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Amalgamating eastern Gondwana: The evolution of the Circum-Indian Orogens

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Abstract

The Neoproterozoic global reorganisation that saw the demise of Rodinia and the amalgamation of Gondwana took place during an incredibly dynamic period of Earth evolution. To better understand the palaeogeography of these times, and hence help quantify the interrelations between tectonics and other Earth systems, we here integrate Neoproterozoic palaeomagnetic solutions from the various blocks that made up eastern Gondwana, with the large amount of recent geological data available from the orogenic belts that formed as eastern Gondwana amalgamated. From this study, we have: (1) identified large regions of pre-Neoproterozoic crust within late Neoproterozoic/Cambrian orogenic belts that significantly modify the geometry and number of continental blocks present in the Neoproterozoic world; (2) suggested that one of these blocks, Azania, which consists of Archaean and Palaeoproterozoic crust within the East African Orogen of Madagascar, Somalia, Ethiopia and Arabia, collided with the Congo/Tanzania/Bangweulu Block at ~650–630 Ma to form the East African Orogeny; (3) postulated that India did not amalgamate with any of the Gondwana blocks until the latest Neoproterozoic/Cambrian forming the Kuunga Orogeny between it and Australia/Mawson and coeval orogenesis between India and the previously amalgamated Congo/Tanzania/Bangweulu–Azania Block (we suggest the name 'Malagasy Orogeny' for this event); and, (4) produced a palaeomagnetically and geologically permissive model for Neoproterozoic palaeogeography between 750 and 530 Ma, from the detritus of Rodinia to an amalgamated Gondwana.

1. Introduction

The amalgamation of Gondwana took place during one of the most dynamic periods known in the evolution of the climate (Hoffman et al., 1998; Evans, 2000), of life (Valentine, 2002), and possibly also the deep Earth (Kirschvink et al., 1997; Evans, 1998). The probability that these events are linked (e.g. Brasier and Lindsay, 2001) and a growing realisation that many of these systems are, in part, controlled by the distribution of continents, mountain ranges and

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oceanic basins has created a need for a better understanding of the geography of the Neoproterozoic world. Associated with this is the importance of Neoproterozoic tectonics on the location and timing of successor Phanerozoic orogenic belts (Boger and Miller, 2004; Cawood, 2005); the present distribution of tectonic belts being largely controlled by past plate configurations.

Eastern Gondwana makes up much of the continents of Australia, India and East Antarctica (Fig. 1). During the Neoproterozoic, these continents came together with the constituent continental blocks that now make up Africa and South America, to form the supercontinent Gondwana (or Gondwanaland, Du Toit, 1937). In this study, we integrate palaeogeographic data, in the form of palaeomagnetic solutions, with geological information from the collision zones that separate the various continental blocks to better delineate the various component continents, constrain the time of amalgamation throughout the system, and generate a permissive tectonic model for the evolution and progressive growth of eastern Gondwana.

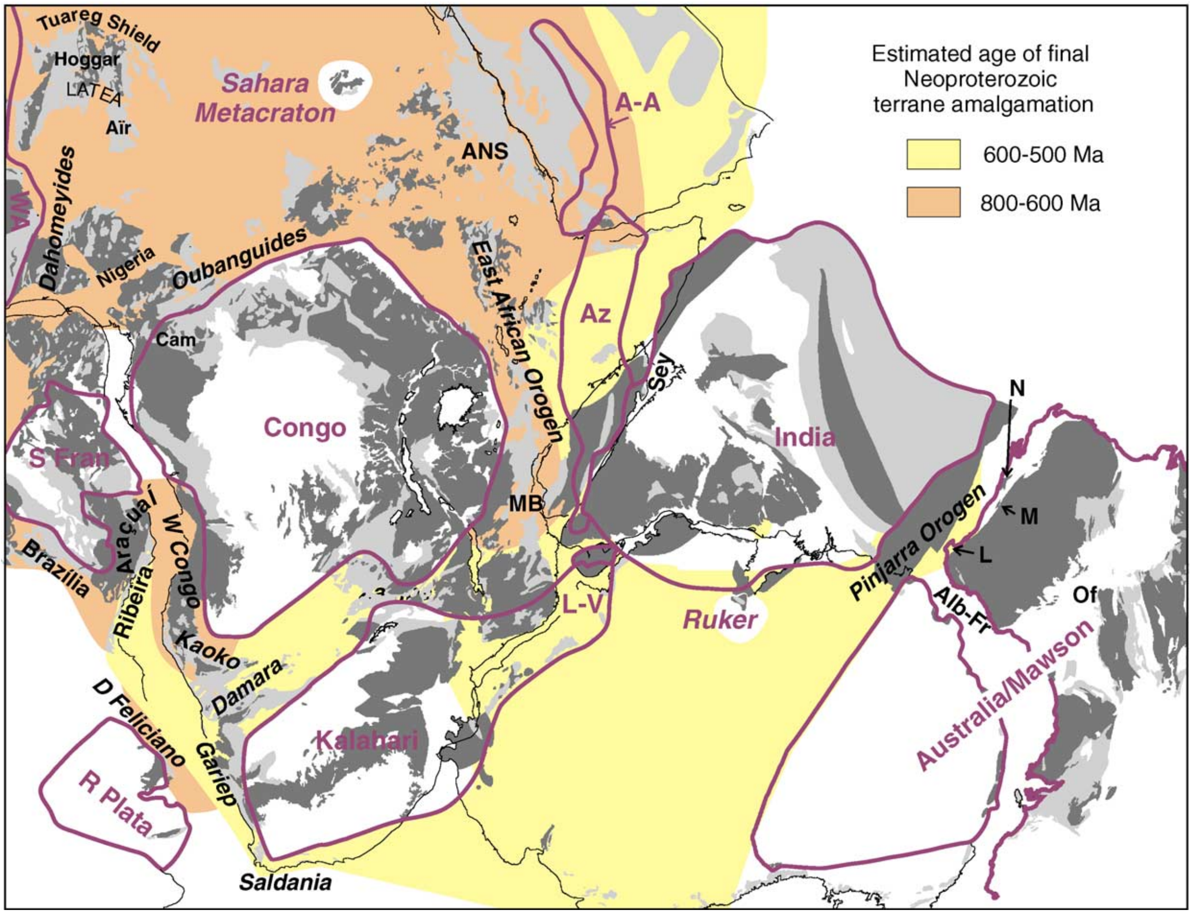

Fig. 1. Geological outcrop map of south, central and northeast Africa, India, East Antarctica and Australia (based on the data sets of IGCP 288, Wolmarans, 1999) rotated into the Gondwana fit of Reeves and De Wit (2000). Precambrian outcrops older than 1000 Ma shown as dark grey, those younger than 1000 Ma shown as light grey. Outlines of the Neoproterozoic continental blocks discussed in this paper and used in Figs. 2–6 are marked in purple. Block abbreviations: A-A = Afif–Abas Terrane; Az = Azania; Congo = Congo/Tanzania/Bangweulu Block; L-V = Lurio–Vijayan Peninsula; R Plata = Rio de la Plata Block; Ruker = Ruker Terrane, Southern Prince Charles Mountains; S Fran = Sâo Francisco Block; WA = West African Block. Orogenic belt and location abbreviations: Alb-Fr = Albany–Fraser Orogen; ANS = Arabian–Nubian Shield; D Feliciano = Dom Feliciano Belt; Cam = Cameroon; L = Leeuwin Complex; M = Mulingarra Complex; MB = Mozambique Belt; N = Northampton Complex; Of = Officer Basin, Sey = Seychelles.

2. Previous models of eastern Gondwanan amalgamation

Early plate-tectonic Proterozoic palaeogeographic models interpreted the limited palaeomagnetic data as supporting the presence of a single super-continent throughout the Proterozoic (Piper, 1976) that was later called Palaeopangea (Piper, 1982, 2000). McWilliams (1981) challenged this view and suggested that the data were more consistent with two Neoproterozoic continental masses, East Gondwana (India–Australia–Antarctica) and West Gondwana (Africa–South America) that collided along the Mozambique Belt to form Gondwana (Fig. 1). In this model, Neoproterozoic India was intimately associated with Australia and Antarctica and any Neoproterozoic orogenesis between them was seen as intra-cratonic (e.g. Harris, 1994; Wilde, 1999).

The suggestion that both East and West Gondwana did not exist as Neoproterozoic supercontinents in their own right, but that the constituents of eastern and western Gondwana came together during the Neoproterozoic is a recent development driven initially by palaeomagnetic considerations (Meert et al., 1995), and more recently supported by reinterpretations of the geological data. In the case of East Gondwana, Fitzsimons (2000a,b) showed that the Rayner Complex of East Antarctica and the Eastern Ghats of India preserved a considerably different history to the Albany-Fraser Orogen of Western Australia. The two orogens had formerly been correlated as part of the Late Mesoproterozoic/Early Neoproterozoic Circum-East Antarctic Orogen (Katz, 1989; Hoffman, 1991; Moores, 1991; Yoshida, 1995), but are separated by the Pinjarra Orogen (Prydz–Denman–Darling orogen of Fitzsimons, 2000b; Prydz–Leeuwin Belt of Veevers, 2000, 2004) (Fig. 1) that would bisect a coherent East Gondwana. The idea that this orogen was an intra-continental Neoproterozoic orogen (Harris and Beeson, 1993; Harris, 1994) was challenged by Hensen and Zhou (1997), who pointed out that metamorphism in the Antarctic extension of the Pinjarra Orogen was similar to that from collisional orogens (Fitzsimons, 1996; Carson et al., 1997; Boger et al., 2000). Fitzsimons (2000b) linked this with the change in age provinces across the orogen to suggest that the Pinjarra Orogen, and its Antarctic continuation, represents the Neoproterozoic collision zone between India and Australia. Palaeomagnetic data from India and the Seychelles support this geological interpretation (Torsvik et al., 2001a,b) as they suggest that India and Australia travelled separately during the Neoproterozoic (Powell and Pisarevsky, 2002; Pisarevsky et al., 2003).

A similar situation occurs along the Zambezi Belt, separating the Kalahari and Congo/Tanzania/Bangweulu Blocks (Fig. 1). Here the Choma–Kalomo Block south of the Zambezi Belt has long been thought of as a southern extension of the Mesoproterozoic Irumide Belt (Hanson et al., 1994; Wilson et al., 1997). If this correlation holds, it follows that the Kalahari and Congo/Tanzania/Bangweulu Blocks must have been together since Mesoproterozoic times and that Neoproterozoic deformation in the Zambezi Belt must only represent intra-continental orogenesis. However, recent U–Pb ion probe dating has shown that the correlation of the Irumide Belt and Choma–Kalomo Block is not valid (De Waele et al., 2003c). Eclogites, arc-volcanics and ophiolites from the Zambezi Belt demonstrate that Mesoproterozoic to Neoproterozoic oceanic-crust existed between the Congo/Tanzania/Bangweulu and Kalahari Blocks (Oliver et al., 1998; Johnson and Oliver, 2000, 2004; John et al., 2003). Closure of this ocean (the Chewore Ocean of Oliver et al., 1998) and collision of Kalahari and Congo/Tanzania/Bangweulu has been interpreted to correlate with 560–510 Ma (Vinyu et al., 1999; Hargrove et al., 2003; John et al., 2003; Johnson and Oliver, 2004) tectonism and high-pressure metamorphism (Johnson and Oliver, 1998, 2002; John et al., 2003), which is similar to the time of metamorphism in the adjacent Mozambique Belt of Malawi (Kröner et al., 2001; Ring et al., 2002). Further west, peak-metamorphism in the Damara Belt is dated to 535–505 Ma (Jung et al., 2000; Goscombe et al., 2004) and is also correlated with incorporation of Kalahari into Gondwana. Foreland basins on the western Kalahari and the Congo/Tanzania/Bangweulu margins preserve distinct records of the timing of basin formation (Prave, 1996). The foreland basin on the west Congo/Tanzania/Bangweulu Block margin developed at 750–600 Ma, whereas that on the western Kalahari margin was delayed until ~550 Ma (Prave, 1996). These observations suggest diachronous orogenesis along southwestern African and support the idea of the Neoproterozoic Kalahari Block being separate from the Congo/Tanzania/Bangweulu

Block and amalgamating with a nascent western Gondwana significantly later than the Congo/Tanzania/Bangweulu–Amazonia collision (Prave, 1996).

