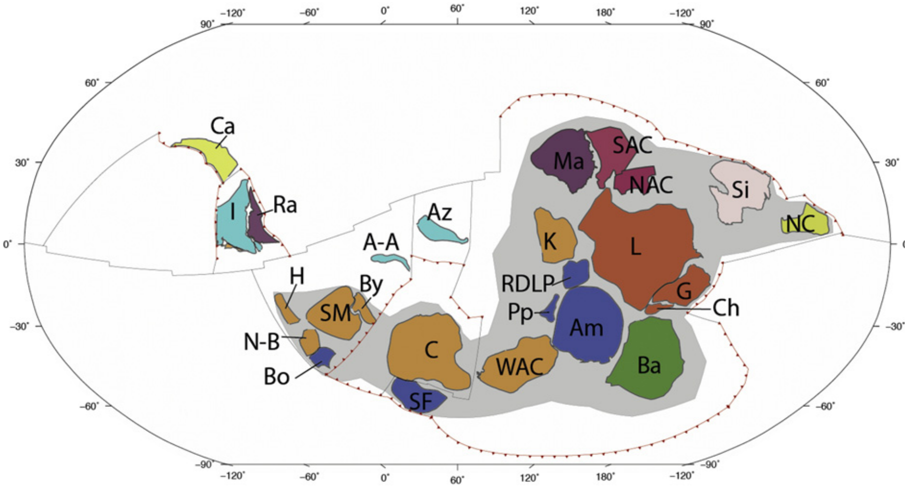

Fig. 5. Tectonic geography at 1000 Ma. A-A, Afif-Abas Terrane; Am, Amazonia; Az, Azania; Ba, Baltica; Bo, Borborema; By, Bayuda; Ca, Cathaysia (South China); C, Congo; Ch, Chortis; G, Greenland; H, Hoggar; I, India; K, Kalahari; L, Laurentia; Ma, Mawson; NAC, North Australian Craton; N-B, Nigeria-Benin; NC, North China; Pp, Paranapanema; Ra, Rayner (Antarctica); RDLP, Rio de la Plata; SAC, South Australian Craton; SF, São Francisco; Si, Siberia; SM, Sahara Metacraton; WAC, West African Craton. Shaded grey area is inferred extent of Rodinia and is meant as a guide only. The longitude is arbitrary and unconstrained, and used here as a relative reference. Cratonic crust is coloured by present day geography: North America, red; South America, dark blue; Baltica, green; Siberia, grey; India and the Middle East, light blue; China, yellow; Africa, orange; Australia, crimson; Antarctica, purple.

with gradual northwards movement through the Neoproterozoic to fit the two younger, robust 770–750 Ma poles. A large ocean basin encompassing the North Pole separates India-South China and Australia-Antarctica-Tarim, we refer to this as the Mawson Sea after Meert (2003), and its closure beneath Rodinia (outboard of Australia, Antarctica and Tarim) and underneath India-South China through the Eastern Ghats, is in part the driving force of pulling India-South China northwards. The Archaean-aged Ruker Terrane collided with the Eastern Ghats-Rayner Province at ca. 960 Ma, although subduction is inferred to continue outboard of this. The present day northwest margin of India and northern margin of the Cathaysia craton of South China underwent substantial growth during the early Neoproterozoic, and

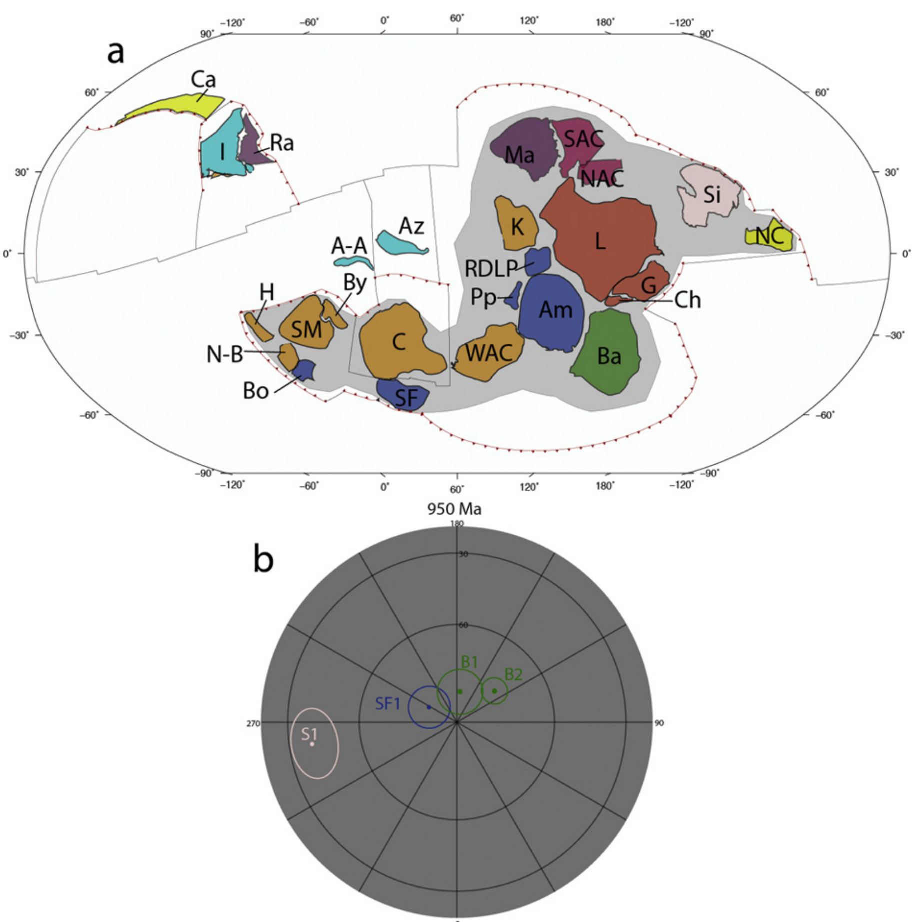

Fig. 6. (a) Tectonic geography at 950 Ma, see Fig. 5 for abbreviations. (b) Palaeomagnetic poles around 950 Ma.

are interpreted here to act as a continuous subduction zone through this time, with accretion occurring for most of the Tonian and including rocks now found in Pakistan and Oman (Alessio et al., in press; Whitehouse et al., 2016).

7.2. 950–850 Ma

At 950 Ma Rodinia remains in a similar latitudinal position to where it was at 1000 Ma (Fig. 4). The plate boundaries are also broadly similar, although we infer a jump in the subduction between Azania and Congo after the collision of the Dabolava Arc with Azania at ca. 950 Ma, with subduction now occurring continuously against the western margin of Azania in the Neomozambique Ocean. This subduction zone is inferred to consist of (from present day south to north) the Antarctic TOAST terrane (Jacobs et al., 2015), the Malagasy Vohibory volcanic arc (Collins et al., 2012) and the Galana Arc (Hauzenberger et al., 2007), and extends northwards towards the SM, through the ANS, correlating it with the early Neoproterozoic (1030–930 Ma) Sa'al Arc (Eyal et al., 2014). This presupposes the legitimacy of extending the Neomozambique Ocean (between Azania and Congo) northwards to between the ANS and SM, as opposed to discrete positions of the ANS and Azania for the early Neoproterozoic. However, it does reconcile mid-Tonian metamorphism and magmatism on the eastern margin of the Bayuda Block (Küster et al., 2008), implying that a long subduction zone was active during the early-mid Tonian along this margin (Fig. 6). The subduction turns off around ca. 900 Ma with the collision of the Malagasy-Vohibory arc with Azania, and this part of the C-SF margin becomes tectonically quiet until ca. 850 Ma. On the present-day western margin of the SM, in the Hoggar Shield, evidence of early sea-floor spreading (pillow basalts), with a transition to a more dominant subduction regime occurring from ca. 900 Ma with the formation of the Iskel magmatic arc (Caby, 2003). Further south, the existence of this subduction zone is also supported by subduction initiating between 900 and 800 Ma preserved in juvenile arcs on the western margin of the Borborema Province (Ganade de Araujo et al., 2012, 2014a), suggesting that by the mid-Tonian the SM was surrounded by large subduction zones. The subduction zone outboard of Borborema is offset slightly, but otherwise occurs along the same margin as the inferred subduction zone outboard of the São Francisco craton due to spreading in the Araçuai Basin.

As with the spatial uncertainties between the ANS and the SM, it is unknown how close the pre-Neoproterozoic crustal elements (Hoggar, Borborema) were to the SM due to an absence of palaeomagnetic data and significant crustal reworking during Gondwana's amalgamation. We place the subduction zone close to Borborema due to the stronger evidence of earlier subduction preserved here and slightly more distal from Hoggar due to the later evidence of subduction. Palaeomagnetic data from Baltica indicate that it lay in polar regions by ca. 870 Ma. Assuming the congruency of Rodinia at this time and to limit plate velocity, we rotate Rodinia from 940 Ma so that by 870 Ma Baltica sits near the geographic south pole (Figs. 4 and 7a) (Walderhaug et al., 1999, 2007). The implication of this is that both the C-SF and the SM portions of crust now occupy equatorial positions, and, given the geometry of Rodinia, subduction is inferred to have encompassed the SM from 940 Ma to accommodate this motion.

North of the SM and separated by an inferred spreading system, the India-South China plate is continuing to move northwards, with the subduction-accretion forming the Yangtze Block and north-western India through this time. Subduction on the other side of India, outboard of the Eastern Ghats, is still interpreted to be ongoing after collision of the Ruker Terrane against the Rayner Province, with granulite facies

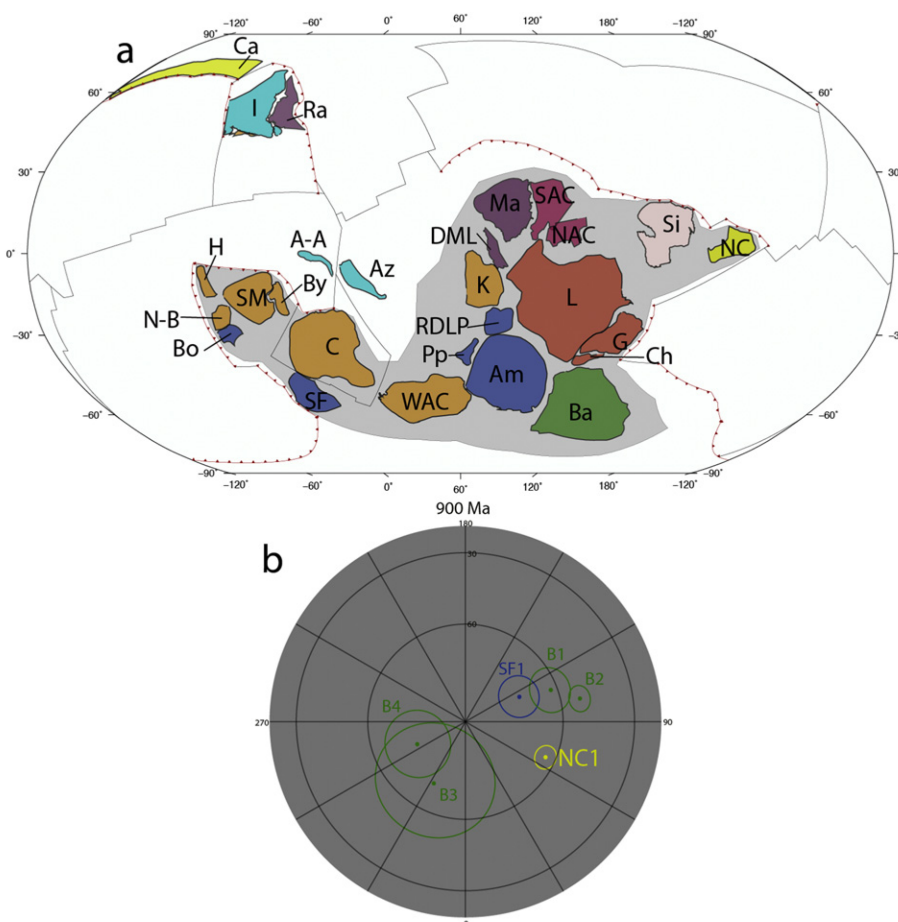

Fig. 7. (a) Tectonic geography at 900 Ma; DML, Dronning Maud Land, for other abbreviations see Fig. 5. (b) Palaeomagnetic poles around 900 Ma.

metamorphism and overprinting preserved (e.g. Gupta, 2012), suggesting that a convergent tectonic regime was still present. This ongoing convergence is a principal reason why we argue that India (and by extension South China) was not part of Rodinia.