3. Palaeomagnetic evidence

Reliable palaeomagnetic information exists for India, Australia, and the Congo in the 750–530 Ma interval (Figs. 2–6, Table 1). According to these data, at ~750 Ma Australia and Congo were in low latitudes and India was in middle latitudes (Fig. 2). Australia is positioned at 755 Ma by the Mundine Well dykes palaeopole (Wingate and Giddings, 2000). The Malani Igneous Suite palaeopole determines the position of India at 770–750 Ma (Torsvik et al., 2001b). In its reconstructed position relative to India, the Seychelles palaeopole, determined from granites precisely dated at 755–748 Ma, overlaps the Malani palaeopole (Torsvik et al., 2001a). The Mbozi Com-

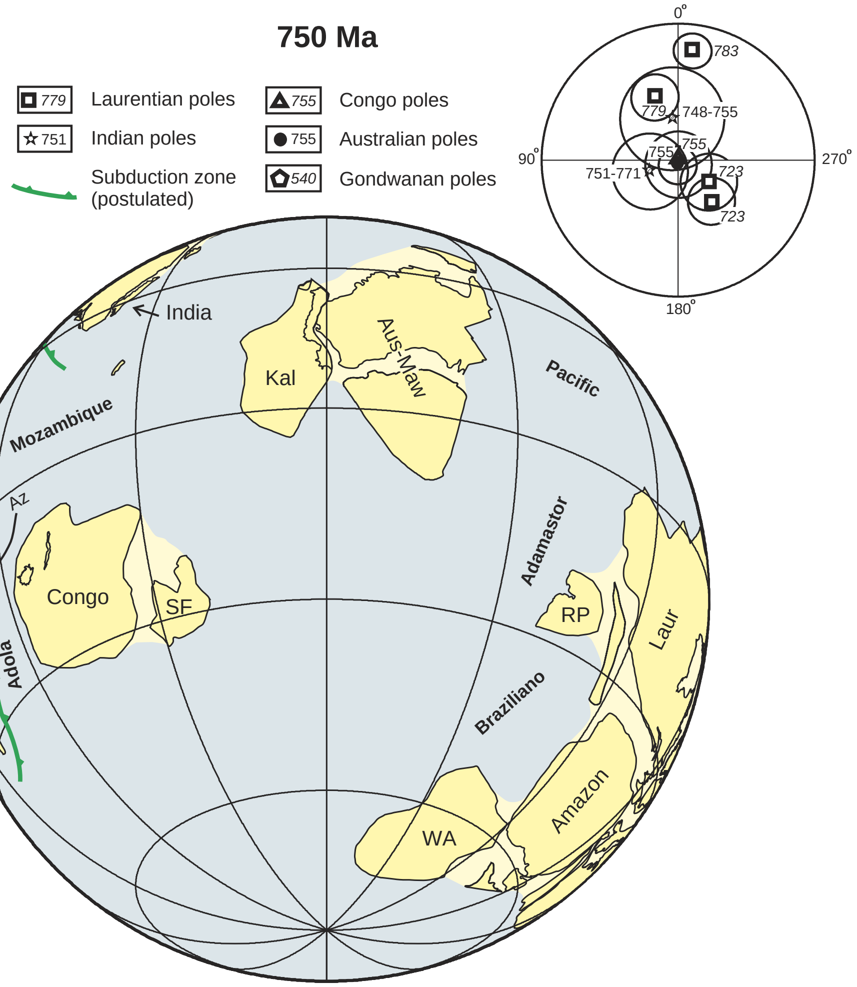

Fig. 2. Palaeogeographic reconstruction at 750 Ma. Deep yellow = continental plates delineated from map, pale yellow = hypothesized extend of continental crust. Amazon = Amazonia; Aus-Maw = Australia/Mawson Block; Az = Azania; Congo = Congo/Tanzania/Bangweulu Block; Kal = Kalahari Block; Laur = Laurentia; RP = Rio de la Plata; SF = São Francisco; WA = West Africa. Adola, Adamastor, Braziliano, Mozambique, Pacific = oceanic basins. A polar 30° latitudinal circle is presented with relevant palaeomagnetic solutions rotated into congruence with the reconstruction—see Table 1 for details and Table 2 for rotation parameters. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

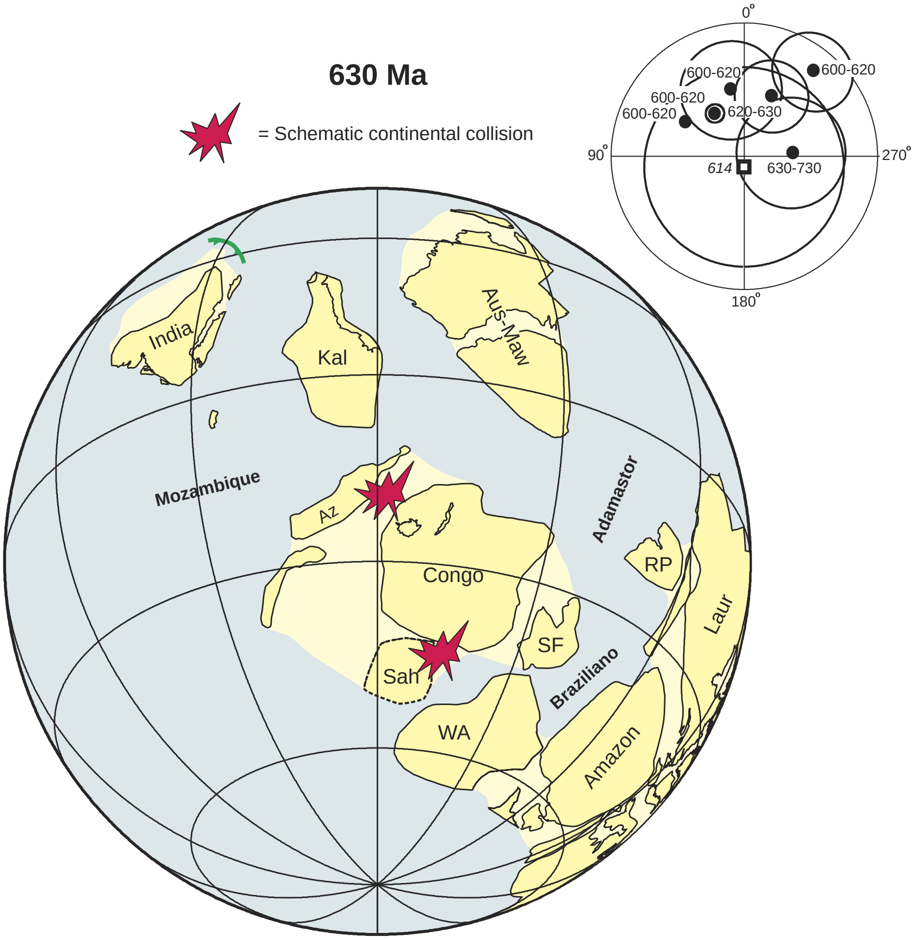

Fig. 3. Palaeogeographic reconstruction at 630 Ma—key and abbreviations the same as for Fig. 2, except: Sah = Sahara Metacraton.

plex palaeopole in the Congo (Meert et al., 1995) has a less well determined age of 755 ± 25 Ma (Evans, 2000, recalculated age).

India and the Congo also have palaeopoles at ~800 Ma. In India, the Harohalli dykes give a palaeopole with an imprecise age of 814 ± 34 Ma (Radhakrishna and Mathew, 1996). In the Congo, Meert et al. (1995) determined a palaeopole from the Gagwe lavas, whose age has recently been reassessed as 795 ± 7 Ma (Deblond et al., 2001). This time frame is beyond the scope of this study, but, importantly, these data suggest a relatively quick, but consistent movement of India from the polar to mid- or low-latitude position between ~800 and 750 Ma. The Congo craton probably remained near equator during the same time interval, but rotated at about 90°.

There are no reliable late Neoproterozoic palaeomagnetic poles from the Congo or Kalahari Blocks. The post-orogenic Sinyai dolerite intrudes the East African orogen and provides a pole for this part of proto-Gondwana (Meert and Van der Voo, 1996). In India, the age of the palaeomagnetic poles from the Bhander and Rewa Series is only broadly determined to be latest Precambrian or Early Cambrian (McElhinny et al., 1978; Evans, 2000). Recently, Chirananda De (2003) reported the discovery of medusoid fossils of Ediacaran affinity at the base of Bhander Group, apparently below the sampling strata of McElhinny et al. (1978). If so, the time range for the Bhander pole might be narrowed. In this study we made a conservative assumption of 530–560 Ma.

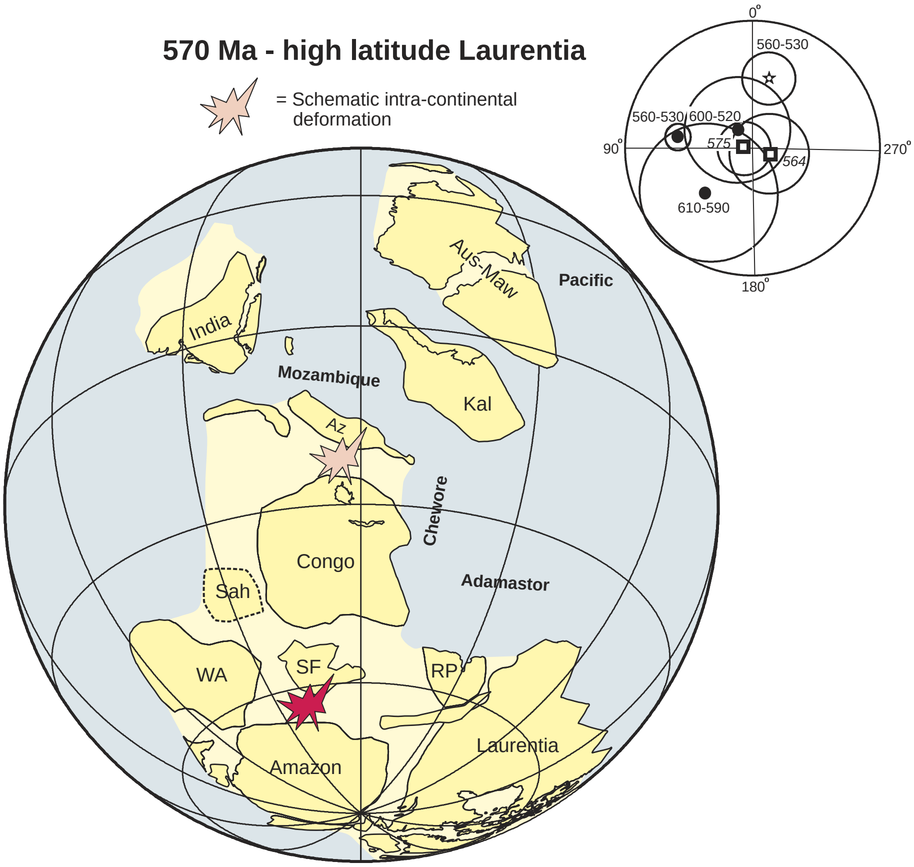

Fig. 4. Palaeogeographic reconstruction at 570 Ma with Laurentia in a high latitudinal position. Key and abbreviations the same as for Figs. 2 and 3.

In Australia, palaeomagnetic information from a stratigraphic drill hole in the Neoproterozoic Officer basin shows that Australia probably inhabited low latitudes since ~820 Ma (Pisarevsky et al., 2001b). From ~650 Ma to 550 Ma there is a swathe of palaeopoles, but age constraints on individual palaeopoles are generally poor. Nevertheless, the swathe forms a consistent pattern that places Australia in low latitudes throughout this time (Figs. 2–6), with the southern margin being near the equator by 600 Ma (Schmidt et al., 1991; Schmidt and Williams, 1995; Sohl et al., 1999).

The Laurentian palaeomagnetic data for ~750–600 Ma are sparse (Table 1). This ~150 my time interval is palaeomagnetically unconstrained. In our reconstructions we conservatively presumed a minimal drift of Laurentia between ~720 and 615 Ma (Fig. 3). In contrast, the latest Neoproterozoic–Early Cambrian palaeomagnetic results from Laurentia display a complicated picture. On one hand, two reliable poles from the Callander Complex (Symons and Chiasson, 1991) and the Catoctin Volcanics, component A (Meert et al., 1994), indicate a high-latitude position of Laurentia at 580–560 Ma. On the other hand, at least two other equally reliable palaeomagnetic results for the same time span from the Sept Iles intrusion (Tanczyk et al., 1987; Higgins and van Breemen, 1998) and the Catoctin Volcanics, component B (Meert et al., 1994) are in favour of the near-equatorial position. The result of Tanczyk et al. (1987) is confirmed by an independent study of Kirschvink et al. (2003). Other results for the time interval around 570–560 Ma (Table 1) are less reliable. Correspondingly, two alternative palaeogeographic models have been proposed (Pisarevsky et al., 2000, 2001a; Meert and Van der Voo, 2001; McCausland et al., 2003). We

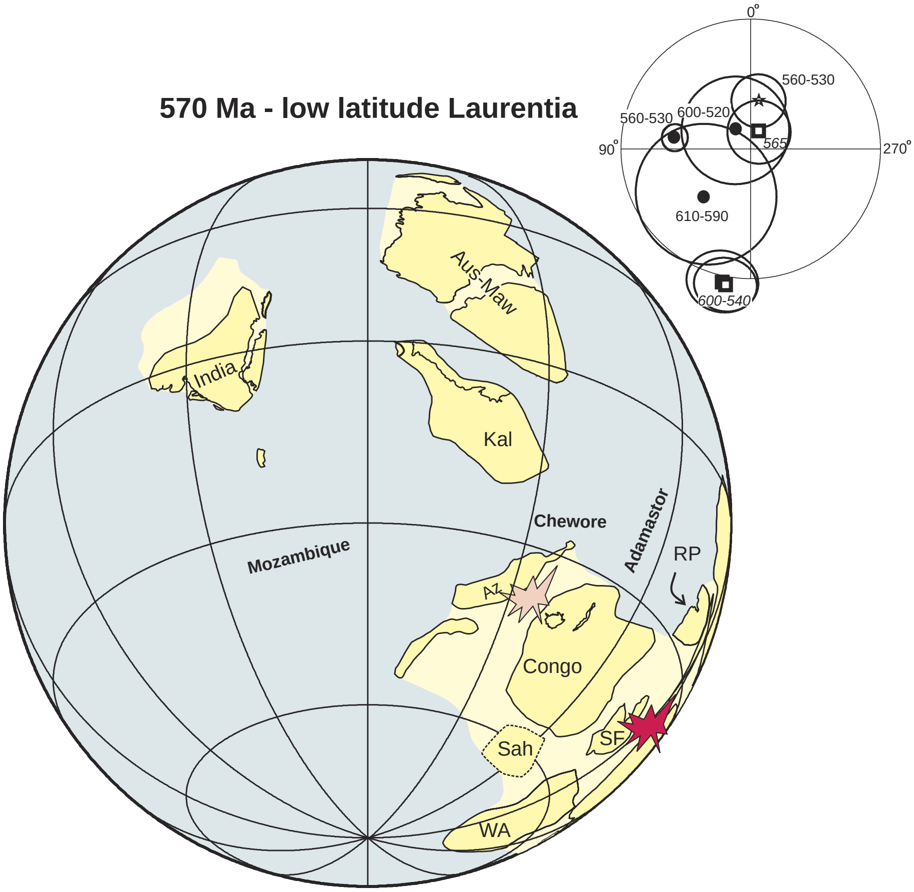

Fig. 5. Palaeogeographic reconstruction at 570 Ma with Laurentia in a low-latitudinal position—key and abbreviations the same as for Figs. 2–4.

follow these and here produce two alternative reconstructions for 570 Ma (Figs. 4 and 5).

4. Components of the Circum-Indian Orogens (the East African Orogen and the Pinjarra Orogen)

4.1. Indian Block

In this contribution cratonic India (including the Dharwar, Bastar or Bhandara, Singhbhum and Bundelkhand cratons of India, the Antongil Block of east Madagascar, and the Napier and Rayner Provinces of East Antarctica) (Figs. 1, 7–9) is considered to have amalgamated before the middle Neoproterozoic. This amalgamation largely took place along the Central India Tectonic Zone and the Eastern Ghats Orogen (Fig. 7). The Central Indian Tectonic Zone formed during the collision of a combined Dharwar/Bastar/Singhbhum Craton (much of present east India) and the Bundelkhand Craton (now in northwest India). The major collision here is thought to have occurred at ~1500 Ma (Yedekar et al., 1990; Roy and Prasad, 2003), although significant crustal shortening is also reported at ~1100 Ma (Roy and Prasad, 2003). Similarly, the southern Eastern Ghats Orogen (the Krishna Province of Dobmeier and Raith, 2003) records crustal thickening and associated deformation and metamorphism in the late Palaeoproterozoic/early Mesoproterozoic (1650–1550 Ma) (Mezger and Cosca, 1999; Rickers et al., 2001; Dobmeier and Raith, 2003) that was likely to be the result of a collision with the Napier Complex of East Antarctica, which also records an ~1.6 Ga tectonothermal event

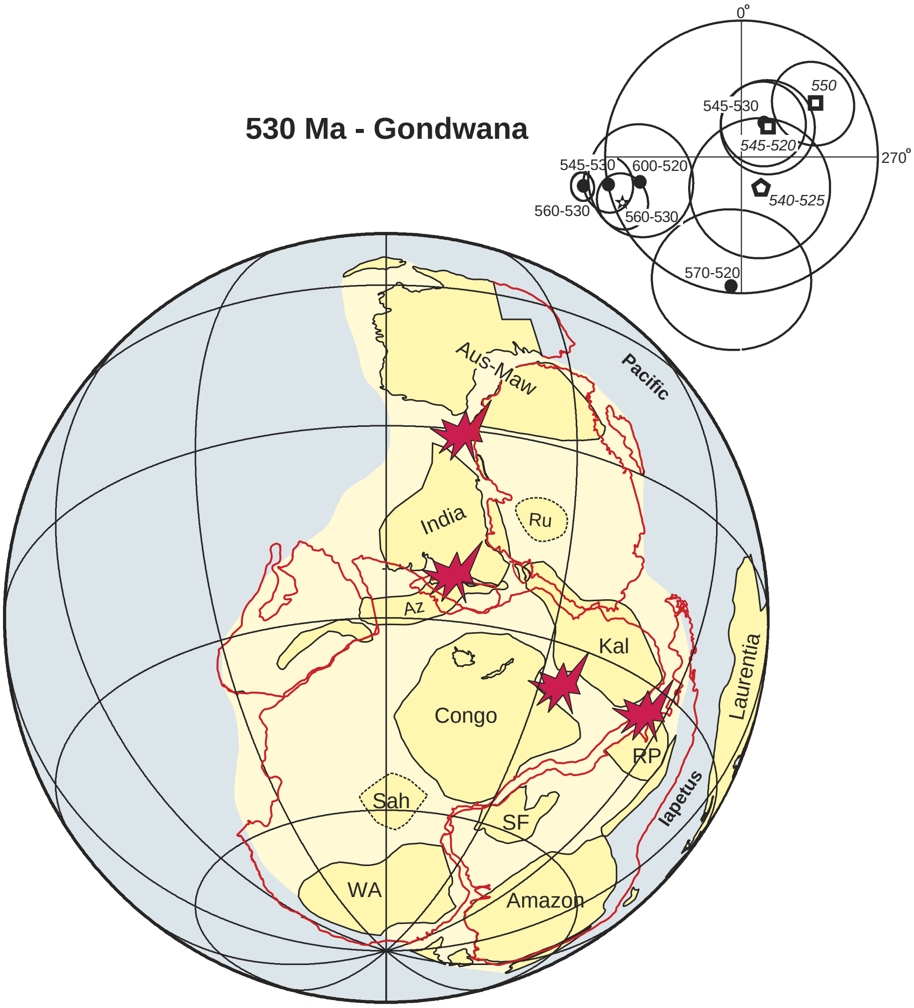

Fig. 6. Palaeogeographic reconstruction at 530 Ma with an assembled Gondwana (present continental outlines reconstructed for Gondwana using the reconstruction of Reeves and De Wit, 2000)—key and abbreviations the same as for Figs. 2–4, except: Ru = Ruker Terrane.

(Grew et al., 2001; Dobmeier and Raith, 2003; Owada et al., 2003). The Eastern Ghats Province of Dobmeier and Raith (2003) covers the more northerly outcrop of the Eastern Ghats Orogen (Fig. 7) and records pervasive high-grade metamorphism and deformation at 985–950 Ma (Shaw et al., 1997; Mezger and Cosca, 1999; Crowe et al., 2003; Dobmeier and Raith, 2003). The Rayner Complex of Eastern Antarctica (stretching south to the northern Prince Charles Mountains) (Fig. 8) records a very similar metamorphic history to that of the Eastern Ghats Province with coeval granulite-facies metamorphism, deformation and charnockite emplacement at 1020–900 Ma (Black et al., 1987; Young and Black, 1991; Beliatsky et al., 1994; Kinny et al., 1997; Shiraishi et al., 1997; Boger et al., 2000). The cause of this latest Mesoproterozoic/early Neoproterozoic orogenesis is unknown, but has been suggested to be due to collision between proto-India (including the Napier and Rayner Provinces) and the Archaean Ruker Terrane of the southern Prince Charles Mountains (Harley, 2003) (Fig. 8), which may in turn be part of a larger continental mass hidden beneath the Antarctic ice cap. This is disputed by Boger et al. (2001) who interpreted the boundary between the Ruker Terrane and the northern Prince Charles Mountains as a Neoproterozoic suture,

Table 1
Neoproterozoic palaeomagnetic poles from Laurentian and Gondwanan blocks

Object	Age (Ma)	Pole (°N)	Pole (°E)	A_{95} (°)	*Q*	Reference
India						
Bhander and Rewa Series	530–560	47	213	6	5	McElhinny et al., 1978
Mahe dykes, Seychelles[a]	748–755	80	79	11	4	Torsvik et al., 2001b
Malani Igneous Suite	751–771	68	88	8	6	Torsvik et al., 2001a
Australia						
Antrim Plateau Volcanics	520–570	9	160	17	4	McElhinny and Luck, 1970
Hawker Group	530–545	21	195	9	5	Klootwijk, 1980
Todd River Dolomite	530–545	43	160	6	7	Kirschvink, 1978
Upper Arumbera Sandstone	530–560	47	153	3	7	Kirschvink, 1978
Albany Belt Dykes	520–600	38	167	12	4	Harris and Li, 1995
Brachina Fm., SA	590–610	33	148	16	6	McWilliams and McElhinny, 1980
Elatina Fm., SA	600–620	51	157	2	5	Embleton and Williams, 1986
Elatina Fm., SA	600–620	54	147	1	6	Schmidt et al., 1991
Elatina Fm., SA	600–620	52	167	11	7	Schmidt and Williams, 1995
Elatina Fm., SA	600–620	39	186	9	7	Sohl et al., 1999
Yaltipena Fm., SA	620–630	44	173	8	7	Sohl et al., 1999
Angepena Fm., SA	630–730	33	164	12	3	McWilliams and McElhinny, 1980
Merinjina Tillite, SA	650–750	45	166	8	5	McWilliams and McElhinny, 1980
Lower Arumbera and Upper Pertatataka Fms., NT	650–750	44	162	10	6	Kirschvink, 1978
Mundine Well dykes	755 ± 3	45	135	4	6	Wingate and Giddings, 2000
Congo						
Mbozi Complex, Tanzania	755 ± 25	46	325	9	6	Meert et al., 1995; Evans, 2000.
Laurentia						
Mean Early Cambrian pole	520–545	4	155	10		McElhinny and McFadden, 2000
Skinner Cove Formation	550 ± 3	15	157	9	5	McCausland and Hodych, 1998
Sept Iles Intrusion	565 ± 4	20	141	7	5	Tanczyk et al., 1987; Higgins and van Breemen, 1998
Catoctin Volcanics A	564 ± 9	−43	128	9	5	Meert et al., 1994; Aleinikoff et al., 1995
Catoctin Volcanics B	564 ± 9	−4	193	10	4	Meert et al., 1994; Aleinikoff et al., 1995
Johnnie Formation	570 ± 10	−10	162	7	4	Van Alstine and Gillett, 1979; Hodych et al., 2004
Buckingham lavas	573 ± 32	−10	161	8	4	Dankers and Lapointe, 1981
Callander Complex	575 ± 5	−46	121	6	6	Symons and Chiasson, 1990
Long Range Dykes	614 + 6/− 4	−11	164	22	5	Murthy et al., 1992; Kamo and Gower, 1994
Franklin dykes	723 + 4/− 2	5	163	5	6	Heaman et al., 1992; Park, 1994
Natkusiak Formation	723 + 4/− 2	6	159	6	6	Palmer et al., 1983; Heaman et al., 1992.
Tsezotene sills and dykes	779 ± 2	2	138	5	5	Park et al., 1989; LeCheminant and Heaman, 1994
Wyoming dykes	782 ± 8 785 ± 8	13	131	4	5	Harlan et al., 1997
Gondwana						
Mean 540–560 Ma	540–560	10	150	7		McElhinny et al., 2003
Mean Early Cambrian pole	525–540	−23	154	15		McElhinny et al., 2003

750–530 Ma palaeomagnetic poles.
Quality factor (after Van der Voo, 1990).
[a] Rotated to India 28° counterclockwise around the pole of 25.8°N, 330°E (Torsvik et al., 2001b).