7.3. 850–800 Ma, Congo-São Francisco displacement

After the bulk of Rodinia transitioned to higher latitudes, the supercontinent drifted back towards equatorial-mid latitudes by ca. 800 Ma, with Laurentia occupying a low-mid latitude position, and Australia-Antarctica and Siberia lying on the equator (Figs. 4 and 9a). The C-SF plate (along with the SM, ANS and Azania) underwent a relative movement with respect to the rest of Rodinia from a mid to equatorial latitude to fit the reliable ca. 750 Ma Mbozi Complex and Luakela Volcanics A poles (Meert et al., 1995; Wingate et al., 2010). The spreading system associated with this movement forms two passive margins. Firstly, along the (present day) northern, eastern and south-eastern margins of the WAC by 850 Ma, accounting for the ~4 km thick passive margin sequences in the north, and the Gourma Aulacogen towards the south (e.g. Bouougri and Saquaque, 2004; Thomas et al., 2002, 2004). Secondly, the conjugate margin here is preserved on the southern flank of the C-SF Plate with the deposition from ca. 850 Ma of the Katanga Supergroup (Bull et al., 2011). Earlier rifting here minimises the spreading velocity required to move C-SF and the SM from the pole to the equator to fit the palaeomagnetic data (~7 cm/yr vs. ~14 cm/yr for 850 and 800 Ma rifting respectively).

To accommodate spreading on the southern side of C-SF, subduction initiates on the northern margin. There is some evidence of intra-oceanic subduction preserved in the oldest rocks of the Western Ethiopian Shield (WES), indicating pulses of magmatism from ca. 850 to 820 Ma (Ayalew and Peccerillo, 1998; Blades et al., 2015), although the precise spatial relationship of the WES to both the SM and ANS prior to the East African Orogeny is still uncertain. Minor arc accretion was occurring in the ANS (e.g. Johnson and Woldehaimanot, 2003; Johnson et al., 2011), although subduction in Azania corresponding with the formation of the extensive Imorona-Itsindro suite from ca. 850 Ma is suggested to be the primary driver of pulling C-SF northwards (Handke et al., 1999; Archibald et al., 2016, in press). Due to the scope of this study, we simplify the accretion of the terranes in the ANS to an extensive subduction zone, since spreading was likely small scale and preserved in back-arc basins while the dominant tectonic regime, suggested through the synthesis of broadscale geological evidence and palaeomagnetic data (assuming a fixed SM-C-SF Plate), is subduction. We propose that future work and regional studies can focus on building a consistent plate model of the different terranes in the area that can be integrated into this larger global model and note that more palaeomagnetic data are needed to better constrain rifting time and the motion of C-SF.

India and South China were located at polar latitudes (on the ‘north’ pole) by 850 Ma (Figs. 4, 7), and subduction in the Eastern Ghats had ceased, although there was still some minor tectonic activity occurring. The motion of the Indian and South China Plate during the late Tonian reached its zenith at 820 Ma with South China sitting on the pole. After this time, India and South China begin their southward drift towards Gondwana. Geologically, the Yangtze and north-western India were fully assembled, and subduction is inferred to have entered a hiatus between 850 and 800 Ma when magmatism along the southern margin of India (Tucker et al., 2001), in the Seychelles (Sato et al., 2010) and on the (present-day) northern margin of the Yangtze Craton (Yan et al., 2004; Zhou et al., 2002) started again.

7.4. 800–750 Ma, Rodinia breakup initiates

The first stage of Rodinia breakup began around ca. 800 Ma, with the opening of the Proto-Pacific Ocean between Australia-Antarctica-Tarim and Laurentia, which had returned to an equatorial-low latitude position, with Baltica-Amazonia-WAC located in mid latitudes further south (Figs. 4 and 9a). The spreading system separating Laurentia and Australia likely extended further north, and is inferred to be the spreading ridge associated with Siberia's dextral movement against Laurentia between 800 and 700 Ma (Pisarevsky et al., 2013). The accommodation of the motion of the Australian Plate away from Laurentia is by inferred subduction outboard of Australia and Antarctica. Assuming a close spatial relationship between Tarim and Australia, the change to retreating subduction preserved in the western Kuruktag area of Tarim (Ge et al., 2014) suggests that the Australian Plate was the overriding one.

The absence of preserved early Cryogenian arcs in Australia could be accounted for by placing the Lhasa terrane against the west coast of Australia as it records Tonian and Cryogenian aged magmatism (Guynn et al., 2006; Qi et al., 2012; Zhang et al., 2012a,b, 2014). In Antarctica, snow and ice cover the suture between Mawson and the other Precambrian constituents of Antarctica, but a prediction of this model is that a series of 800–700 Ma arc related rocks exist, separated by N-S sutures between the parts of Antarctica associated with the Mawson craton and Neoproterozoic India (e.g. Chron Craton of Boger, 2011).

Siberia's motion is accommodated by a cessation of subduction along Greenland (the Valhalla Orogeny), with the main subduction of oceanic crust between Siberia and Baltica occurring outboard of Baltica, although the presence of this arc is unknown. The (present day) eastern and southern margins of Laurentia remained mostly intact during this time, although separation due to small scale spreading of WAC and Amazonia is inferred from the presence of a ca. 757 Ma ophiolite preserved in the Araguaia Belt between Amazonia and the WAC (Paixão et al., 2008). Kalahari remained attached to Laurentia, and its relationship with the Australian Plate is along a transform fault, based on Cryogenian palaeomagnetic data from Australia that show it remained in low-mid latitudes (Fig. 4). This boundary accounts for the absence of passive margin sedimentation along the (present-day) eastern margin of the Kalahari Craton. During this time, subduction in the ANS was (present-day) N-S aligned with the amalgamation of the western oceanic domain by ca. 750 Ma, inboard of the Afif-Abas Block.

India and South China were located at mid-high latitudes at 800 Ma and palaeomagnetic data suggests that they underwent a ~35° counterclockwise rotation by 720 Ma. Subduction had ceased on the (present-day) eastern margin of India, although the western margin records magmatism in the Seychelles and the Malani Igneous Suite (e.g. Tucker et al., 2001) until ca. 750 Ma, and along the Yangtze Craton, but not on the north-western margin of India, hence the two subduction zones are offset by a small transform fault. The preservation of an ophiolite in the Manumedu Complex of southern India indicates that subduction extended to the southern margin of India and is interpreted here as the initiation of the closure of the Mozambique Ocean and the onset of Gondwana amalgamation. The breakup of Rodinia and onset of southerly growth, motion and subduction of India is inferred to alter the paradigm of seafloor spreading in the Mirovoi Ocean (analogous to how Pangaea breakup changed Panthalassa spreading?), and the triple junction that existed until now in the Mirovoi Ocean is replaced a by a single ridge system.

7.5. 750–700 Ma, Congo-São Francisco rifting

By 750 Ma C-SF is located adjacent to the Kalahari Craton in a position similar to Li et al. (2008) to fit palaeomagnetic data (Figs 4, 10a and b). The Proto-Pacific spreading system separating the Australian Plate from Rodinia extended further southwards, breaking C-SF from the Kalahari. We posit that this is plausible as Australia is inferred to have ‘unzipped’ from Laurentia from the north towards the south in order to maintain a low latitude position and to necessitate the ‘fan-like collapse’ of Rodinia during the transition to Gondwana (Hoffman, 1991). The southerly extent of this unzipping is the equatorial and most distant craton from the Rodinian core, C-SF. The extension of the spreading ridge would most likely occur on a pre-existing zone of weakness, in this case the transform fault that separated the C-SF plate from Kalahari. From 750 Ma, the combined C-SF and SM plate is almost completely surrounded by subduction, with only the (present-day) southern margin of the Congo Craton preserving a passive margin (e.g. McGee et al., 2012a). There is evidence of subduction preserved in the northern ANS with the final suturing of the central and western terranes. Assuming a close spatial relationship between the ANS and SM, we continue this subduction zone outboard of the SM (early development of peri-Gondwanan terranes?) and into Hoggar, where the (oblique) subduction and accretion of the LATEA and Iskel terranes were taking place (Caby, 2003; Liégeois et al., 2003). Subduction outboard of the (present day) northern SM margin is inferred, partly to accommodate C-SF spreading from Rodinia, and partly to link the subduction on each side of its margin. Similar to our simplification of the early tectonism in the ANS as a primarily convergent regime, we depict the assembly of Hoggar and LATEA as a convergence dominated tectonic regime with the expectation that future iterations of the reconstruction can more closely model the terrane amalgamation.

The Australian plate continued to grow from the spreading centre between it and Laurentia. At the same time, the Mawson Sea (between India and Australia-Antarctica) shrunk. India and South China remained at similar latitudes as their 35° counter-clockwise rotation finished by ca. 720 Ma, with subduction ceasing on its northern margin along the Yangtze, and occurring solely on its southern margin (e.g. Collins et al., 2014; Yellappa et al., 2010).

On the eastern margin of Rodinia, Amazonia-Baltica-Laurentia continued to remain stable, with the proto-Avalonian and Cadomian arcs developing outboard of West Africa and Baltica. Sedimentation outboard of Greenland continued (e.g. Malone et al., 2014), and also outboard of the Timan margin of Baltica (e.g. Siedlecka et al., 2004) during the Cryogenian, indicating that there was no arc-continent collision here at this time. We follow the depiction of Malone et al. (2014) by extending the Valhalla Orogeny outboard of this margin, allowing for a thick passive margin, while preserving the cirucm-Rodinia subduction system. The remaining portion of Rodinia finishes a (subtle) 20° clockwise rotation to better fit the Franklin Dyke Swarm's palaeomagnetic pole at ca. 720 Ma, with West Africa occupying a polar latitude, and Amazonia and Baltica in mid latitudes (Figs. 4 and 11a).

7.6. 700–600 Ma, Kalahari rifting

Rodinia remains in a similar position, with Laurentia in equatorial-low latitudes, Baltica-Amazonia in mid latitudes and West Africa in high latitudes (Fig. 4). Rifting between the Kalahari Craton and Laurentia occurred at 700 Ma, although this is not constrained palaeomagnetically, and poorly constrained geologically, such that rifting time anywhere from 725 to 675 Ma is acceptable. We opted for 700 Ma for three reasons; firstly, the earliest evidence of subduction in the Damara-Lufilian-Zambezi Belt is ca. 675 Ma (John et al., 2004a). Secondly, assuming a Neoproterozoic connection between Kalahari and Laurentia, later rifting (post 650 Ma) of Kalahari is problematic, as Rodinia begins a clockwise rotation to fit 615 Ma palaeomagnetic data (Long Range and Egersund Dykes from Laurentia and Baltica respectively), which moves the (present-day) southern margin of Laurentia away from Gondwana. If Kalahari remains attached in its 'usual' position, then the ocean basin requiring closure becomes incredibly large (~9000 km) and narrow, due to the proximity of Australia-Antarctica, something we consider unlikely given the distribution of plate boundaries at this time. Finally, as Australia-Antarctica occupy a similar relative palaeolongitude (based on spreading rates) and since the trajectory of Australia-Antarctica spreading must change in order for it fit Ediacaran palaeomagnetic data, we infer a ridge jump for Australia, with the new orientation of the spreading ridge matching the inferred rifting trajectory of Kalahari off Laurentia. Following the unzipping of the supercontinent, we infer that the extension of this rift to the south continued along the RDLP and Amazonian margin of Rodinia, possibly rifting off other micro-blocks or terranes from Rodinia and transferring them to the C-SF. We currently infer that the Paranapanema Block to be one of these, based on the collisional orogenies that ring it (e.g. Riberia belt against the São Francisco Craton, Brasília Belt against Amazonia), suggesting that it was, for a time in the Neoproterozoic, unconnected to the larger South American cratons that are presently juxtaposed against it.

The C-SF and SM plate is surrounded by subduction by 680 Ma. As mentioned above, the earliest evidence for subduction in the Damara-Lufilian-Zambezi Belt between Congo and Kalahari is preserved in eclogitic oceanic crust in the Zambezi Belt, dated to between 675 and 590 Ma (John et al., 2003, 2004a). Subduction along the southern margin of São Francisco is preserved in the Ribeira Belt, inferred here to represent collision of the Paraná Block with São Francisco by 630 Ma. The collision of the Coastal Terrane from 650 Ma in the Kaoko Belt, outboard of Congo, is synchronous with the collision of the Oriental Terrane slightly further north in the Ribeira Belt, outboard of São Francisco. The tectonic evolution of the Dom Feliciano Belt, which preserves long lived subduction from the Cryogenian through to the Cambrian, is more difficult to untangle in a global context and would likely benefit from a more detailed regional model. Two key issues are where the Dom Feliciano Belt developed (adjacent to the RDLP or as small continental terranes in an open ocean) and how the belt preserves sinistral deformation given that the overwhelming regime of transition from eastern Rodinia to western Gondwana is dextral (e.g. Transbrasiliano lineament). If it developed purely on the margin, or slightly outboard, of the RDLP (where it is preserved) the traditional position of RDLP in Rodinia (e.g. Gaucher et al., 2011) would not be valid as it is surrounded by other portions of cratonic crust until the opening of the Iapetus Ocean, and experienced extension rather than convergence at this time. Alternatively, having the RDLP as a 'separate' cratonic element to Rodinia is problematic in a plate-modelling scenario as there are no palaeomagnetic data to constrain its position, and so it would become essentially a 'ghost' craton somewhere in an ocean. A third option is to have the Dom Feliciano Belt evolve in a position completely removed from RDLP, and their close spatial relationship be a consequence of late Ediacaran-early Cambrian tectonics that juxtaposed them (e.g. Rapela et al., 2011). We opt to leave RDLP as part of Rodinia and have the Dom Feliciano Belt evolve outboard within the Adamastor Ocean until 600 Ma, perhaps as a consequence of rifting between C-SF-Kalahari at 750 Ma, and perhaps acting as the (most) southerly extension of the Kaoka and Ribeira belts from 700 Ma. Here the exotic (to RDLP) Nico Perez Terrane was perhaps a micro-continental slither that rifted off Rodinia and became the locus for subduction. If so, it could be tectonically related to both the Coastal and Oriental terranes further north. Rapela et al. (2011) showed that most of the sinistral transpression deformation occurred post 600 Ma, and inferred the ca. 540 to 520 Ma RDLP collision with Kalahari to be the driver of this deformation. This is somewhat problematic, given RDLP's position in Rodinia, since it should be approaching from the (present day) south, along a similar path to the Kalahari Craton, and not sliding from the north (as if it were adjacent to C-SF). Instead, we suggest that the sinistral deformation evident here is driven by closure of the Araçuaí Ocean between Paraná-São Francisco and Congo that closed mostly between 600 and 570 Ma, leading to tectonic escape towards the south.

Subduction on the western margin of São Francisco in the Brasília Belt was underway with the development of the Goiás magmatic arc and accretion of the Goiás Massif. The relationship of the Goiás Arc and massif to the broader palaeogeography beyond their suturing to São Francisco is unknown. Palaeomagnetic data between 750 and 720 Ma place Amazonia (palaeomagnetic constraints from Franklin-Natkusiak in Laurentia) and C-SF (Mbozi Complex) orthogonal to one another and separated by ~30° of latitude. By ca. 650 Ma with the C-SF rifting from the southern margin of Rodinia, even assuming slow spreading (20 mm/yr), a ~2000 km ocean basin would separate the two. Furthermore, the separation of the Brasília Belt and younger Araguaia and Paraguay belts by the Transbrasiliano Lineament, suggests that the late Cryogenian-Ediacaran development of the Brasília Belt was probably unrelated to events occurring on the margin of Rodinia during this time, and that the Goiás Arc and massif evolved away from Amazonia. Assuming a (relatively) close and fixed spatial position between the C-SF and the SM-Borborema-Hoggar cratons, the juvenile, oceanic Goiás magmatic arc, and Archaean Goiás Massif would act as a southerly extension of the magmatism and accretion of terranes in the Hoggar Shield. Here also, juvenile arc development is associated with accretion of older thin slithers of Archaean and Palaeopoterozoic terranes (e.g. Caby, 2003; Liégeois et al., 2003). We (tentatively) propose a correlation between the westerly growth of the two areas. We infer that this subduction is preserved in between these areas along the margin of Borborema, where arc-related granitoids are preserved from 670 Ma, and detrital zircons with an age range of 780 to 617 Ma and a peak at ca. 690 Ma suggest that subduction could be traced back to the early Cryogenian (Ganade de Araujo et al., 2016). We depict some transform motion (i.e. Transbrasiliano Belt) between the growing Gondwana and dwindling Rodinia from 650 Ma due to the clockwise rotation of Rodinia to fit palaeomagnetic data. The 650 Ma Bassaride Belt on the (present-day) southwestern margin of the WAC formed with the closure of the narrow ocean basin between it and Amazonia.

Closure of the Mozambique Ocean under Azania and the ANS, and closure of the Neomozambique Ocean, under Congo and SM, were occurring, with India and South China being pulled south. Subduction outboard of Australia, and closure of the Mawson Sea is inferred to have ceased by 670 Ma due to proximity, without collision, between Australia and India. From ca. 650 Ma India begins sliding past Australia along the sinistral Darling Fault (Collins, 2003; Fitzsimons, 2003), and we rotate Australia counter-clockwise to help this motion and also to maintain its low latitude and to match its APWP during the Ediacaran, which concludes with the ca. 570 Ma Wonoka Pole and suggests an almost 180° rotation from its position in Rodinia. The inferred mid ocean ridge that, in part, drives the rotation of Rodinia is extended through to Australia (Fig. 12). In the east, it creates some small-scale displacement, (perhaps accounting for sedimentation on the eastern margin of Australia) and is offset via a transform fault from here, northwards, to between cratonic Australia (+Lhasa?) and Tarim. This spreading forces the 40° intraplate rotation of Li and Evans (2011) of the NAC to the SAC and forms the Petermann and Paterson Orogenies (Bagas, 2004; Raimondo et al., 2010) and also causes the rifting of Tarim away from Australia by Cambrian times.

Juvenile arcs in the Avalonian and Cadomian terranes formed then slowly moved towards Baltica, West Africa and Amazonia. Avalonia collided with Amazonia by 650 Ma, as exhibited by the timing of peak metamorphism. Subduction remained ongoing after this collision, likely stepping away from the suture zone to form a continental margin. Cadomia is less well known, but a similar event is inferred (Garfunkel, 2015; Murphy et al., 2013; Nance et al., 2008). There is some minor, small-scale, separation and rotation between Baltica and Amazonia between 615 and 600 Ma, although the Iapetus Ocean remained closed during this time. On the other side of Laurentia, rifting of Siberia and North China away from Rodinia started. There was a large counter-clockwise rotation of Siberia to fit Ediacaran- to Cambrian-aged palaeomagnetic data (Pisarevsky et al., 1997), necessitating large sinistral faults against northern Laurentia. This was driven, in part, by spreading between Siberia and an unknown craton, inferred here to be North China, as Baltica remained attached to Laurentia. However, North China is inferred out of convention and tradition rather than from data, since there are no palaeomagnetic constraints on it for this time. In the Cryogenian, subduction outboard of the Siberian margins

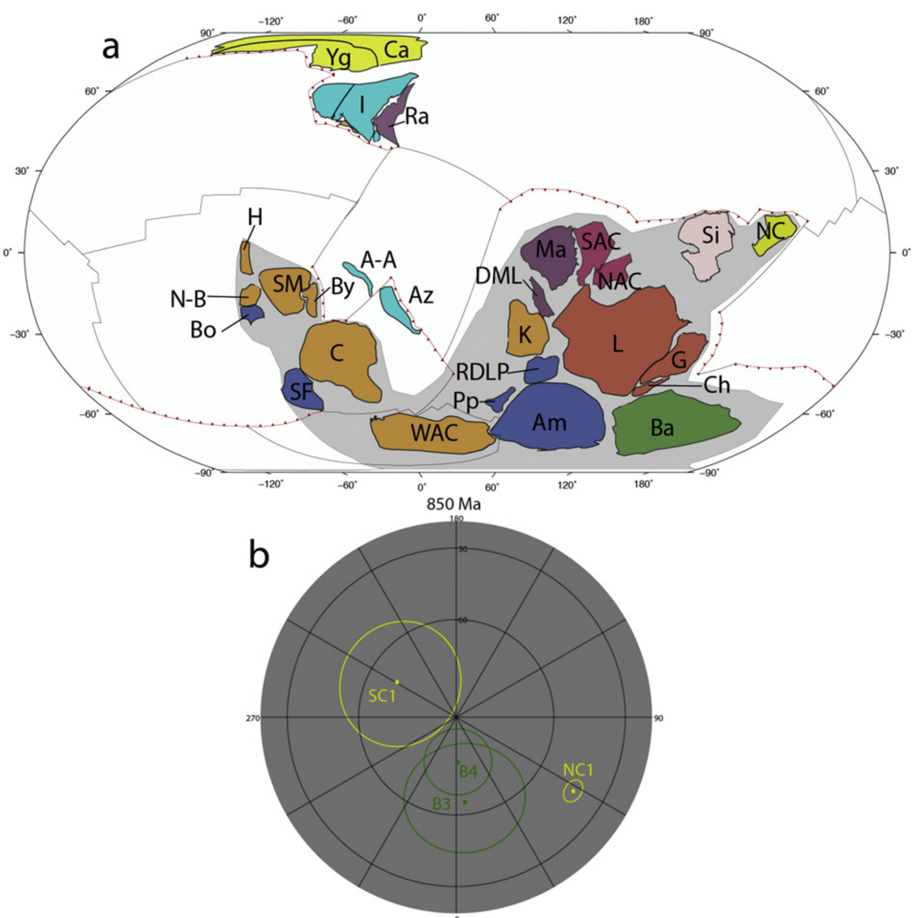

Fig. 8. (a) Tectonic geography at 850 Ma; Yg, Yangtze Craton (South China), for other abbreviations see Figs. 5 and 6. (b) Palaeomagnetic poles around 850 Ma.

(e.g. Yenesei Ridge) occurred, with the accretion of a series of terranes (e.g. Kara, Angara; Metelkin et al., 2012, Vernikovsky et al., 2004) that continued from this time right up to the Cambrian.