Table 2
Euler rotation parameters used in Figs 2–6

Craton/block/terrane	Age (Ma)	Pole (°N)	Pole (°E)	Angle (°)
Laurentia to absolute framework	750	19.63	−138.02	−95.39
	630	16.50	−128.69	−105.72
(high-latitude model)	570	−3.22	−139.32	−135.13
(low-latitude model)	570	7.08	−132.72	−74.53
	530	12.91	−127.76	−97.60
Greenland to Laurentia	750–530	70.40	−94.10	−18.00
Amazonia to Laurentia	750–630	11.97	−47.01	−110.65
(high-latitude model)	570	11.97	−47.01	−110.65
(low-latitude model)	570	10.78	−46.52	−109.64
	530	0.73	−42.46	−102.55
West Africa to Amazonia	750–530	53.00	−35.00	−51.00
Florida to Amazonia	750–530	53.34	44.15	30.54
Yucatan to Amazonia	750–530	2.18	−78.62	−90.37
Baltica to Laurentia	750–630	75.83	−94.18	−59.46
Rio de La Plata to Laurentia	750	10.39	−46.48	−94.42
	630	10.72	−46.82	−99.43
(high-latitude model)	570	10.15	−49.66	−107.87
(low-latitude model)	570	8.89	−49.18	−106.95
	530	0.73	−42.46	−102.55
Pampean to Laurentia	750	11.32	−44.86	−95.76
	630	11.05	−46.18	−99.97
(high-latitude model)	570	10.22	−49.50	−107.99
(low-latitude model)	570	8.97	−49.01	−107.08
	530	0.73	−42.46	−102.55
Rockall to Laurentia	750–530	75.33	159.57	−23.47
Oaxaquia to Amazonia	750–630	12.10	81.72	53.39
	570	11.57	71.46	54.88
	530	9.88	74.53	55.86
Chortis to Amazonia	750–570	5.73	−78.50	139.74
	530	3.64	−79.73	143.56
Australia to absolute framework	750	52.42	77.28	75.83
	630	50.30	98.64	83.02
	570	44.86	108.52	85.15
	530	35.23	124.11	85.47
Mawson to Australia	750–530	1.29	37.71	30.31
Kalahari to absolute framework	750	61.96	66.31	142.58
	630	53.29	82.06	139.00
(high-latitude model)	570	36.16	106.64	145.81
(low-latitude model)	570	37.53	109.38	145.26
	530	32.29	125.87	140.06
Dronning Maud Land to Kalahari	750–530	9.67	148.78	−56.28
India to absolute framework	750	73.35	42.17	63.76
	630	69.34	128.79	78.39
(high-latitude model)	570	62.73	142.12	91.01
(low-latitude model)	570	62.23	149.22	89.51
	530	50.99	165.23	126.19
Rayner to India	750–530	4.85	−163.36	−93.16
Congo to absolute framework	750	10.53	82.03	140.09
	630	18.61	113.89	146.41
(high-latitude model)	570	51.22	124.60	134.63
(low-latitude model)	570	19.37	133.95	144.06
	530	32.29	125.87	140.06

Table 2 (*continued*)

Craton/block/terrane	Age (Ma)	Pole		Angle (°)
		(°N)	(°E)	
São Francisco to Congo	750–530	53.00	−35.00	51.00
Azania to Congo	750	28.76	1.16	7.35
	630–570	3.02	125.5	0.07
	530	2.79	−140.58	−8.35
Afif–Abas to Congo	750	9.02	−175.29	−3.68
	630–530	55.21	−3.58	−5.21

thereby suggesting that Neoproterozoic India did not include the Ruker Terrane. If the latter interpretation is correct, Neoproterozoic orogenesis migrates south from the latest Mesoproterozoic/early Neoproterozoic Rayner Complex to Neoproterozoic/Early Palaeozoic metamorphism and deformation in the Prince Charles Mountains (Fitzsimons, 2000a,b; Boger et al., 2001). Late Neoproterozoic/Palaeozoic deformation in the Eastern Ghats and Rayner Complex themselves are concentrated along discrete shear zones (Clarke, 1988; Crowe et al., 2001) and did not pervasively affect the orogen (Dobmeier and Raith, 2003).

The northeastern extent of Neoproterozoic India is again poorly constrained. Late Neoproterozoic deformation reworks pre-existing structures in the northeastern Eastern Ghats Orogen (Crowe et al., 2001), but no new crust or pervasive deformation is reported. Rb–Sr whole rock and mineral isochrons from calc-alkaline granitoids and pegmatites in Assam and Meghalaya (Fig. 7) yield imprecise Neoproterozoic ages (van Breeman et al., 1989; Ghosh et al., 1991; Panneer Selvan et al., 1995). These rocks intrude Palaeo-Mesoproterozoic gneisses (Ghosh et al., 1994) and may have intruded in an active continental margin setting, near the boundary of Neoproterozoic India. Thrust slices of Neoproterozoic India lie within the Himalaya, where both Palaeoproterozoic and Neoproterozoic granitoids and metabasalts have been reported (Le Fort et al., 1986; Miller et al., 2000; Singh et al., 2002).

Early Neoproterozoic granitoids and rhyolites in the northwest of India (Aravalli–Delhi Belt) may represent the early stages of an early–mid Neoproterozoic arc (Deb et al., 2001; Pandit et al., 2003) that later migrated oceanward to include the mid-Neoproterozoic (~750 Ma) granitoids of the Seychelles (Tucker et al., 2001; Ashwal et al., 2002) and the Bemarivo Belt of northernmost Madagascar (Fig. 9) (Tucker et al., 1999a; Collins and Windley, 2002).

4.1.1. Vestfold Hills

The Vestfold Hills region of Princess Elizabeth Land in East Antarctica (Fig. 8) consists largely of 2520–2480 Ma orthogneisses (Black et al., 1991; Snape et al., 1997) that experienced granulite-facies metamorphism prior to 2470 Ma (Snape et al., 1997) and were subsequently intruded by a series of dykes and larger igneous bodies throughout the Palaeoproterozoic and Mesoproterozoic (Lanyon et al., 1993). Palaeoproterozoic monazite cooling ages (Kinny et al., unpublished data, referred to in Harley, 2003) demonstrate that this region was not pervasively affected by the high-temperature Neoproterozoic tectono-thermal events that characterise much of this part of East Antarctica. In this contribution we include the Vestfold Hills with India. However, we realise that it is just as likely that this region formed a separate Neoproterozoic continent of unknown extent or palaeogeographic location.

4.2. Congo/Tanzania/Bangweulu Block

The Congo/Tanzania/Bangweulu Block (Fig. 10) (Johnson et al., in press) includes the Congo–Angola–Kasai Craton, the Tanzania Craton and the Bangweulu Block. The Tanzania and Congo–Angola–Kasai Cratons are juxtaposed along the Mesoproterozoic Kibaran Belt (Rumvegeri, 1991; Kokonyangi et al., 2001), whereas the Bangweulu Block accreted to the Tanzania Craton along the ~2.1–1.9 Ga Ubendian/Usagaran Orogen (Lenoir et al., 1994b; Möller et al., 1995; Collins et al., 2004).

The Bangweulu Block is enigmatic with little occurrence of pre-2.1 Ga rocks, suggesting that much

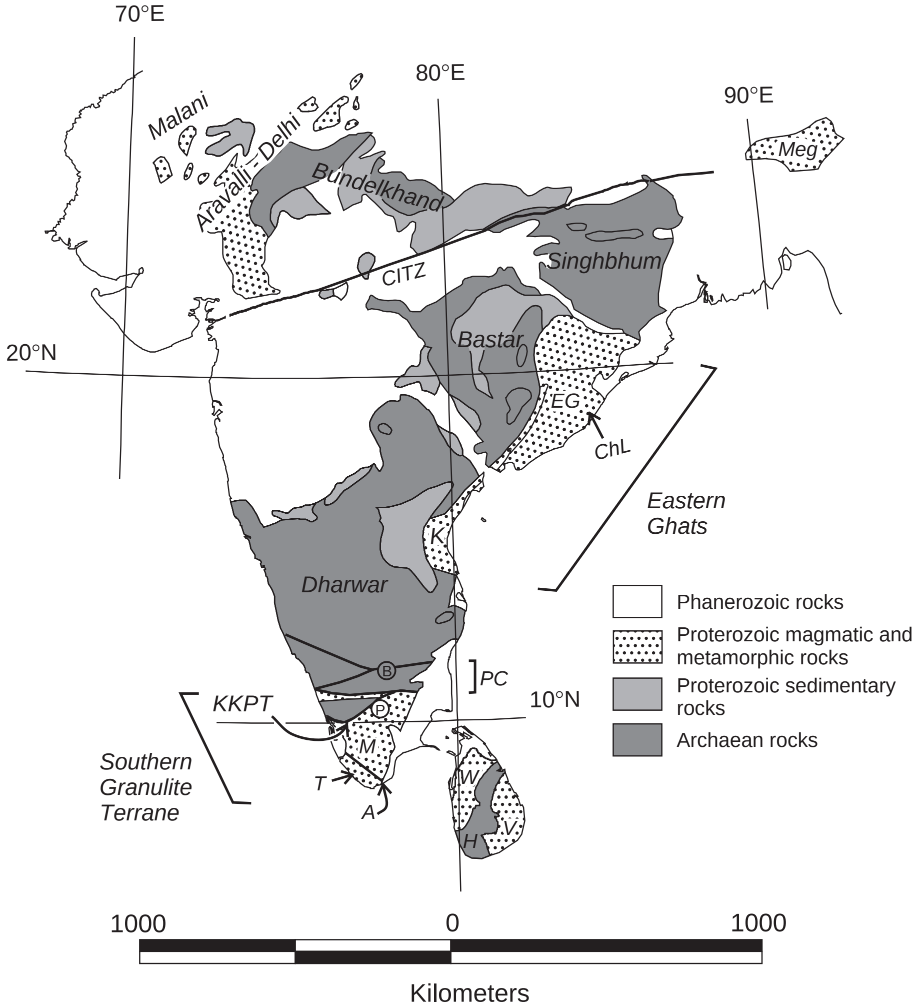

Fig. 7. Location and geological province map of India (geology based on, Commission for the Geological Map of the World, 2000). Bastar, Bundelkhand, Dharwar and Singhbhum = main Archaean cratons. Eastern Ghats = Eastern Ghats Orogen, EG = Eastern Ghats Province and K = Krishna Province of the Eastern Ghats Orogen, ChL = Chila Lake Domain of the Eastern Ghats Province (Dobmeier and Raith, 2003). A = Achancovil shear zone, B = Bhavani, CITZ = Central Indian Tectonic Zone, H = Highland Complex, KKPT = Karur–Kamban–Painavu–Trichur shear zone, M = Madurai Block, Meg = Assam and Meghalaya, P = Palani, PC = Palghat–Cauvery shear zone system, T = Trivandrum Block, V = Vijayan Complex, W = Wanni Complex.

of the ‘block’ may consist of a Palaeoproterozoic accretionary complex. Recent reports of a 2738 ± 4 Ma granite (De Waele et al., 2003a) and early Palaeoproterozoic T_{DM} ages from ~2.0–1.8 Ga granites (De Waele et al., 2003b) do, however, suggest older continental rocks are involved. Neoproterozoic deformation reactivated the boundary between the Tanzania Craton and Bangweulu Block, but is restricted to localised shear zone movement (Lenoir et al., 1994b; Theunissen et al., 1996). Therefore, in this contribution we consider the Congo/Tanzania/Bangweulu Block to be one continental block throughout the Neoproterozoic (Figs. 2–6).

4.3. Australia/Mawson Block

The Australia/Mawson Block includes cratonic Australia and corollaries in Antarctica (Figs. 1 and

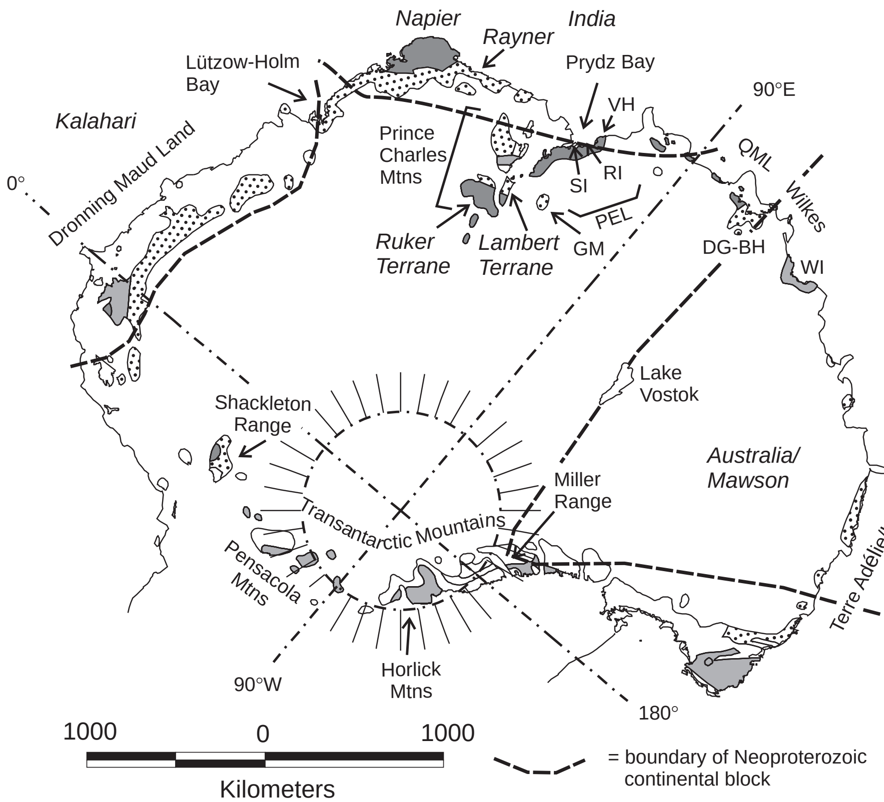

Fig. 8. Location and geological province map of East Antarctica (geology based on Commission for the Geological Map of the World, 2000). The location of interpreted Neoproterozoic continental boundaries is drawn. DG-BH = Denman and Scott Glaciers and Bunger Hills, GM = Grove Mountains, KGV = King George V Land, PEL = Princess Elizabeth Land, QML = Queen Mary Land, RI—Rauer Islands and Terrane, SI = Søstrene Island, WI = Windmill Islands, Wilkes = Wilkes Land, VH = Vestfold Hills. Key as Fig. 7.