7.7. 600–520 Ma, opening of Iapetus Ocean and Gondwana amalgamation

By 600 Ma, most of the subduction zones that facilitated Gondwana amalgamation had formed. However, the opening of the Iapetus and the motion of Amazonia and the WAC into the C-SF and the SM drove the youngest Gondwana-forming orogenies (Figs. 13 and 14). The eastern Iapetus (between Baltica and Laurentia) opened first at ca. 600 Ma, with small degrees of extension between Amazonia and Baltica. The western arm of Iapetus, between Amazonia and Laurentia, began rifting at 590 Ma. In the WAC, the extensive oceanic crust that developed outboard of its Tonian-Cryogenian passive margin acted as a strong pulling force as it subducted, dragging it towards the SM and closing the Pharusian Ocean. Here subduction dipped away from WAC, beneath the growing Gondwana, and transpressive tectonics and oblique convergence are inferred to have occurred between 610 and 580 Ma in both the Dahomeyide and Pharuside belts (e.g. Caby, 2003; Paquette et al., 1998; Berger et al., 2014). Some minor motion between Amazonia and WAC is invoked to shift the WAC from its suture against Amazonia preserved in the Bassaride Belt, to the more southerly 550 Ma Rokelide Belt. The Araguaia and Paraguay belts, along the present day eastern margin of Amazonia, were also active during this time. The (present-day) northerly movement of the WAC and Amazonia into Gondwana is here used to account for the stronger deformation and higher metamorphic grade preserved in the Araguaia Belt than is seen in the Paraguay Belt, as it was on the leading edge of the plate when it collided with C-SF. The Paraguay Belt only preserves greenschist facies metamorphism (Pimentel et al., 1996), which is explained by its location on a trailing or peripheral margin of the plate. The poorly dated rift-drift transition that is preserved in the Paraguay Belt could represent a motion of RDLP away from Amazonia (as RDLP sits adjacent to the location of the Paraguay belt in Rodinia), and the late deformation (550–520 Ma) here could be a response to the collision of RDLP with the rest of Gondwana and closure of the Adamastor Ocean along the 550–540 Gariep belt (e.g. Frimmel and Frank, 1998). The Araguaia Belt, which preserves a ca. 750 Ma ophiolite, records a slightly longer history of sedimentation and deformation than the Paraguay Belt, and the timing of collision here, at 550 Ma, is inferred to represent the primary suture between the two largest cratonic crust components of Gondwana (C-SF and Amazonia-WAC). The dextral Transbrasiliano Lineament is the structural boundary that separates the two sections of crust (Figs. 11 and 12). We acknowledge that the notion of a long (~6000 km!) strike-slip fault is problematic, although it accounts quite well for the relative motion between Laurentia-Amazonia-WAC and Gondwana during the Ediacaran, both during the subtle clockwise rotation of Rodinia to fit 615 Ma palaeomagnetic data, and in juxtaposing Amazonia-WAC in their Gondwana configuration from an equatorial Laurentia. It is our expectation that the lineament probably espouses some component of oblique subduction, and that a more realistic interpretation would be a series of small transform faults linking and offsetting oblique subduction in the Paraguay, Araguaia and Brasília belts, and perhaps the Dom Feliciano and Dahomeyide belts. Such a scenario would benefit from further untangling the spatial and temporal relationships of the small exotic terranes caught up in the orogenies along the margins of the (present day) African cratons (e.g. Coastal, Oriental, Nico Perez terranes) to create a more nuanced model for the amalgamation of western Gondwana. Further east, closure of the Khomas Ocean

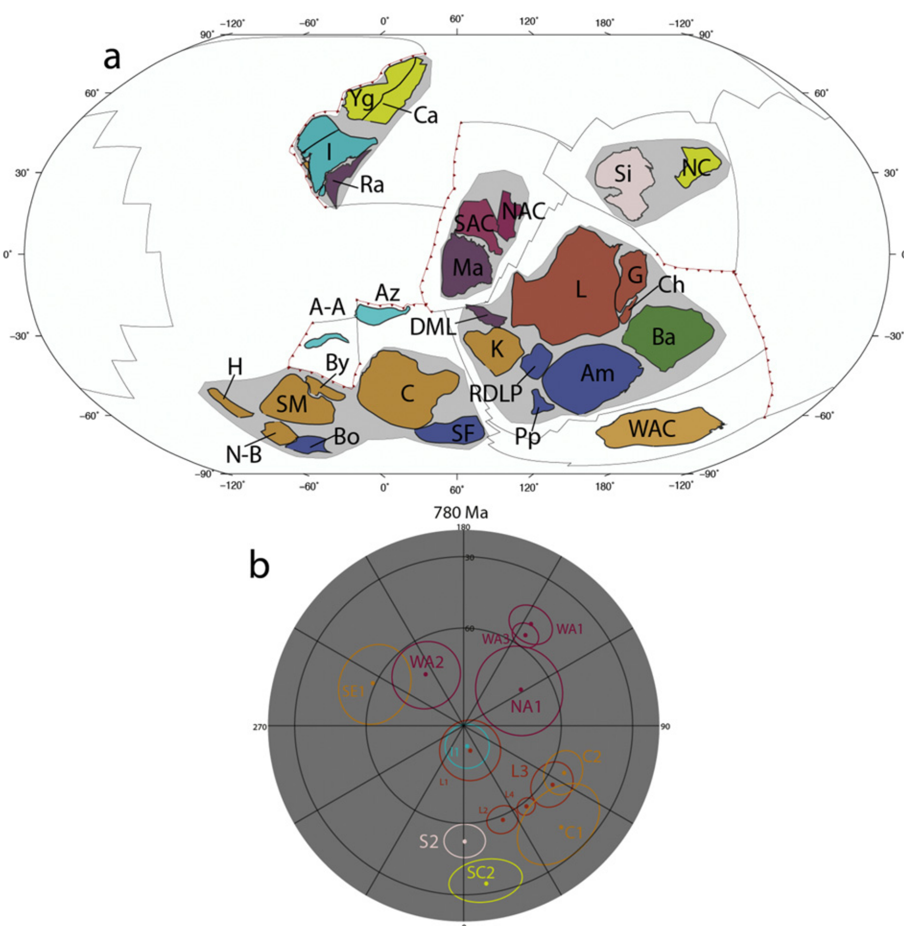

Fig. 9. (a) Tectonic geography at 780 Ma, see Figs. 5, 6 and 7 for abbreviations. (b) Palaeomagnetic poles around 780 Ma.

between Congo and Kalahari was completed by 550 Ma with the development of the Damara-Lufilian-Zambezi Orogeny, although some later transpressive deformation may have occurred with the collision of RDLP along the Mwembeshi Shear Zone (e.g. Naydenov et al., 2014).

The relationship and suture between Australia-Antarctica and Dronning Maud Land (attached to Kalahari) is unknown due to ice cover. However, from permissible positions of each craton from palaeomagnetic data (e.g. Wonoka Pole, mean Gondwana APWP), indicate that the motion would be along a transform fault as their latitudes prior to collision preserved in the Pinjarra Orogeny are similar to their Gondwana fit (Fig. 4). The enigmatic Coats Land Block and Crohn Craton of Boger (2011) could perhaps record some of this deformation and collision, although otherwise the suturing of Australia-Antarctica to Gondwana at 520 Ma marks the final major Gondwana building orogen (Fig. 14). Further north in Australia, the intraplate rotation of the NAC into the SAC finishes by 550 Ma, a similar time to the closure of the Neomozambique and Mozambique oceans between Azania and Congo, and India and Azania, respectively. Further north (present day), the ANS finished assembly by 550 Ma and is accreted onto the SM. Outboard on the (present-day) northern margin of the SM, WAC and Amazonia the peri-Gondwanian terranes are developing (e.g. Avalonia and Cadomia, Armorica, Florida, Iberia etc.).

Finally, the kinematic evolution of the non-Gondwanan constituents (Laurentia, Siberia and Baltica) suffers from two key problems. Firstly, the current 'Ediacaran nightmare' (cf. Meert, 2014b) of palaeomagnetic data from both areas makes it difficult to reliably trace the latitudinal wandering of these continents. Secondly, a lack of continent-continent collisions during the Ediacaran and early Cambrian makes it difficult to pin them to a tectonic environment or to another continent (especially Baltica and Siberia). Thus, the motion of Baltica, post Iapetus opening, is problematic and forms one of the most poorly constrained motions of any major continent in the model. An appropriate solution is probably to look at the post-Iapetus opening evolution of Baltica as a single entity. Here we have its motion between 550 and 520 Ma depicted to connect to its 500 Ma position described in Domeier (2016). Baltica rifts off from Laurentia at mid-high latitude and moves towards an equatorial position, to fit high quality ca. 550 Ma palaeomagnetic data (e.g. Lubnina et al., 2014), in close proximity to Siberia. We infer two drivers for this motion, firstly subduction in the Timan area of Baltica acts as the leading edge, with slab roll-back pulling Baltica northwards. Secondly, the opening of the ocean between Gondwana and Baltica (orthogonal to the Iapetus Ocean) is inferred to exert more pushing force on Baltica than the rift between Baltica and Laurentia, leading to a northerly motion of Baltica. Nonetheless, this position is puzzling, as with two large continents (Baltica and Siberia) so close to one another, we would expect an eventual collision. By 550 Ma Siberia is finishing its large rotation to fit the palaeomagnetic pole from the Kesyussa Formation (Pisarevsky et al., 1997). From this point we infer a spreading system opening between Siberia and Laurentia to begin pushing Siberia further east to collide with Chinese blocks that rift from Gondwana during the Palaeozoic (e.g. Domeier and Torsvik, 2014). This re-organisation and spreading ridge is (loosely) inferred to act as a driver to push Baltica away from both Siberia and Laurentia and back towards high latitude and its 500 Ma position.

8. Discussion

Below we briefly summarise some characteristics of the plate model that have relevance to geodynamic understanding of plate tectonics in the Neoproterozoic. Complete topological plate reconstructions can be

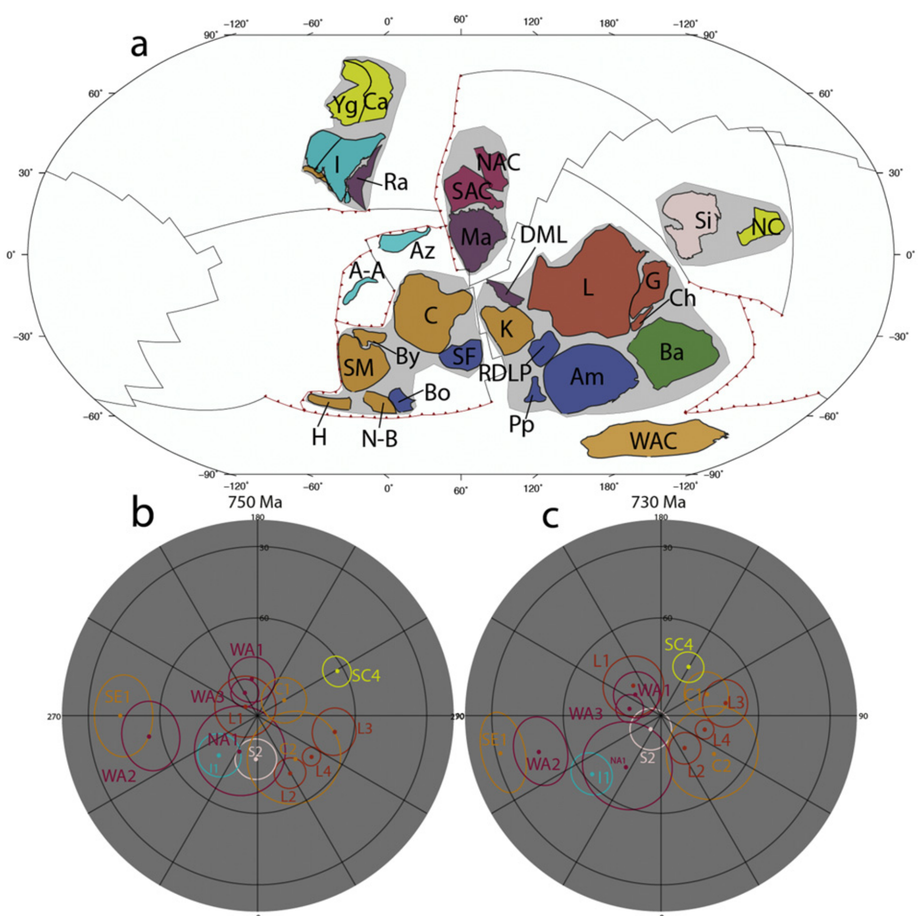

Fig. 10. (a) Tectonic geography at 750 Ma, see Figs. 5, 6 and 7 for abbreviations. (b) Palaeomagnetic poles around 750 Ma and (c) palaeomagnetic poles around 730 Ma.

used to compute a number of characteristics that describe the behaviour of plates and plate boundaries and these quantities for our Proterozoic reconstruction can be compared to models both for the Phanerozoic and for the present-day Earth. We divide the characteristics into two classes; those based on plate geometries, and those based on plate kinematics, the latter being more directly dependent on the motions of the continents. As shown by Domeier and Torsvik (2014), True Polar Wander (TPW) corrections will affect computed plate velocities, and our results here do not consider TPW effects. Considering the large overall uncertainties and that oceanic plates and mid-ocean ridges are artificial and inferred rather than observed, the plate model characteristics must be treated with a high level of caution. We thus offer the comments below as a way of comparison with Phanerozoic reconstructions (e.g. Domeier and Torsvik, 2014; Matthews et al., 2016), with the aim of illustrating the degree to which our reconstructions share similar characteristics, as well highlighting areas of future improvement. We extract velocity values only from cratonic crust and not from full topological plates due to the uncertainty surrounding pre-Pangaea oceanic domains, as well as absence of hard evidence of plate boundaries in the oceanic domain.

8.1. Plate geometries

8.1.1. Number and size of plates

The total number of plates modelled in the Neoproterozoic varies from six to twenty, with a smaller number of plates more common in the early Neoproterozoic, and a higher number more prevalent just prior to Gondwana amalgamation (Fig. 15a). The least number of plates occurs in the intervals 860 and 850 Ma and 820 and 800 Ma, both times during the life span of Rodinia. The largest number of plates occur in the transition between the two supercontinents (ca. 600 to 580 Ma). While this represents fewer plates than were modelled for the Palaeozoic (e.g. Domeier and Torsvik, 2014) and for the Cenozoic and Mesozoic (Mallard et al., 2016; Matthews et al., 2016; Morra et al., 2013, Fig. 15b), the change in number of plates associated with the supercontinent cycle is broadly similar. Unsurprisingly, the number of plates is lowest during times of supercontinent existence, (e.g. 1000–800 Ma and 560–520 Ma), while during dispersal and assembly the number is typically higher. Matthews et al. (2016) suggested that missing plates could be accounted for, in part, by the lack of regional analyses some areas receive, such that the evolution of back-arc basins and areas of regional accretion could account for a subset of missing plates. The model presented here is a simplification that excludes smaller micro-plate accretion and instead focusses on the major constituents of continental crust to produce a global model, within which, regional improvements can be anchored (and, ideally, improve the global model). Specifically, a number of tectonically complex areas that are likely made up of small plates were simplified (e.g. Hoggar Shield, ANS, Kara and Angara microterranes, Lhasa, Indochina). More complex tectonics in most, if not all, of these areas likely occurred through the entire supercontinent cycle (e.g. Caby, 2003; Fritz et al., 2013; Johnson et al., 2011, Metelkin et al., 2012; Vernikovsky et al., 2016; Robinson et al., 2014).

The comparison of plate size, and the relationship between this and the number of plates, is slightly more nuanced and informative about shortcomings in the model. The size of the largest plate during the Neoproterozoic is slightly larger than that seen during the past 200 Ma (e.g. Morra et al., 2013). From here, however, plate size decreases more rapidly than that predicted during younger times, and the trend identified by both Morra et al. (2013) and Matthews et al. (2016) of ~8 large plates, with the rest being of considerably smaller

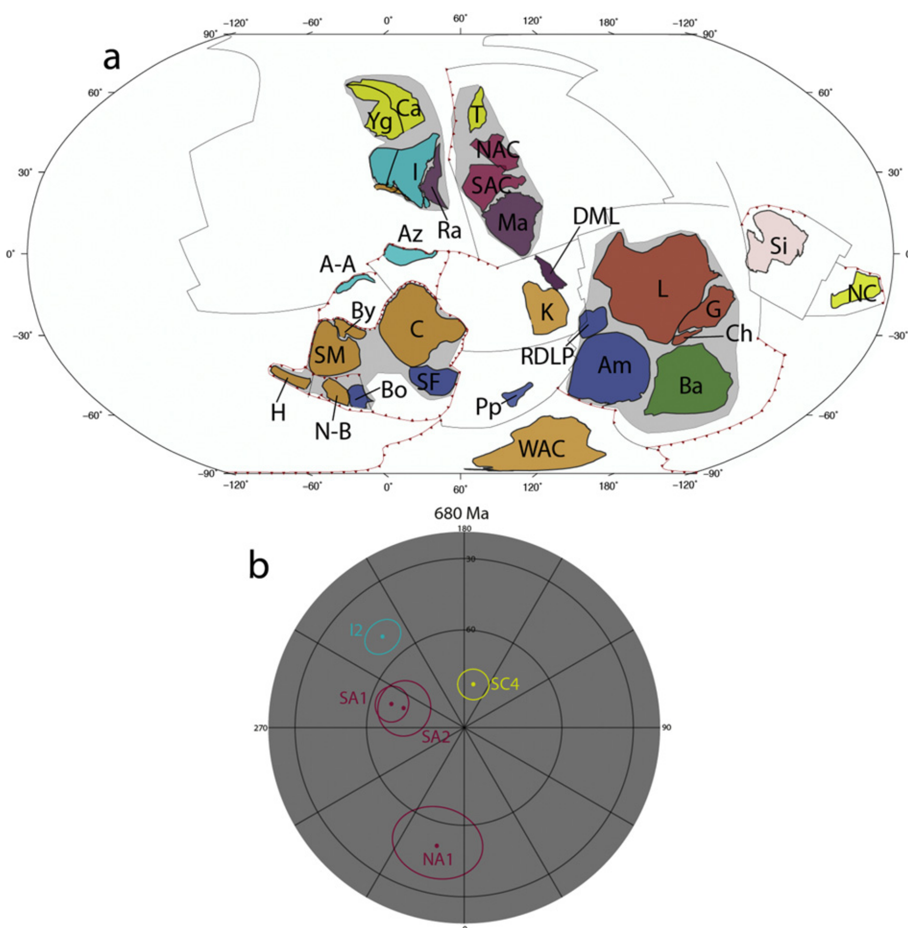

Fig. 11. (a) Tectonic geography at 680 Ma; T, Tarim, for other abbreviations see Figs. 5, 6 and 7. (b) Palaeomagnetic poles around 680 Ma.

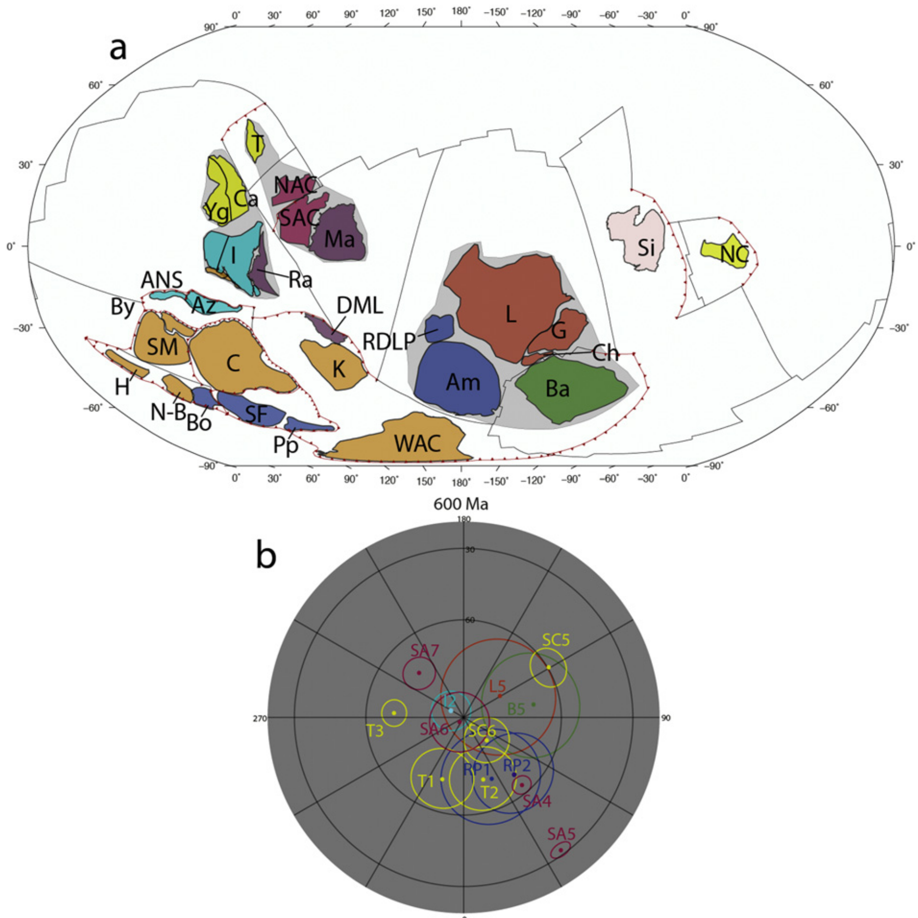

Fig. 12. (a) Tectonic geography at 600 Ma, see Figs. 5, 6, 7 and 10 for abbreviations. (b) Palaeomagnetic poles around 600 Ma.

size, is not observed here (i.e. there is no inflexion point after eight plates). For earlier times (1000–800 Ma) the inflexion point leading to a notable decrease in plate size occurs after the largest four plates. Comparably, in younger times (e.g. 700–520 Ma) the inflexion does not occur until the ultimate or penultimate smallest plate (i.e. there is a linear trend of cumulative plate count vs. plate size). We also find a similar trend with homogenous vs. heterogeneous tessellation of plates during the Neoproterozoic (Fig. 15c). Morra et al. (2013) proposed that immediately after supercontinent breakup the global system developed towards a homogenous organisation of plates, where the large plates were all of a similar size (i.e. a low standard deviation of plate size amongst the largest eight plates). This relaxes over 50–100 Myr to a heterogeneous distribution of plates with a high standard deviation more commonly associated with a growing supercontinent (i.e. one large plate). This is analogous to supercontinent breakup modelled here, as we 'unzip Rodinia' from the outside. Here, as one large supercontinent consisting of >80% of the available continental crust (heterogeneous tessellation) broke-up, smaller continental fragments rifted off incrementally at similar times from the outside (i.e. Australia first, then C-SF, then Kalahari), leading to a homogenous distribution of large plate sizes. Early in this process the bulk of the supercontinent remains (e.g. Fig. 9, Australia and Siberia have rifted off, but the rest of Rodinia is intact). Hence, the remaining supercontinent then still forms a large plate. As rifting and break-up continues, the size of the remnant supercontinent diminishes, and is taken up by other plates that continue rifting. For example, at 680 Ma (Fig. 11), Australia, C-SF, Kalahari, Siberia all occupy their own plates, and the remnant Rodinia is much smaller, now consisting only of Baltica, Laurentia and Amazonia, leading to plate sizes that are more similar.