8). It was assembled in the Palaeoproterozoic and Mesoproterozoic along a series of orogens that crop out in central and west Australia and along the Wilkes Land coast of East Antarctica (Fanning et al., 1995; Myers et al., 1996; Fitzsimons, 2003b; Giles et al., 2004). The extent of the Mawson Craton under the Antarctic ice cap is unknown. In this contribution we follow Fitzsimons (2003b) by considering that coastal exposures of Archaean and Palaeoproterozoic rocks in Terre Adélie and King George V Land (Oliver et al., 1983; Monnier et al., 1996; Oliver and Fanning, 1999; Peucat et al., 1999; Fitzsimons, 2003b), Archaean rocks metamorphosed in the ~1.7 Ga Nimrod Orogeny in the Miller Range, within the Transantarctic mountains (Fig. 8) (Borg et al., 1990; Goodge and Fanning, 1999; Goodge et al., 2001) and Archaean–Palaeoproterozoic rocks reworked in the Albany–Fraser Orogen in the Bunger Hills and Windmill Islands (Sheraton et al., 1992, 1993; Post et al., 1997) to be the only exposures of the Antarctic part of the Neoproterozoic Mawson Craton. In this scenario, the western margin of the Mawson Craton stretches directly south from the Denman–Glacier/Bunger Hills region through the sub-glacial Lake Vostok, intersecting the Transantarctic mountains in the Horlick–Pensacola Mountains region (Fitzsimons, 2003a). Aeromagnetic and gravity data from the Lake Vostok region supports this interpretation by suggesting that a crustal boundary exists in the region that is the result of thrust sheet emplacement on a pre-existing passive margin (Studinger et al., 2003).

An alternative outline of the Neoproterozoic Mawson Craton include a continent encompassing most of East Antarctica, limited in the west by a Neoproterozoic orogen cutting west through the Prince Charles Mountains and joining the East

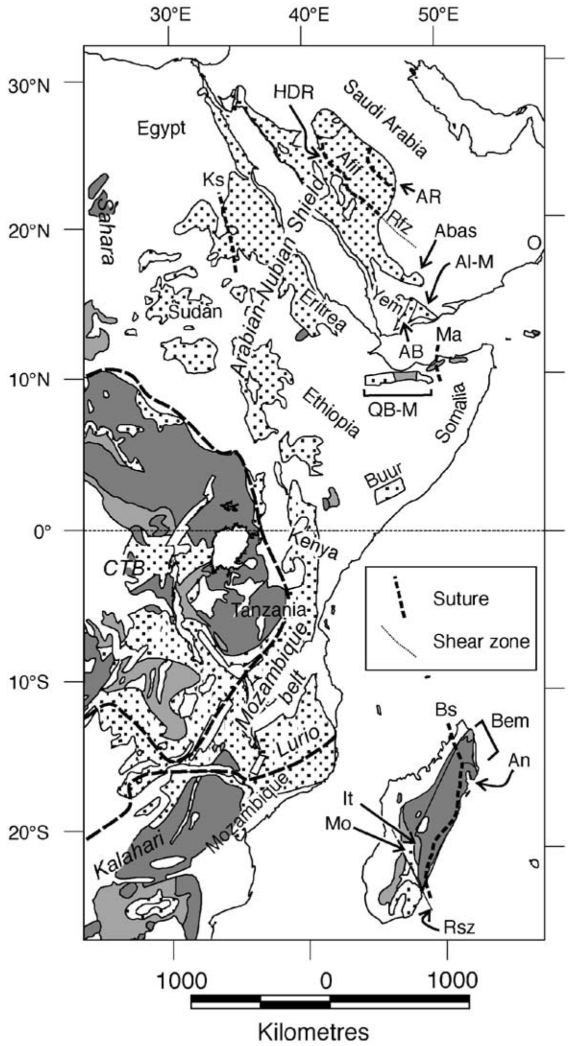

Fig. 9. Location and geological province map of East Africa, Arabia and Madagascar (geology based on Commission for the Geological Map of the World, 2000). AB = Al-Bayda Terrane, Abas = Abas Terrane, Al-M = Al-Mafid Block, An = Antongil Block, AR = Al-Rayn Terrane, Bem = Bemarivo Belt, Bs = Betsimisaraka Suture, HDR = Hulayfah–Ad Dafinah–Ruwah Suture, Buur = Buur Massif, CTB = Congo/Tanzania/Bangweulu Block, It = Itremo Group, KS = Keraf Suture, Lurio = Lurio Belt, Ma = Mait (or Maydh) Complex, Mo = Molo Group, QB-M = Qabri–Bahar and Mora Complexes, Rfz = Ruwah fault zone, Sahara = Saharan Metacraton, Rsz = Ranotsara shear zone, Yem = Yemen. Key as Fig. 7.

African Orogen in Dronning Maud Land (Boger et al., 2001; Boger and Miller, 2004), thereby including the Archaean rocks of the Ruker Terrane of the Southern Prince Charles Mountains (Fig. 1) (Kovach and Belyatsky, 1991; Boger et al., 2001) and probably the Palaeoproterozoic–Mesoproterozoic rocks in the Read Complex of the Shackleton Range (Talarico and Kroner, 1999) in the Mawson Craton (Goodge et al., 2001; Boger and Miller, 2004). In this contribution we consider these regions separate continental blocks (see below).

A considerably smaller Mawson Craton was postulated by Giles et al. (2004) in their Mesoproterozoic reconstruction of Australia and Laurentia. Their model can be related to the dimensions of the Neoproterozoic Mawson Craton used in this study if extensive continental material was accreted to the west side of an Archaean/Palaeoproterozoic nucleus in late Mesoproterozoic times. Support for this Mesoproterozoic increase in the size of the Mawson Craton lies in the Mesoproterozoic orogenesis in the Bunger Hills (Sheraton et al., 1992) and Windmill Islands (Post et al., 1997) and the change in basement Nd age provinces along the Shackleton Range (Borg and DePaolo, 1994).

4.4. Kalahari Block—the Lurio–Vijayan peninsula

In north Mozambique, the east-directed Lurio Belt (Fig. 9) thrusts ~615 Ma granulites over a 1000–1200 Ma foreland region (Sacchi et al., 1984; Costa et al., 1992; Kröner et al., 1997) that preserves no evidence of high-grade metamorphism in Neoproterozoic times (Kröner et al., 1997). The thrust base of the Lurio Belt contains siliceous metasedimentary rocks and basic and ultrabasic rocks that have been interpreted as fragments of an ophiolite (Sacchi et al., 1984).

A similar relationship with Neoproterozoic granulites thrust over a 1000–1200 Ma unit is found in southeastern Sri Lanka where the Highland and Wanni Complexes are thrust east over the Vijayan Complex (Fig. 7) (Kriegsman, 1995). Because of their proximity in the tight-fit reconstruction of Lawver et al. (1998), Collins and Windley (2002) linked the Vijayan Complex of Sri Lanka and the Lurio foreland to form a poorly known Mesoproterozoic tectonic block caught up in the East African Orogen, which they called they Lurio–Vijayan Block. Manhica et al. (2001) found no evidence for suture zone rocks separating rocks of central Mozambique and Natal from the Kalahari Block, which would suggest that, rather than being a separate Neoproterozoic continen-

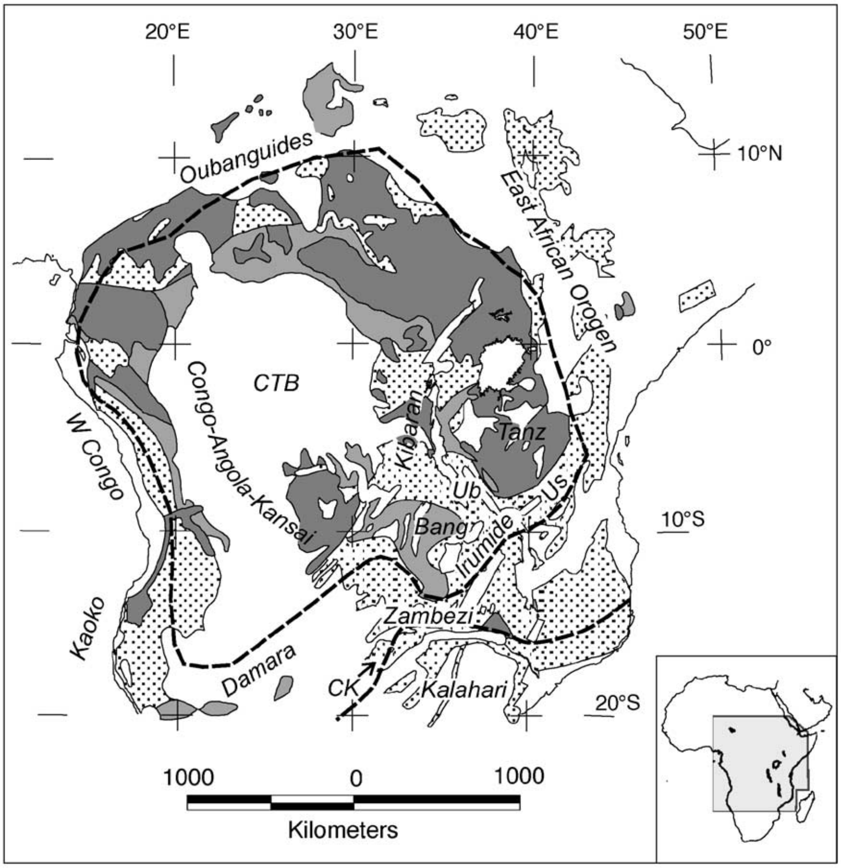

Fig. 10. Location and geological province map of central Africa (geology based on Commission for the Geological Map of the World, 2000). Bang = Bangweulu Block, CK = Choma–Kalomo Block, Congo–Angola–Kansai = Congo–Angola–Kasai Craton, CTB = Congo–Tanzania–Bangweulu Block, Damara = Damara Belt, Irumide = Irumide Belt, Kalahari = Kalahari Block, Kaoko = Kaoko Belt, Kibaran = Kibaran Belt, Oubanguides = Oubanguides Belt, Tanz = Tanzania Craton, Ub = Ubende belt, US = Usagaran Belt, W Congo = West Congo Belt, Zambezi = Zambezi Belt. Key as Fig. 7.

tal terrane, the Lurio–Vijayan Block formed a peninsula of the Kalahari Block (Fig. 1) thermally reworked to amphibolite-facies conditions in Neoproterozoic/Cambrian times.

The Lurio–Vijayan peninsula is progressively metamorphosed to higher grades to the south and stretches into central Dronning Maud Land (Fig. 8) where similar ~1170–1060 Ma rocks are pervasively reworked by 620–510 Ma magmatism, high-grade metamorphism and deformation (Moyes et al., 1993; Grantham et al., 1995; Jacobs et al., 1996, 1998, 2003a,b; Bauer et al., 2003; Paulsson and Austrheim, 2003; Ravikant et al., 2004). Jacobs et al. (1998) suggested that the similarity in ages between the Mozambiqe Belt and Dronning Maud Land was enough to extend the East African Orogen to form an extended East African–Antarctic Orogen.

4.5. Azania—the central East African Orogen Block

Collins and Windley (2002) summarised geochronological data from central Madagascar, southern and northern Somalia and east Ethiopia and identified an extensive ribbon of Archaean and Palaeoproterozoic crust that was thermally and structurally reworked in the Neoproterozoic/Cambrian East African Orogeny (Fig. 1). The Neoarchaean Al-Mafid Block of Yemen is likely to form the Arabian continuation of this extensive continental unit (Fig. 9) (Windley et al., 1996; Whitehouse et al., 1998; Collins and Windley, 2002), but a continuation into Saudi Arabia is masked by the scarcity of modern geochronological data and limited outcrop (Johnson and Woldehaimanot, 2003). This terrane may also pass south into the Madurai Block of southern India (Fig. 7) where ~2500 Ma igneous protoliths occur reworked in the Neoproterozoic/Cambrian (Collins and Santosh, 2004; Ghosh et al., 2004).

This continental block, here called Azania after the classical name for the East African coast, predominantly consists of 2.90–2.45 Ga rocks (Küster et al., 1990; Lenoir et al., 1994a; Paquette and Nédélec, 1998; Teklay et al., 1998; Kröner et al., 1999, 2000; Collins et al., 2001; Whitehouse et al., 2001). Palaeoproterozoic–Mesoproterozoic zircons, interpreted as xenocrysts, in gneisses from the Qabri Bahar Complex of northern Somalia (Kröner and Sassi, 1996) and eastern Ethiopia (Teklay et al., 1998; Yibas et al., 2002) suggest age heterogeneity in this pre-Neoproterozoic continental block. In Madagascar, a metasedimentary sequence (the Itremo Group) unconformably overlies orthogneisses (Cox et al., 1998) that elsewhere have been dated as late Archaean (Tucker et al., 1999b). The Itremo Group is inferred to have been deposited between ~1700 and 1500 Ma based on zircon and monazite age profiles, stromatolite characteristics and carbon isotope data (Cox et al., 1998, 2004; Fernandez et al., 2003). Elsewhere in Azania, the Mora Complex of northeast Somalia consists of marbles, quartzites and amphibolites and is interleaved with orthogneisses of the Qabri Bahar Complex (Kröner and Sassi, 1996). Xenocrysts from leucosomes developed in the Mora metasedimentary rocks yielded $^{207}Pb/^{206}Pb$ evaporation ages of 1405 ± 5 Ma and 1422 ± 22 Ma (Kröner and Sassi, 1996) suggesting that these rocks were deposited after this date and before 842 ± 4 Ma, the interpreted lower constraint on high-grade metamorphism (Kröner and Sassi, 1996). Mid-Neoproterozoic (740–820 Ma) intrusions occur throughout Azania that are commonly reworked by subsequent deformation and metamorphism (Kröner and Sassi, 1996; Handke et al., 1999; Tucker et al., 1999b; Kröner et al., 2000; Whitehouse et al., 2001; Yibas et al., 2002). In central Madagascar and southeast Ethiopia, these mid-Neoproterozoic rocks preserve geochemical evidence of having been intruded above a subduction zone (Handke et al., 1997; Brewer et al., 2001; Yibas et al., 2003).