The plate size and number of plates are inherently linked. We expect our oversimplified model to underestimate the number of plates, and that in more detailed future analyses the number of small plates in such a model will increase. Mallard et al. (2016), demonstrated that ongoing subduction at the margin of larger continental plates is a driver of plate fragmentation. Therefore, we expect that continent-margin subduction would create an ensemble of smaller plates. An example of this in the model may be the large oceanic embayment from 1000 to 850 Ma in Rodinia, just north of the WAC, where it joins the Congo Craton, which is here assigned to Rodinia, but could be subdivided into smaller plates (for example by further extending Azania).

8.1.2. Length of plate boundaries

We extracted the length of plate boundaries throughout the Neoproterozoic (Fig. 16) and compared them with estimates from Phanerozoic reconstructions. We evaluate the total length of subduction zones, and also the combined total length of mid ocean ridge (MOR) and transform boundaries - we refrain from specifically extracting ridge and transform lengths, due to the largely synthetic construction of divergent plate boundaries into ridge and transform segments within pre-Pangaea reconstructions.

The model presented here exhibits, on average, slightly less subduction than is evident at present-day, and ~60% of the total length of present-day MORs and transform boundaries. It is reasonable to consider our computed lengths as conservative estimates, as we do not include regional accretion models that may incorporate a series of back-arc basin opening and closures (e.g. Turaeg Shield, Avalonia, Cadomia, Western Ethiopian Shield, ANS), which account for a small amount of

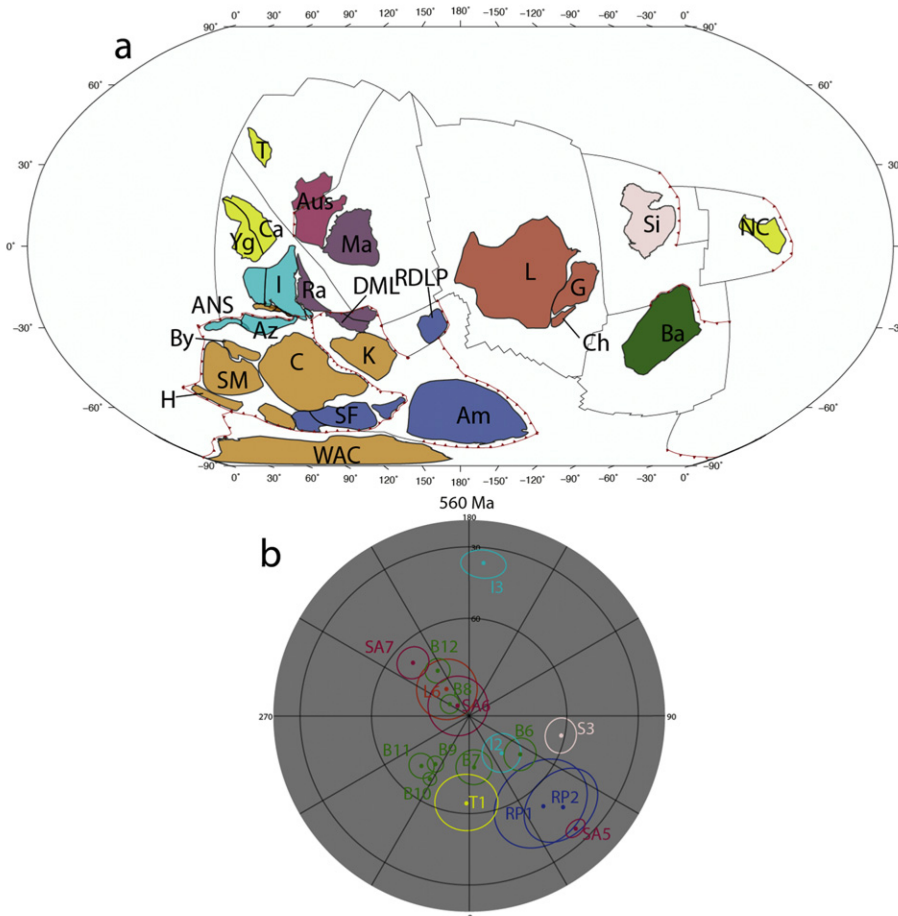

Fig. 13. (a) Tectonic geography at 560 Ma; Aus, Australia, for other abbreviations see Figs. 5, 6, 7 and 10. (b) Palaeomagnetic poles around 560 Ma.

global subduction, and we have little evidence to constrain oceanic-oceanic subduction zones.

The Neoproterozoic model has a similar length of plate boundaries to that of the Palaeozoic, and, in particular, the evolution of MORs and transform boundaries during the Neoproterozoic follows a similar, but slightly more protracted, trend (Fig. 16). There is a gradual increase in the length of MORS and transform boundaries until 600 Ma, reflecting the breakup of Rodinia and culminating with the opening of the Iapetus Ocean and the inversion of the ocean basins between the cratonic constituents of Gondwana (e.g. Fig. 11a), a pattern similar to the Mesozoic increase in total plate boundary length during the breakup of Pangaea and initiation of many new MORs. The peak in the mid Tonian is associated with the motion of the C-SF plate (e.g. Figs. 7a and 8a). The decrease post 600 Ma is due to the amalgamation of Gondwana, and is likely exacerbated by the uncertainty of the evolution of the Panthalassa Ocean, and could be rectified by models of the Early Palaeozoic that map the evolution of the Panthalassa Ocean backwards in time from the Late Palaeozoic.

Our length of subduction zones also shows a similar trend to the estimates of continental arc length presented by Cao et al. (2017), although these authors presented their synthesis of geological data using a different reconstruction model (Scotese, 2016). They depict a doubling of arc length from 750 to 550 Ma, where it drops abruptly (associated with collision of India and Africa), before dropping again at 530–520 Ma, associated with the collision of Australia and Gondwana (Figs. 11a, 12a and 16). Cao et al. (2017) record the length of continental arcs during the height of Gondwana building (650–550 Ma) as being roughly equal to the length of continental arcs at the present day. We find a similar relationship for the total subduction zone lengths within our model; from 750 to 550 Ma, the length of subduction zones double, with total subduction length between 600 and 550 Ma roughly the same length as total subduction at present-day (~55,000 km). Assuming that the proportion of continental vs. oceanic arcs remains stable over time, their measurements are, proportionally, in good agreement with our own.

8.2. Plate kinematics

8.2.1. Latitudinal distribution of subduction

The concept of inertial-interchange true polar wander (IITPW) describes the process where mantle instabilities arising from mass distributions at different latitudes, acting to reduce (or destabilise) the angular velocity of the earth on its spin axis, the outer layers of the earth (crust and mantle) can rotate relative to the spin axis (Evans, 2003; Goldreich and Toomre, 1969). Centrifugal forces due to the rotation of the Earth push positive inertia anomalies to the minimum moment of inertia (around the Equator). Two important processes that redistribute mass within the Earth are subduction, the process of introducing cold, dense material into the mantle, and plume upwellings, the process of pushing hot, buoyant material further away from the centre of the earth (Raub et al., 2007).

Raub et al. (2007) postulated that superplumes (like those arising from the thermal insulation of a supercontinent) would have a stronger effect on IITPW than individual sinking slabs due to the concentration of mass differentials. Hence, high latitudinal superplumes would encourage IITPW; as would extensive low latitude subduction if there was a high latitude positive anomaly as well. Conversely, low latitude superplumes stabilise the moment of inertia, negating IITPW. Critically, the thermal insulation that a supercontinent creates in the mantle (e.g. Ganne et al., 2016) will generate a positive inertia anomaly. Typically,

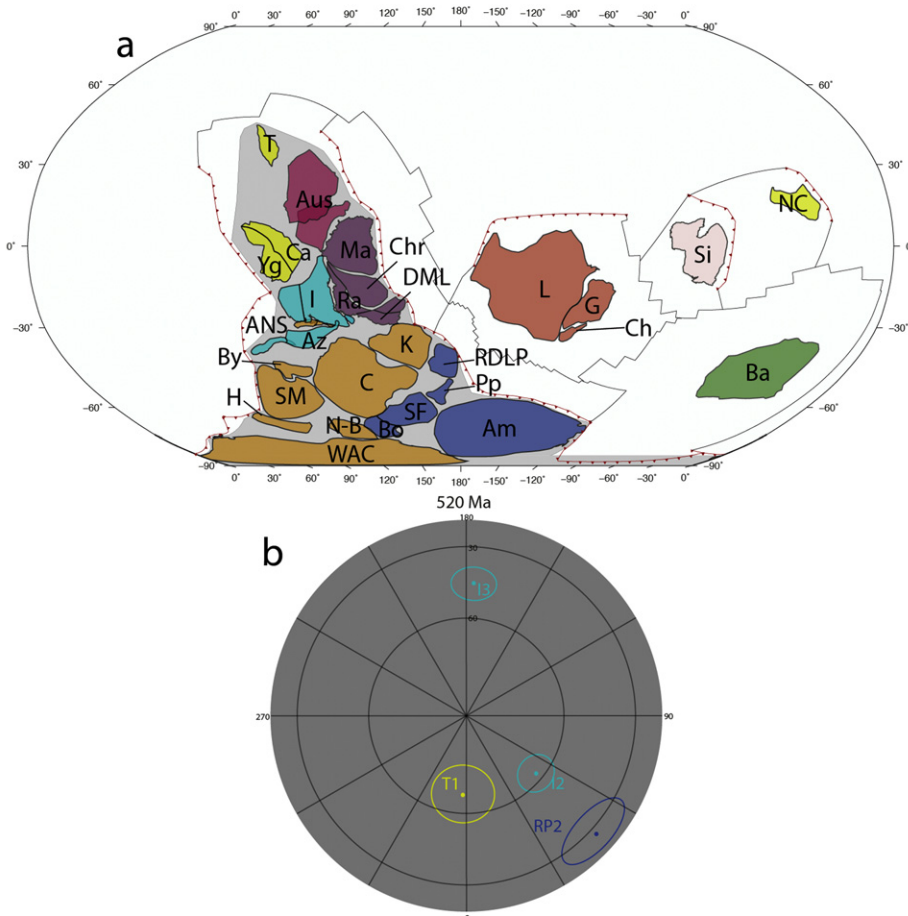

Fig. 14. (a) Tectonic geography at 520 Ma; Chr, Chron Craton, for other abbreviations see Figs. 5, 6, 7, 10 and 12. (b) Palaeomagnetic poles around 520 Ma.

TPW is inferred through measuring abrupt latitudinal changes of palaeomagnetic data within APWPs across the entire globe. Such episodes been interpreted in Phanerozoic times (e.g. Steinberger and Torsvik, 2008; Steinberger et al., 2017) and are suspected during the Neoproterozoic (e.g. Evans, 2003).

Our reconstruction provides an estimate the spatial and temporal distribution of subduction zones, (albeit with the caveats concerning interpolation of plate boundaries outlined previously), which may hold clues to redistributions of mass within the deep Earth and related true polar wander episodes. The latitudinal distribution of subduction zones through the Neoproterozoic is shown in Fig. 17. Low latitude subduction is prevalent throughout the Tonian, evident in the equatorial Valhalla Orogeny and early subduction in Azania and the ANS, and by the mid-latitude accretion of India-South China. In particular, there is a relative dearth of high latitude subduction between 850 and 700 Ma, one of the key times postulated to be affected by IITPW (e.g. Evans, 2003; Li et al., 2004).