Both the east and west margins of Azania are marked by Neoproterozoic sedimentary sequences and rocks consistent with having formed in an oceanic environment. To the east of central Malagasy Azania, ~800–550 Ma metasedimentary rocks occur with podiform peridotites and gabbros in the Betsimisaraka Suture (Fig. 9) (Collins et al., 2003c). Further north, in northern Somalia, the Maydh (or Mait) Complex (Fig. 9) marks the eastern margin of Azania and consists of Neoproterozoic pillow lavas, microgabbros and greenschist-facies metasedimentary rocks (Utke et al., 1990). The western margin of Azania is marked in Madagascar by highly strained metasedimentary rocks of the Molo Group that are considered to be deposited between ~620 and 560 Ma (Cox et al., 2004). In southern Ethiopia, quartzofeldspathic gneisses, marbles and sillimanite–kyanite schists may correlate with lower-grade clastic metasedimentary rocks in Kenya that were deposited between 1200 and 800 Ma (Mosley, 1993; Yibas et al., 2002). Neoproterozoic ophiolites and juvenile volcanic rocks are found in a broad, southern tapering band throughout eastern Tanzania, Kenya, Ethiopia, Eritrea, Sudan, east Egypt and the Arabian peninsula (Vearncombe, 1983; Shackleton, 1986; Beraki et al., 1989; Mosley, 1993; Sassi et al., 1993; Stern, 1994, 2002; Maboko, 1995; Teklay et al., 1998; Yibas et al., 2002, 2003; Johnson and Woldehaimanot, 2003; Kröner et al., 2003b).

The Itremo and the Molo Groups have zircon age spectra consistent with being derived from Azania and East Africa (Cox et al., 2004) and contrast with Neoproterozoic metasedimentary rocks in eastern Madagascar that are consistent with being derived from the Dharwar Craton of southern India, suggesting that the Betsimisaraka Suture marks the site of a significant strand of the Mozambique ocean (Collins et al., 2003c).

The detrital zircon record of the Itremo Group metasedimentary rocks suggest that Azania (or at least the central Malagasy part of it) was adjacent to the Congo/Tanzania/Bangweulu Block in the late Palaeoproterozoic/early Mesoproterozoic (Cox et al., 2004).

A poorly dated phase of deformation occurred in central Madagascar, between ~1700–1550 Ma and

~800 Ma, and created ~20 km amplitude recumbent folds in the Itremo Group (Collins et al., 2003b). This may be the result of a lateral continuation of the ~1020 Ma (De Waele et al., 2003c) Irumide Orogen, or active margin deformation in the ~800–750 Ma Andean-style continental arc that developed in the region.

Azania separated from the Congo/Tanzania/Bangweulu Block sometime after deposition of the Itremo Group, possibly due to roll-back of the Mozambique Ocean slab under central Madagascar from a subduction zone marked by the eastern Malagasy Betsimisaraka Suture (Collins et al., 2000, 2003c; Collins and Windley, 2002). In this model, the basin between Azania and the Congo/Tanzania/Bangweulu Block would have initiated as a back-arc basin that developed within and behind the ~800–750 Ma continental arc developed on Azania.

4.6. Afif–Abas Block

The Afif–Abas Block (Fig. 1) here consists of the Khida subterrane, part of the Afif Terrane of Saudi Arabia (Johnson and Kattan, 2001; Johnson and Woldehaimanot, 2003), and the Abas Terrane of Yemen (Fig. 9) (Windley et al., 1996; Whitehouse et al., 1998, 2001). Correlation between the basement terranes of Yemen and Saudi Arabia is controversial due to regions of non-exposure, the presence of terrane-cutting strike-slip faults and the paucity of geochronological data (Windley et al., 1996; Whitehouse et al., 1998, 2001; Johnson and Kattan, 2001; Johnson and Woldehaimanot, 2003). In this contribution we link the Afif and Abas Terranes because they both contain rocks with Palaeoproterozoic Nd model ages (Windley et al., 1996; Whitehouse et al., 2001), they have similar Pb-isotope values (Whitehouse et al., 2001), their along-strike offset is consistent with offset along the sinistral Ruwah fault zone (Johnson and Kattan, 2001), and both terranes preserve evidence for ~800–750 Ma metamorphism and magmatism (Whitehouse et al., 1998; Johnson and Woldehaimanot, 2003). The Afif–Abas Block is separated from the Al-Mafid Block (northern component of Azania) in the Yemen by the Al-Bayda terrane that is interpreted as a Neoproterozoic arc (Windley et al., 1996). Nd isotope data, however, do suggest that Neoproterozoic intrusions in the Al-Bayda Terrane have sampled older material at depth (Whitehouse et al., 2001), presenting the possibility that Azania and the Afif–Abas Block may have been linked in the Neoproterozoic.

Neoproterozoic crust also exists beneath Phanerozoic rocks in central and eastern Arabia (Johnson and Woldehaimanot, 2003). Granitoids cropping out in Oman have 1000–750 Ma crystallisation ages (Gass et al., 1990), but it is unsure whether these are juvenile Neoproterozoic rocks, or whether they intrude older Proterozoic basement.

4.7. Saharan Block(s)

Pre-Neoproterozoic rocks have been reported throughout the central Saharan region from the river Nile in the east to the Tuareg Shield in the west (summarised in Abdelsalam et al., 2002). Collectively this region has been called the Saharan Metacraton (Fig. 1) (Abdelsalam et al., 2002), although the lack of exposure and geological data make it hard to determine whether the region was a single Neoproterozoic continental block, or several terranes. At its western margin, in the Tuareg Shield, a number of pre-Neoproterozoic terranes have been identified, separated by juvenile Neoproterozoic crust (Black et al., 1994; Caby, 2003; Liégeois et al., 2003).

4.8. Other pre-Neoproterozoic continental crust in the Circum-Indian Orogens

4.8.1. Northampton, Leeuwin and Mullingara Blocks, Western Australia

The three inliers of the Pinjarra Orogen along the Western Australian coast (Fig. 1) all preserve rocks with Mesoproterozoic or older ages.

Both the Northampton and Mullingarra Blocks preserve Mesoproterozoic metasedimentary rocks that were metamorphosed to amphibolite and granulite facies during latest Mesoproterozoic times (Bruguier et al., 1999; Cobb et al., 2001). The Mullingarra Block preserves an unmetamorphosed Palaeoproterozoic (2181 ± 10 Ma) monzogranite (Cobb et al., 2001) whilst 1068 ± 13 Ma post-tectonic granites and ~990 Ma pegmatites cut the Northampton Block (Bruguier et al., 1999). Recent interpretations have suggested that both the Mullingara and Northampton Blocks were formed elsewhere and transported to their present locations during the Neoproterozoic (discussed more fully in Fitzsimons, 2003b).

Unlike the more northern blocks, the Leeuwin Block was thoroughly metamorphosed and deformed during latest Neoproterozoic/Cambrian times (Myers, 1990; Wilde and Murphy, 1990; Wilde, 1999; Collins, 2003). Protolith ages range between 1200 and 500 Ma with U–Pb age maxima at ~1100 and 691 Ma (Collins, 2003). The Mesoproterozoic ages are restricted to the central Leeuwin Block (Nelson, 1998; Janssen et al., 2003), which may represent a separate terrane within the gneiss complex (Janssen et al., 2003).

4.8.2. East Prydz Bay, East Antarctica

Pre-Neoproterozoic protoliths have been found in the Rauer Island Group directly south of the Vestfold Hills, and from Søstrene Island in southern Prydz Bay (Fig. 8) (Hensen and Zhou, 1995). No evidence for pre-Neoproterozoic protoliths have yet been found in the ~150 km of scattered outcrops along the coast of eastern Prydz Bay between these two localities (see recent reviews of Hensen and Zhou, 1997; Fitzsimons, 2000b; Harley, 2003; Zhao et al., 2003).

Two distinct pre-Neoproterozoic rock packages have been identified in the Rauer Islands of eastern Prydz Bay (Kinny et al., 1993) (Fig. 8). The oldest rock package consists of 3300–2800 Ma tonalitic orthogneisses, layered igneous complexes and paragneisses that include marbles, pelites and quartzites (Harley, 2003). These are interleaved with 1060–1000 Ma felsic and mafic orthogneisses and Mesoproterozoic supracrustal rocks. These rocks have very different ages to the Vestfold Hills Block (see above), which has led to the identification of a discrete Rauer Terrane (Kinny et al., 1993; Harley and Fitzsimons, 1995; Hensen and Zhou, 1997). Controversy exists over whether the components of the Rauer Terrane amalgamated with each other and the Vestfold Hills Block at ~1000 Ma, or whether their juxtaposition occurred during the latest Neoproterozoic/Cambrian (Kinny et al., 1993; Harley and Fitzsimons, 1995; Hensen and Zhou, 1997; Harley et al., 1998). In the former case an amalgamated Vestfold Hills/Rauer Terrane would have existed before the Neoproterozoic events discussed in this paper. However, the observation that only some units of the Rauer Terrane preserve evidence for ~1000 Ma metamorphism, whilst the whole terrane is affected by 550–500 Ma isotopic disturbance suggests that a latest Neoproterozoic/Cambrian amalgamation is more likely (Harley and Fitzsimons, 1995) and that two discrete terranes existed in the Neoproterozoic.

Further south in Prydz Bay, evidence of pre-600 Ma deformation and metamorphism are restricted to Søstrene Island where Sm–Nd garnet-whole rock isochrons from mafic granulites that suggest high-grade metamorphism at ~990 Ma (Hensen and Zhou, 1995) and 960–920 Ma zircons from orthogneisses have been interpreted as protolith ages (Zhao et al., 1995). These rocks have been metamorphosed in the late Neoproterozoic/Cambrian (Hensen and Zhou, 1995) and are likely to represent a relatively small exotic block within the Neoproterozoic Prydz Belt.

4.8.3. Queen Mary's Land, East Antarctica

Enigmatic Archaean tonalitic orthogneisses from ~50 km west of the Denman Glacier (Fig. 8) have zircon populations at 3003 ± 8 Ma and 2889 ± 9 Ma that have been interpreted as emplacement and metamorphic ages respectively (Black et al., 1992). The tectonic position of these rocks is unknown, but lower intercept ages of 630–510 Ma and the lack of evidence for Mesoproterozoic tectonism led Fitzsimons (2000b) to suggest that they were juxtaposed with rocks east of the Denman and Scott Glaciers during Neoproterozoic deformation and lie close to the east margin of the Pinjarra Orogen.

4.8.4. Southern India

India south of the Palghat–Cauvery Shear Zone system (here including the numerous shear zone strands identified between Bhavani and Palani, Fig. 7, Chetty et al., 2003) is commonly divided into two crustal domains, the Madurai and Trivandrum Blocks, with the latter also known as the Kerala Khondalite Belt (Harris et al., 1994; Braun and Kriegsman, 2003). These two domains are separated by the Achankovil Shear Zone, an enigmatic structure with a controversial kinematic evolution and tectonic significance (Radhakrishna et al., 1990; Sacks et al., 1997, 1998; Rajesh et al., 1998).

Recently, Ghosh et al. (2004) proposed an isotopic boundary, separating Archaean rocks from Proterozoic rocks, that runs northeast–southwest through the towns of Karur, Kabam, Painavu and Trichur (and therefore called the KKPT shear zone) in the central Madurai Block. These authors suggest that this boundary is the southern margin of the Dharwar Craton (see also Bhaskar Rao et al., 2003). However, considerable isotopic, lithological, seismic and structural evidence suggests that the Palghat–Cauvery Shear Zone system juxtaposes terranes of different ages and tectonothermal histories (Chetty, 1996; Harris et al., 1996; Bartlett et al., 1998; Raith et al., 1999; Meißner et al., 2002; Chetty et al., 2003; Reddy et al., 2003). U–Pb and $^{207}Pb/^{206}Pb$ zircon ages of 2100–2500 Ma (Jayananda et al., 1995; Bartlett et al., 1998; Ghosh, 1999) from the region between the Palghat–Cauvery Shear Zone system and the KKPT are very similar to the Nd model ages from this area (Harris et al., 1994; Jayananda et al., 1995; Bartlett et al., 1998; Bhaskar Rao et al., 2003) suggesting that this region represents a juvenile ~2.5 Ga crustal unit (Bartlett et al., 1998), isolated from both the Dharwar Craton and the Proterozoic terranes south of the KKPT. This history is similar to that of Malagasy Azania and we propose that this Archaean part of the Madurai Block is a southern extension of Azania, translated east by the Malagasy Ranotsara shear zone (Fig. 1).

Metasedimentary rocks dominate south of the KKPT. However, Palaeoproterozoic ^{207}Pb–^{206}Pb zircon evaporation, electron-probe U–Pb monazite and zircon ages and Nd model ages have been used to suggest a Palaeoproterozoic phase of crustal growth and high-grade metamorphism (Harris et al., 1994; Bartlett et al., 1998; Braun et al., 1998; Braun and Kriegsman, 2003; Santosh et al., 2003). Most of the electron-probe analyses have been from monazite cores of metasedimentary rocks that are probably detrital grains. The Nd model ages provide an average crustal residence age of the source regions of the sedimentary protolith and the uncertainty in concordance inherent in the zircon evaporation technique means that the resultant age only reflects a minimum age. To date the evidence of Palaeoproterozoic high-grade metamorphism south of the KKPT is equivocal. Recent U–Pb Secondary Ion Mass Spectroscopy (SIMS) analyses on metasedimentary rocks in the Trivandrum and Madurai Blocks show that many of the detrital zircons originate from Late Archaean, Palaeoproterozoic and Mesoproterozoic sources with peaks at ~2700, 2260, 2105, 2000, 1855 and 1435 Ma (Collins and Santosh, 2004). Concordant Neoproterozoic grains show that at least some of these rocks were deposited in Neoproterozoic times (Collins and Santosh, 2004). Further U–Pb ion-probe work is needed to clarify the thermo-tectonic record of this region.