Li and Zhong (2009) proposed that a scenario where a supercontinent that is surrounded by a girdle of subduction can generate antipodal superplumes, one of which will form beneath the supercontinent itself. Superplumes forming beneath a supercontinent at high latitudes (e.g. Li et al., 2008) would be more likely to generate episodes of IITPW. Our model omits both South China and India from Rodinia (and thus any requirement to fit Rodinia to palaeomagnetic data from these regions), with the consequence that Rodinia lies at low latitudes throughout the time that any superplume may have developed beneath it. Palaeomagentic constraints on the Congo have been respected by not having Rodinia form until ca. 750 Ma, immediately before it begins to break-up. The geological evidence used to argue for the existence of a superplume in Rodinia between 850 and 750 Ma relate predominantly to dyke swarms throughout Australia, eastern Laurentia and South China, bimodal magmatism in South China and magmatic activity throughout India (Li et al., 2003, 2008). As our model separates India and South China from Rodinia, we suggest that the magmatic record through this period needs to be objectively investigated to ensure that a 'smoking-gun' for the proposed superplume exists. We also suggest that a significant amount of this magmatism should be within the core of Rodinia (east Australia, west Laurentia, south Congo, Kalahari, Rio de la Plata), rather than on the periphery where voluminous subduction-related arc magmatism occurred.

During the Ediacaran, nearly every piece of known continental crust, along with nearly all subduction zones constrained by observations from the geological record, were in the southern hemisphere (e.g. Figs. 13a, 14a and 17). Given the conventional interpretation of palaeomagnetic data and the position of continents, the northern hemisphere of the globe, and particularly the medium-high latitudes, consisted primarily of oceanic plates at this time. While determining the nature of IITPW during this time is elusive, in part, because of discrepancies in palaeomagnetic data (e.g. the 'Ediacaran nightmare' of Baltica and Laurentia; Meert, 2014b, compared to Australia which has a relatively consistent APWP that does not seem to exhibit IITPW; Schmidt, 2014), the shift towards a monohemispheric distribution of continental crust and subduction zones, creates implications for both the mass distribution of the Earth and, consequently, TPW.

8.2.2. Plate velocities

The concept of a 'plate tectonic speed limit' has been proposed for plate motions since the Jurassic (e.g. Zahirovic et al., 2015), although it is less certain whether the same limits apply for earlier times (e.g. Gurnis and Torsvik, 1994; Meert et al., 1993). Limits derived from reconstruction

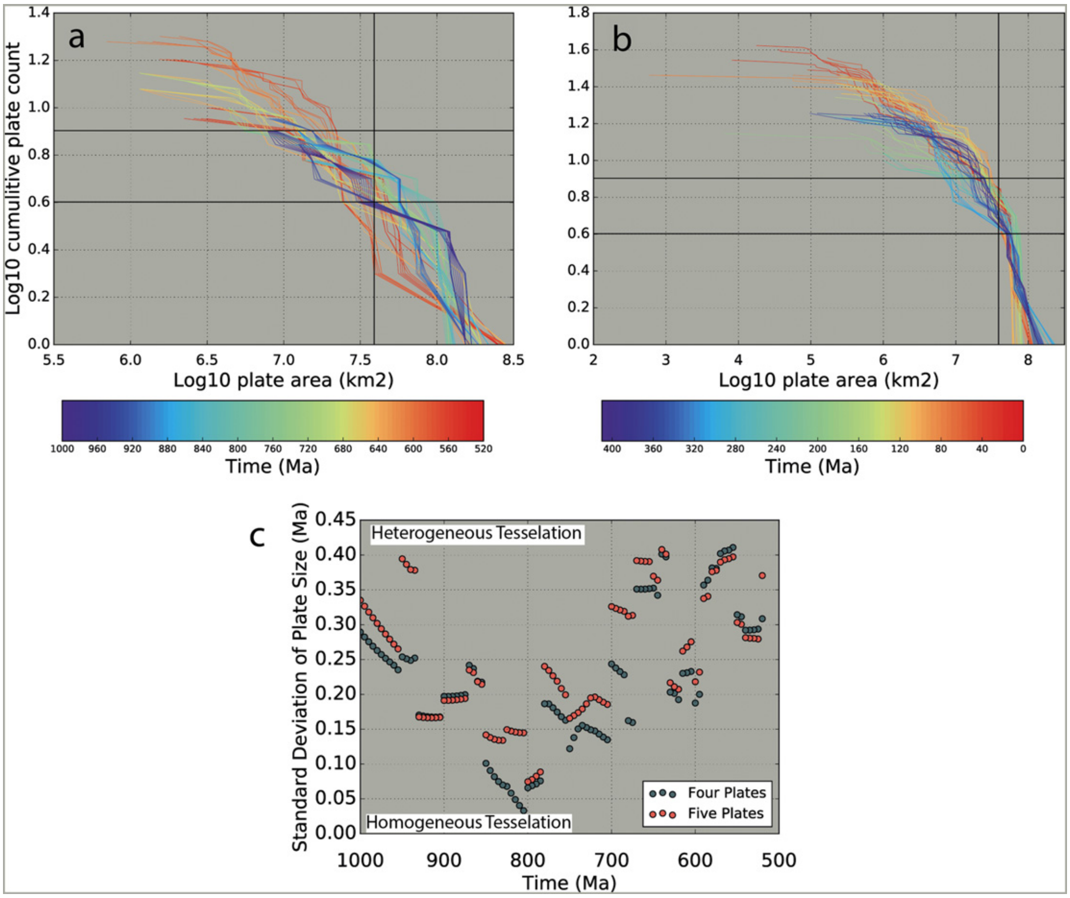

Fig. 15. Log_{10} of plate size vs. log_{10} of cumulative plate count (from largest to smallest) for; (a) Neoproterozoic model presented here; (b) the Late Palaeozoic, Mesozoic and Cenozoic after Matthews et al. (2016); and (c) standard deviation of the largest four and five plates vs. time for the Neoproterozoic model. Straight black lines in (a) and (b) represent markers for $log_{10}(4)$ and $log_{10}(8)$ (horizontal lines) and the vertical line represents the cut off of area for large vs. small plates after Mallard et al. (2016) (7.59 on the scale). Sizes are extracted at 5 Myr intervals.

models of the Cenozoic and Mesozoic suggest a maximum velocity of ~200 mm/yr for plates consisting of less than 50% continental crust, with cratonic and continental crust typically moving no faster than ~150 mm/yr (Zahirovic et al., 2015). Furthermore, Root-Mean Square (RMS) velocities of plates consisting of >25% cratonic crust are ~28 mm/yr, and plates with any amount of continental crust have a RMS velocities of ~58 mm/yr. Models of the Palaeozoic (Domeier and Torsvik, 2014), Matthews et al. (2016) indicate that the global RMS velocity of continental crust peaked at ~140 mm/yr at ca. 360 Ma, but otherwise averaged around ~100 mm/yr. Fig. 18 depicts the RMS velocities of continental crust extracted from the model for the Neoproterozoic. The model presented here maintains an average RMS velocity for continental crust of ~37 mm/yr, roughly in line with plate motions over the past 200 Ma.

The lower global RMS velocity that we have extracted from the Neoproterozoic compared to the Palaeozoic may be a consequence of both the fewer amount of palaeomagnetic data available for the Neoproterozoic, as well as a lesser quality of poles from the available poles (e.g. Domeier, 2016; Domeier and Torsvik, 2014, are based on a compilation of Torsvik et al., 2012, consisting of 150+ quality filtered

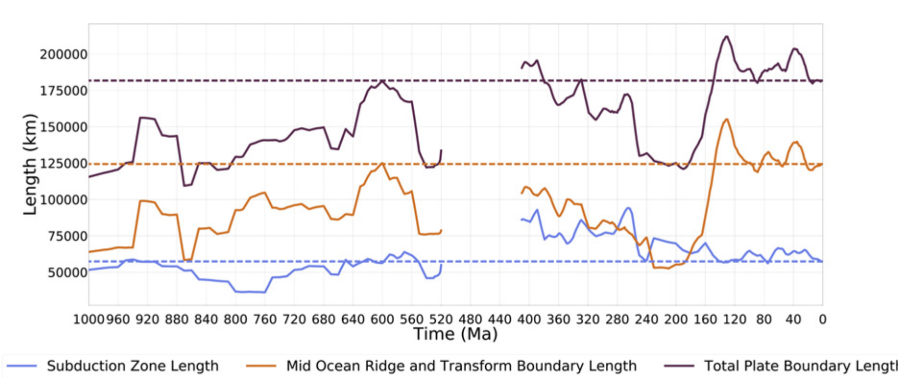

Fig. 16. Length of plate boundaries, extracted at 1 Myr intervals and calculated as 10 Myr rolling average. The present-day length of plate boundaries are plotted as straight dashed lines. Boundary lengths are from the model presented here and the model from Matthews et al. (2016), which is a compilation of the Domeier and Torsvik (2014) and Müller et al. (2016) models.

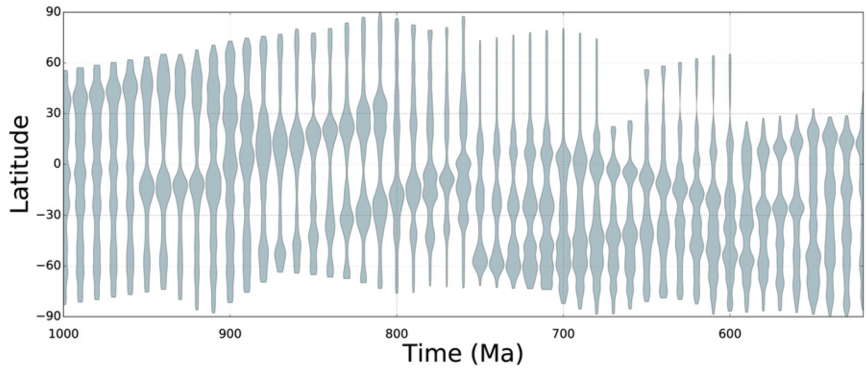

Fig. 17. Latitudinal distribution of subduction zones through the Neoproterozoic.

poles, comparatively we use 51). In particular, omitting poles from Baltica and Laurentia in the Late Ediacaran and Cambrian (Section 6.1) simplified the motions of the both plates. A stricter fitting of palaeomagnetic poles here would dramatically increase the distance that each continent moved, increasing their velocity.

Peaks of global RMS velocity occur following the rifting of plates that have a significant constituent of inferred oceanic crust. For example, Australia and Siberia, both moving relative to Laurentia from 800 Ma, have significant lengths of modelled oceanic subduction along their leading edges (as there is no preserved subduction on the appropriate continental margins) (e.g. Figs. 9a and 10a). Similarly, the increase of RMS velocity at 700 Ma is associated with the rifting of Kalahari and the initiation of the closure of the Khomas Ocean, where a tectonic plate with a leading edge of oceanic crust was subducted (e.g. Figs. 11a and 12a). Comparatively, the rifting of C-SF-SM at 750 Ma had limited oceanic crust due to (preserved) subduction on its leading edges (e.g. ANS, SM, Hoggar region, Figs. 10a and 11a), so there is no increase in RMS velocity. The peak at ca. 670 Ma is likely the result of inferred southerly subduction of India and closure of the Mozambique and Neomozambique Ocean (e.g. Fig. 11a), where the lead of the plate also consisted of oceanic crust, perhaps analogous to the closure of the Neo-Tethys in the Cenozoic.

We also extracted the velocity from centroid points of individual cratons (Fig. 19a–e), to more clearly identify phases of rapid continental motion and discuss how well constrained these phases of motion are. Generally, the velocities are less than 80 mm/yr, although there are some exceptions where individual plate velocities extend up to ~120 mm/yr. For example, the Congo Craton (and Borborema, São Francisco, the SM, Hoggar, Nigeria-Benin and Bayuda blocks, which are children of Congo within the plate hierarchy) exhibits a phase of rapid motion between 780 and 750 Ma. Although the velocity is high for such a large plate (~120 mm/yr), the speed is required to satisfy palaeomagnetic data and known geological constraints, though dating on many of the sedimentary sequences is poor (Section 3.5). Shifting the rifting time to before 900 Ma, with a transition to drift by 850 Ma (still permitted by palaeomagnetic data), would reduce the velocity, however, we have opted to constrain the movement to a time coeval with the onset of the Imorona-Itsindro subduction zone outboard of Azania (Handke et al., 1999; Archibald et al., 2016, in press).