4.8.5. Ruker Terrane, Southern Prince Charles Mountains, East Antarctica

The Southern Prince Charles Mountains (Fig. 8) contain orthogneisses, whose protoliths crystallised at ~3150 Ma and contain xenocrysts that reach back to 3386 ± 6 Ma (Boger et al., 2001; Mikhalsky et al., 2001). These rocks were deformed before being intruded by an undeformed ~2645 Ma pegmatite (Boger et al., 2001) and are overlain by metasedimentary rocks of unknown age (Mikhalsky et al., 2001; Fitzsimons, 2003b). The southern extent of the terrane is also unknown and covered by ice. The northern boundary is marked by the Neoproterozoic Lambert Terrane and interpreted as a Neoproterozoic–Cambrian suture (Boger et al., 2001).

5. The timing of Neoproterozoic metamorphism and deformation in the Circum-Indian Orogens

In the following sections we discuss the timing of terrane amalgamation and Neoproterozoic metamorphism in the orogens that surround India (directions refer to relative positions in Gondwana) associated with the amalgamation of eastern Gondwana. Throughout, we have separated orogen (geographic extent) from orogeny (a specific mountain building event) as has been adopted in other deformed and metamorphosed regions (e.g. the Capricorn Orogen of Australia, Cawood and Tyler, 2004).

5.1. Neoproterozoic orogens west of India (the East African Orogen)

5.1.1. Arabian/Nubian Shield

The Arabian/Nubian Shield (Fig. 9) largely consists of intra-oceanic arc terranes that began to collide with each other at ~780–760 Ma (Johnson and Woldehaimanot, 2003). By ~650 Ma a number of these arc terranes had collided with the pre-Neoproterozoic Afif Terrane of Saudi Arabia along the

Hulayfah-Ad Dafinah-Ruwah Suture (Johnson and Woldehaimanot, 2003). The Afif Terrane probably also sutured with the Al-Rayn Terrane further east by 650 Ma (Stacey et al., 1984; Johnson and Woldehaimanot, 2003). ^{40}Ar–^{39}Ar hornblende ages of 614 ± 11 Ma from the Al-Bayda Terrane in Yemen also suggest terrane amalgamation between the pre-Neoproterozoic Afif and Al-Mahfid Terranes was complete before this time (Whitehouse et al., 1998). Final amalgamation between the various components of the Arabian/ Nubian Shield appears to have occurred along the Keraf Suture of the Sudan (Fig. 9) (Abdelsalam et al., 1998), which was active until ~600 Ma (Abdelsalam et al., 1998, 2003; Johnson and Woldehaimanot, 2003) (Fig. 11).

5.1.2. East Africa

The timing of peak metamorphism in Ethiopia and Somalia is incompletely known, largely due to the inaccessibility of much of the region. However, recent studies in southern Ethiopia are beginning to fill this gap. Yibas et al. (2002) reported a number of tectonothermal events in southern Ethiopia throughout the Neoproterozoic and related them to subduction/ accretion processes. The youngest is associated with granulite-facies metamorphism and post-dates a 540–520 Ma granite (Genzebu et al., 1994) and has been interpreted as being related to the amalgamation of Gondwana (Yibas et al., 2002). Yihunie (2002) suggested that peak metamorphism was associated with crustal shortening in the Kenticha sequence of southern Ethiopia and pre-dated a 554 ± 23 Ma granite. East of this region, In the poorly exposed and inaccessible Buur region of southern Somalia (Fig. 9), upper amphibolite–granulite facies metamorphism is dated between ~600 and 530 Ma (Küster et al., 1990; Lenoir et al., 1994a). Further north, in northern Somalia, post-kinematic granitoids are dated at ~630 Ma (Küster et al., 1990; Lenoir et al., 1994a). Metamorphic isotopic-resetting is reported throughout northern Somalia between 600 and 520 Ma (Sassi et al., 1993).

High-grade metamorphism in Tanzania is well dated between ~655 and 610 Ma (Coolen et al., 1982; Muhongo and Lenoir, 1994; Maboko and Nakamura, 1995; Möller et al., 1997, 2000; Muhongo et al., 2001; Kröner et al., 2003b; Sommer et al., 2003) and may correlate with similar age metamorphism in northern Mozambique (Kröner et al., 1997) and southern Kenya (Meert and Van der Voo, 1996). Both anti-clockwise (Appel et al., 1998) and clockwise (Sommer et al., 2003) P/T paths have been described from these rocks. Möller et al. (2000) showed that the Tanzanian granulites cooled slowly (2–5 °C/Ma) over <100 my, which supported the interpretation of Appel et al. (1998) that the metamorphism occurred at an active continental margin. Whatever the origin, the timing of high-grade metamorphism in eastern Tanzania is considerably older than that from more eastern components of the East African Orogen (see below).

5.1.3. Madagascar

Large 560 ± 7 Ma zircon rims developed in quartzites from west-central Madagascar are interpreted to reflect the timing of high-grade metamorphism and related high-strain deformation in this region (Cox et al., 2001, 2004). These ages are ~80 Ma younger than the age of syn-metamorphic zircon in eastern Tanzania and indicate an eastward younging in high-grade metamorphism through the East African Orogen. A zircon rim with a U/Pb age of 518 ± 9 Ma (Collins et al., 2003c) and identical SIMS U/Pb monazite ages (517 ± 1 Ma, Fitzsimons et al., 2004b) in ~800–550 Ma kyanite schists from the Betsimisaraka Suture suggest high-grade metamorphism in eastern Madagascar was ~40 my younger than that in western Madagascar.

Peak metamorphism in southern Madagascar is associated with large sheath folds and ductile thrusts (De Wit et al., 2001) and reached high-pressure/high-temperature conditions of 7–12 kbar and 750–940 °C (Ackermand et al., 1989; Nicollet, 1990; Markl et al., 2000) followed by a phase of medium pressure/ medium temperature metamorphism at 3–5 kbar and 650–730 °C (Markl et al., 2000). Zircons and monazites from southern Madagascar indicate that metamorphism occurred between 647 and 520 Ma (Paquette et al., 1994; Kröner et al., 1996, 1999; Ashwal et al., 1999; De Wit et al., 2001). De Wit et al. (2001) proposed that the high-pressure metamorphism occurred at 647–627 Ma and was followed by an extended period of medium-pressure granulite-facies conditions that lasted until ~520 Ma (see also Ashwal et al., 1999). Without U–Pb microprobe analyses of key grains, this suggested elevated thermal environ-

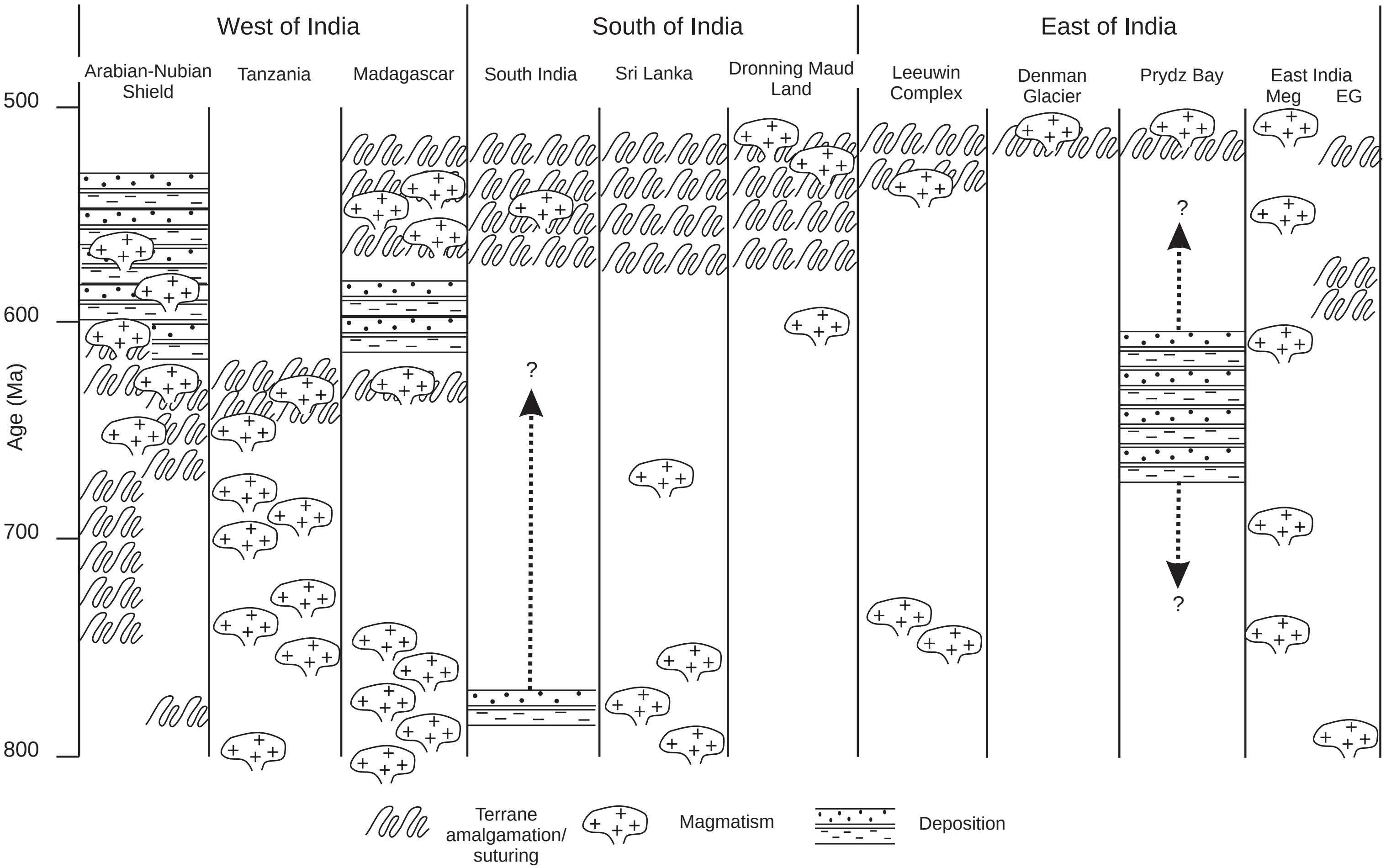

Fig. 11. Time–space plot for the middle to late Neoproterozoic tectonothermal events in orogens west, south and east of Gondwanan India.

ment remains a hypothesis that is indistinguishable from a more punctuated thermal history.

5.2. Neoproterozoic orogens east of India (Fig. 11)

5.2.1. Western Australia (Pinjarra Orogen)

Late Neoproterozoic granulite-facies metamorphism in Leeuwin Complex of south Western Australia (Fig. 1) was originally dated at ~615 Ma (Nelson, 1995, 1996). Collins (2003) pointed out a number of problems with this age interpretation and instead presented chemical and isotopic evidence to suggest that the zircon rim age of 522 ± 5 Ma was a better estimate of this high-grade metamorphism.

5.2.2. East Antarctica (Denman Glacier, Prydz Bay regions)

East Antarctic rocks deformed and metamorphosed in the Neoproterozoic Pinjarra Orogen [as defined by Fitzsimons (2003b) and synonymous with the Prydz–Denman–Darling Orogen of Fitzsimons (2000a) and the Prydz–Leeuwin belt of Veevers (2000)] occur west of the Denman Glacier in Queen Mary Land, south and east of the Vestfold Hills in Prydz Bay, in the Lambert Glacier region of the Prince Charles Mountains and in the Grove Mountains of Princess Elizabeth Land (Fig. 8).

Rocks that crop out in the southern Prydz Bay region preserve evidence for compressional deformation, peak metamorphism and partial melting at conditions of 800–860 °C and 5.5–7 kbars (Stuwe and Powell, 1989; Fitzsimons and Harley, 1992; Fitzsimons, 1996; Carson et al., 1997) at ~535–525 Ma (Zhao et al., 1992, 2003; Hensen and Zhou, 1995; Carson et al., 1996; Fitzsimons et al., 1997). The Grove Mountains, over 200 km east, preserve rocks that experienced a similar history with near peak metamorphic conditions of 850 °C and ~6 kbar (Liu et al., 2003) associated with zircon rims that yielded U–Pb ages of 529 ± 14 Ma (Zhao et al., 2003). Further north, in the Rauer Terrane, an Archaean layered metaigneous complex preserves evidence for U–Pb zircon resetting and new zircon growth at 519 ± 8 Ma (Harley et al., 1998). This widespread orogenic event that involved significant crustal thickening has been interpreted as a result of collision at a convergent plate boundary (Fitzsimons, 1996; Carson et al., 1997; Zhao et al., 2003). Zhao (2003) has suggested, based on immobile trace element compositions (Yu et al., 2002), that mafic and ultra-mafic lenses in the Grove Mountains (and by implication, southern Prydz Bay) represent the metamorphosed accreted remains of mid-ocean-ridge basalt and ocean–island basalt caught up in an oceanic accretionary system.