Siberia and North China (also linked in the reconstruction hierarchy) record a phase of rapid motion at a similar time (ca. 780 Ma) (Fig. 9a). The rapid Siberia and North China motion (~110 mm/yr) follows their displacement along the (present-day) northern margin of Laurentia as a response to the 800 Ma rifting of Australia from the west coast of Laurentia. This motion is constrained by palaeomagnetic data (e.g. Pisarevsky et al., 2013) suggesting that Siberia was located closer to Laurentia at 760 Ma than in the Tonian (e.g. Figs. 9a and 10a). As with C-SF, earlier rifting of Siberia would reduce the velocity. The location of the Euler Pole for the rifting of Australia from Laurentia, suggests that rifting initiated to the (present-day) north of Australia, closer to Siberia, and moved south as it progressed (i.e. the fan-like collapse). This phase of rapid motion could be reduced by modelling the opening of the ocean basin between Australia and Laurentia further north (ca. 825 Ma), outboard (and west) of Siberia, then propagated southwards to initiate rifting of Australia from Laurentia by 800 Ma (Fig. 9a).

Rapid phases of motion are modelled for both India-South China-Rayner (100 cm/yr at ca. 650 Ma), and the RDLP (120 mm/yr at ca. 520 Ma), both implied within our model to result from a similar mechanism (Figs. 13a and 14a). In both cases, the cratons are initially separated from a growing supercontinent by an ocean basin; then, subduction initiates around the supercontinent periphery resulting in closure of the ocean basins and pulling the smaller continental blocks towards Gondwana. In both cases, the leading edge of the plate is inferred to consist of a significant area of oceanic crust, similar to the evolution of India in the Cenozoic. This creates a strong pulling force and can explain the faster than expected motions. Finally, Baltica exhibits extremely fast motions in the latest Ediacaran and early Cambrian (Fig. 14a); we note that this is partly a reflection of uncertainties with palaeomagnetic data from both Baltica and Laurentia during this time and the subsequent uncertainty about their position. For example, Domeier (2016) depicts a high latitude Baltica at 500 Ma outboard of Gondwana (our 520 Ma position is in transition to this position), but this necessitates a fast transition and rotation from the reliable 560 Ma poles that place the continent at a lower latitude.

9. Future work

The study that we present here is, by necessity, incomplete and therefore, represents just a first approximation to the true tectonic geography of the nearly half a billion years of Earth history represented by the Neoproterozoic. We do assert, however, that this methodology is an important way to test the numerous speculated links between Neoproterozoic plate tectonics and the greater Earth system. We also confident that this is the most complete attempt to date at merging

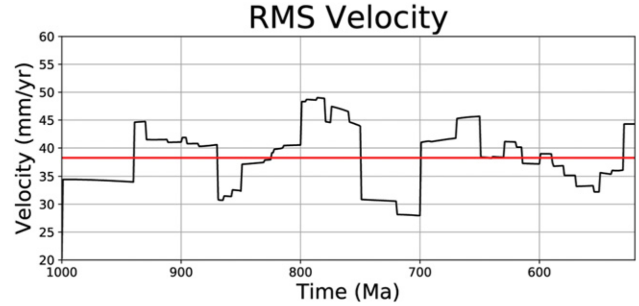

Fig. 18. RMS velocity of continental crust in the Neoproterozoic. The black line is the average from all continents extracted at a 5 Myr interval. The red line is the average across the Neoproterozoic.

tectono-geographic data derived from geology with palaeomagnetic data for the Neoproterozoic and, to our knowledge, this is the first attempt at amalgamating these into a full globe, whole-plate topological model of the planet over this critical period of Earth history. We hope that this will stimulate many improvements to palaeogeographical reconstructions over the next few years.

Beyond the gathering of more (and better quality) palaeomagnetic and geological data, we identify three key steps that should be undertaken in the future to improve the reconstruction presented here. Firstly, we are conservative with the amount of subduction occurring during the Neoproterozoic, relative to what we know occurs at the present-day. Assuming that Neoproterozoic subduction systems were similar to the modern Earth, there is a considerable scope for uncovering other subduction systems during this time. Part of this will be through developing more tectonically complex and complete regional models for key areas (e.g. Azania, SM, ANS and Siberia) that more closely approximate the spatial and temporal constraints of arc collision, back-arc rifting and subduction initiation. This could be achieved by the coupling of models with whole earth system databases, such as a global geochemistry database. Secondly, the development of alternate models to test end member scenarios is critically important to completely evaluate the reliability of this (and any other model), especially with respect to times with more accessible to data (e.g. 650–500 Ma). Finally, the coupling of this reconstruction to younger models for the Phanerozoic

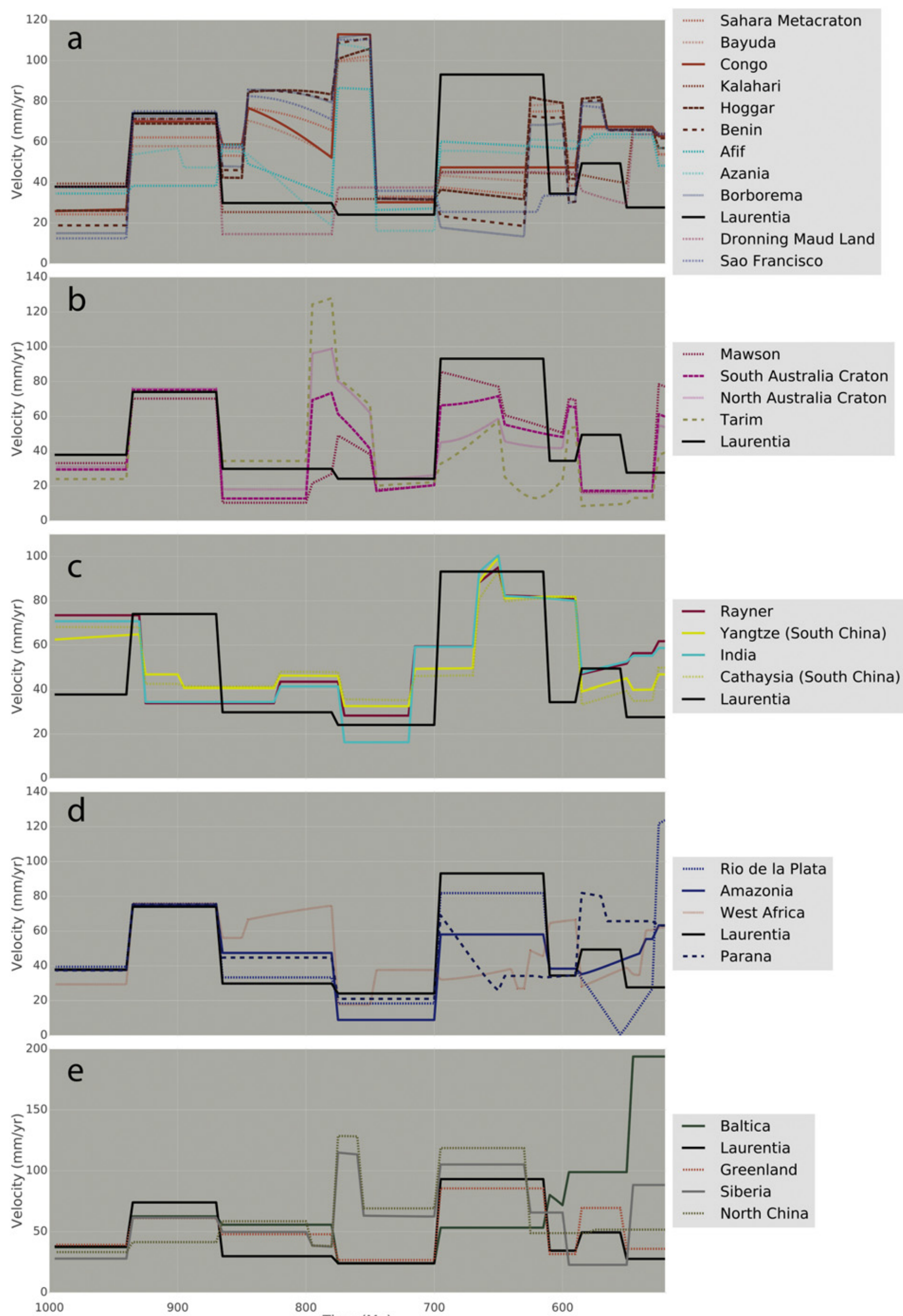

Fig. 19. Velocity of continental crust fragments in the Neoproterozoic. Velocities were calculated from the centroid of each portion of continental crust. Each panel broadly reflects a common Neoproterozoic journey; (a) Congo-São Francisco and the Sahara Metacraton (extra south Rodinia); (b) Australia and Mawson (west Rodinia); (c) India and South China; (d) Amazonia (east Rodinia); (e) Laurentia and northern Rodinia. Colour is based on present day geographical position; red – North America; dark blue – South American; green, Europe; grey, Siberia; light blue, India, Madagascar and the Middle East; yellow, China; purple, Australia and Antarctica. Laurentia is plotted in black in each panel as it is considered the heart of Rodinia.

paves the way for the first full plate models that encompass entire supercontinent cycles and will be instrumental for further understanding the long-term nature of the mantle. So, a priority and future studies should prioritise linking Neoproterozoic reconstructions with the younger Palaeozoic models (e.g. Domeier, 2016; Domeier and Torsvik, 2014).

10. Conclusions

The Neoproterozoic was dominated by a supercontinent cycle beginning with the formation of Rodinia, its subsequent breakup, and the consequent amalgamation of Gondwana. Here we have presented the first full plate global reconstruction of the Neoproterozoic that synthesises both palaeomagnetic and geological data with plate tectonic theory to model the evolution of plate boundaries and plates during this time. We present an updated set of Euler rotations for the kinematic motion of these plates during the Neoproterozoic. In particular, we present a new model for India and South China that reconciles the similar accretionary histories of the two cratons in the Tonian and fits the sparse palaeomagnetic data available from both cratons. This places South China in a position similar to its Gondwana position in the Cambrian, just north of India. A new model for the Congo-São Francisco craton is also proposed that ties disparate palaeomagnetic data (from the Tonian and Cryogenian) with a rifting event and counter-clockwise rotation of C-SF relative to Rodinia, helping explain the thick mid-Tonian sedimentation evident on the southern margin of Congo and eastern margin of the WAC, 50 Ma before the earliest inferred Rodinia rifting.

A key contribution of the reconstruction presented here is to model plate boundaries through time, and, by extension, model tectonic plates. From this we are able to extract basic data about the geometries and kinematics of these boundaries and plates, and make decisions about plate motions that minimise factors such as velocity, whilst also fitting palaeomagnetic and geological constraints. Our model maintains an average RMS velocity of continental crust during the Neoproterozoic of ~3.8 cm/yr.

The methodology has also allowed us to investigate the lengths of plate boundaries through time. We note that, relative to present-day, we have underestimated the length of subduction zones by 5000 to 10,000 km. This probable underestimation is likely due to poorly preserved subduction missing from the geological record (e.g. ocean-ocean convergence), and the oversimplification of key areas of subduction in the Neoproterozoic that we incorporated into this initial model. We expect that this model will be useful and act as a foundation, for future regional studies in poorly constrained areas of the Neoproterozoic, as well as providing an essential plate tectonic model that can be used to investigate the link between plate tectonics and other Earth systems, such as climate, biological evolution, oceanography and mantle geodynamics.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gr.2017.04.001.

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