5.2.3. Eastern India (Assam, Meghalaya and the northern Eastern Ghats Orogen)

Neoproterozoic deformation, metamorphism and magmatism has been identified in two main regions of east India: 1) northeast India (Assam and the Meghalaya Plateau); and, 2) the northern Eastern Ghats, where it is related to shear zone development (Fig. 7).

Neoproterozoic granitoids intrude poorly dated schists in the basement exposed in far north-east India. These rocks are only dated by Rb–Sr methods, which yielded ages between 900 and 450 Ma (van Breeman et al., 1989; Ghosh et al., 1991, 1994). The age of the deformation and metamorphism in the country rocks is not known.

Mid- to Late-Neoproterozoic metamorphism, magmatism and deformation in the Eastern Ghats Orogen occurs in the Rengali and northern Eastern Ghats Provinces of Dobmeier and Raith (2003). Anorthosites and granitoids were emplaced in the Chilka Lake Domain between 795 and 740 Ma (Krause et al., 2001; Dobmeier and Simmat, 2002). This eastern domain was then metamorphosed to granulite-facies conditions between 690 and 660 Ma (Dobmeier and Simmat, 2002). Neoproterozoic deformation and metamorphism along the northern margin of the Eastern Ghats (the Rengali and northern Eastern Ghats Provinces) is associated with shear zone reactivation and occurred at ~550–500 Ma (Mezger and Cosca, 1999; Crowe et al., 2001; Dobmeier and Raith, 2003).

5.3. Neoproterozoic orogens south of India (Fig. 11)

5.3.1. Southern India

Both the Madurai and Trivandrum Blocks (Fig. 7) have been metamorphosed to granulite facies with ultra-high temperature assemblages of 900–1000 °C preserved in the Madurai Block (Brown and Raith, 1996; Raith et al., 1997; Satish-Kumar, 2000; Tsunogae and Santosh, 2003). Sm–Nd mineral isochrons (Santosh et al., 1992), garnet-whole rock ages (Choudhary et al., 1992), U–Pb electron-probe monazite ages (Braun et al., 1998; Santosh et al., 2003), and $^{207}Pb/^{206}Pb$ single zircon evaporation ages (Bartlett et al., 1998) from charnockites, garnet–biotite gneisses and garnet–sillimanite gneisses in the Trivandrum Block indicate high-grade metamorphism at 580–510 Ma. High-grade metamorphism in the Madurai Block also spans the Neoproterozoic/Cambrian boundary with garnet-whole rock ages indicating near-peak metamorphism at 553 ± 15 Ma from the central Madurai Block (Jayananda et al., 1995) and between ~610 and 560 Ma from the Palghat–Cauvery shear zone system (Meißner et al., 2002). Madurai Block $^{207}Pb/^{206}Pb$ single zircon evaporation ages of 547 ± 17 Ma, were interpreted as dating zircon rims (Bartlett et al., 1998), U–Pb electron probe monazite ages span 480–580 Ma (Santosh et al., 2003) and U–Pb SIMS and Thermal Ionisation Mass Spectroscopy (TIMS) zircon and monazite ages from syn-tectonic granites and metamorphic rims range from ~600–540 Ma (Ghosh et al., 2004). SIMS U–Pb data from 15 zircon rims from a quartzite at Ganguvarpatti in the central Madurai Block yielded an age of 506.5 ± 6.5 Ma that was interpreted as dating high-grade metamorphism in this rock (Collins and Santosh, 2004). These Middle Cambrian ion-microprobe ages are some of the youngest estimates of high-grade metamorphism in the Circum-Indian Orogens and suggest that metamorphism here outlasted that of other parts of the orogenic system.

5.3.2. Sri Lanka

Granulite-facies metamorphism in both the Highland and Wanni Complexes (Fig. 7) is dated at ~610 to 550 Ma (Baur et al., 1991; Hölzl et al., 1994). The Vijayan Complex was metamorphosed to amphibolite-facies between ~591 and 456 Ma (Hölzl et al., 1994; Kröner et al., 2003a).

Kröner et al. (2003a) proposed that Highland Complex was a microcontinent that collided with both the Wanni Complex and the Vijayan Complex during the accretion of Gondwana. Following Collins and Windley (2002), we suggest that the Vijayan Complex correlates with the Lurio foreland in Mozambique, and that together these formed the easternmost part of a Neoproterozoic peninsula of the Kalahari Block (the Lurio–Vijayan Peninsula—see above). Therefore, in this model the Sri Lankan basement consists of a series of exotic microcontinental blocks amalgamated in the late Neoproterozoic on the northeast margin of Neoproterozoic Kalahari.

5.3.3. East Antarctica (Dronning Maud Land)

A southern continuation of the East African Orogen into East Antarctica was proposed by Jacobs et al. (1998) as the East African–Antarctic Orogen. This was based on the recognition of high-grade Late Neoproterozoic–Cambrian metamorphism and deformation pervasively overprinting Late Mesoproterozoic/Early Neoproterozoic rocks in Dronning Maud Land (Moyes and Groenewald, 1996; Jacobs et al., 1998; Bauer et al., 2003). Granulite-facies conditions of ~6.8 kbar and 830 °C (Piazolo and Markl, 1999) occurred at ~570–550 Ma (Jacobs et al., 1998, 2003a; Bauer et al., 2003). This early collisional phase was followed by granitoid magmatism and high-temperature metamorphism at ~530–490 Ma (Jacobs et al., 1998, 2003a; Paulsson and Austrheim, 2003) correlated with extensional collapse and lateral extrusion of the orogen (Jacobs et al., 2003c; Jacobs and Thomas, 2004).

6. Discussion

6.1. Integrating palaeomagnetic and geologic databases

Palaeomagnetic determination combined with precise thermochronology is the only tool we have to determine the absolute location of continental blocks on the Neoproterozoic globe. However, the Neoproterozoic palaeomagnetic record is very poor from many of the blocks that amalgamated to form Gondwana. Consequently, geological techniques are extremely valuable in helping resolve the many palaeomagnetically permissive permutations. This complementary approach has been used here to construct a geologically and palaeomagnetically feasible model for the amalgamation of India into Gondwana.

Further tests can, and should, be applied; including the location of climate specific lithologies (e.g. carbonate build-ups and glacial deposits)—although, the postulated extreme fluctuations in Neoproterozoic global climate need to be considered when doing this.

6.2. A note of caution on using isotopic ages of high-grade metamorphism to infer tectonic processes

Isotopic ages date the time diffusion of parent–daughter isotopes effectively ceased in a mineral phase. This can be because of cooling of a mineral through a characteristic closure temperature (different for each mineral species and isotopic system), or because of the growth of a mineral below its closure temperature. The ages obtained do not directly date any specific tectonic process. Because of this, correlating isotopic ages of high-grade metamorphism with collisional events is an inherently difficult process. For example, in a case where an isotopic age is a close approximation to the time of peak metamorphism of a rock (as it is often assumed with 'metamorphic' U–Pb zircon ages), metamorphism will have occurred an unknown period of time after the actual tectonic collision that caused it. To try and avoid this problem, in this study we have concentrated on linking available geochronological data with lithological, sedimentological, geochemical and palaeomagnetic data to develop a more holisitic tectonic model.

6.3. 800–700 Ma palaeogeography

Palaeomagnetic solutions for India, Congo/Tanzania/Bangweulu and Australia between ~800 Ma and ~750 Ma (Table 1, Fig. 2) suggest that during this time a high-latitude India (Torsvik et al., 2001b) moved south towards the near-constant sub-equatorial latitudes of Congo/Tanzania/Bangweulu and Australia (Powell and Pisarevsky, 2002; Pisarevsky et al., 2003). This model suggests that India was not a part of Rodinia, which is consistent with it being surrounded by late Neoproterozoic/Cambrian orogenic belts that formed during Gondwana amalgamation. Li et al. (2004) challenged this model and suggested that the apparent southward path of India and rotation of Congo and South China were better explained by inertial interchange true polar wander (Kirschvink et al., 1997). We note that the geological evidence is strongly in favour of India colliding with the Australia/Mawson Block in latest Neoproterozoic/Cambrian times, supporting the Neoproterozoic isolation of India and, in our opinion, negating the need for a non-uniformitarian explanation.

An 800–750 Ma high-latitude India with sub-equatorial Australia/Mawson and Congo/Tanzania/Bangweulu continents has considerable implications for the palaeogeography of the Mozambique and Mawson Oceans. The earlier Neoproterozoic reconstructions (e.g. Dalziel, 1997) envisaged a broadly orthogonal approach between East and West Gondwana, separated by a diminishing Mozambique Ocean. The recognition that India was likely to have been separated from Australia/Mawson in the Neoproterozoic led to the suggestion of a 'Mawson Sea' separating the two components of eastern Gondwana (Meert, 2003). In our reconstruction at ~750 Ma (Fig. 2), only one ocean separates the Congo/Tanzania/Bangweulu and Australia/Mawson Blocks. An oceanographic separation between a 'Mozambique ocean' and a 'Mawson Sea' only becomes meaningful later in the Neoproterozoic as India moves south relative to these continents.

Neoproterozoic dyke swarms in Western Australia have been used to suggest that a continental block rifted off Western Australia as Rodinia broke-up at ~750 Ma (Wingate and Giddings, 2000; Wingate and Evans, 2003). Palaeomagnetically permissive possibilities include Congo/Tanzania/Bangweulu, Kalahari (Pisarevsky et al., 2003), South China (Li et al., 2004) and the Tarim (from northern Western Australia, Chen et al., 2004). Similar ages of metamorphism in Dronning Maud Land (Jacobs et al., 1998) and the Northampton Complex of west Western Australia (Bruguier et al., 1999) have been used to support the case for Kalahari against Western Australia in Rodinia (Fitzsimons, 2002; Pisarevsky et al., 2003). We have followed this, and note that the increased dimensions of Kalahari with the Lurio–Vijayan peninsula are still compatible with this fit (Fig. 2).

6.3.1. Eastern Africa–Western India

The central Malagasy Itremo Group (Fig. 9) is interpreted as being sourced from the Congo/Tanzania/Bangweulu Block (Cox et al., 2004; Fitzsimons et al., 2004a) and was probably deposited at ~1700 Ma (Cox et al., 2004). If correct, this places Azania against East Africa at this time. Rifting between Azania and the Congo/Tanzania/Bangweulu Block occurred sometime after deposition of the Itremo Group, and may correlate with the ~870 Ma intrusion of anorthosites (Tenczer et al., 2004) in east Tanzania and deposition of rift-related sediments in Kenya at ~840–770 Ma (Key et al., 1989). This basin may have been floored with oceanic crust (Frisch and Pohl, 1985; Berhe, 1990; Wallbrecher et al., 2004) and is co-incident with juvenile 800–630 Ma granitoids found along the eastern granulite belt of Tanzania (Muhongo et al., 2001; Kröner et al., 2003b). These magmatic and basin formation processes in East Africa occurred at broadly the same time as 820–740 Ma arc-related gabbros and granitoids intruded Azania (Whitehouse et al., 1998; Handke et al., 1999; Kröner et al., 2000; Brewer et al., 2001) that have been related to subduction of the Mozambique Ocean along the site of the Betsimisaraka Suture in east Madagascar (Collins and Windley, 2002; Collins et al., 2003c). We suggest that these processes are linked, and that Azania rifted off East Africa because of roll-back of the Mozambique Ocean slab from a subduction zone located along the Betsimisaraka Suture.

Subduction of the Mozambique Ocean also occurred in this time frame along the west Indian margin (Ashwal et al., 2002). Extrusion and emplacement of the Malani Igneous Suite in NW India (Fig. 7) occurred between ~770 and 730 Ma (Crawford and Compston, 1970; Rathore et al., 1996; Torsvik et al., 2001b) and was coeval with magmatism in the Seychelles (Stephens et al., 1997; Tucker et al., 2001) and both magmatism and volcanism in the Bemarivo Belt of far north Madagascar (Tucker et al., 1999a).

6.3.2. Eastern India–Western Australia/Mawson

Magmatism along the eastern margin of India was restricted to the intrusion of anorthosites and granitoids in the Chilka Lake Domain (Krause et al., 2001; Dobmeier and Simmat, 2002) and granitoids in the Meghalaya Plateau in far northeast India (Fig. 7) (van Breeman et al., 1989; Ghosh et al., 1991, 1994). The Meghalaya granitoids have calc-alkaline chemistries (Ghosh et al., 1991) and may indicate that this region was an active margin in mid-Neoproterozoic times.

Kalahari is interpreted to have rifted off Western Australia at ~750 Ma, in the event that caused rift-related (Wilde and Murphy, 1990; Wilde, 1999) granitoid magmatism in the Leeuwin Complex (Nelson, 2002; Collins, 2003) and dyke swarms within cratonic Western Australia (Wingate and Giddings, 2000).

6.4. 700 – 600 Ma palaeogeography

Australia and Laurentia both lay in tropical latitudes during this time period (Table 1, Fig. 3), but reliable palaeomagnetic constraints do not exist for India, Kalahari or Congo/Tanzania/Bangweulu. Therefore, geological constraints for this time range are important in helping locate these continents. Tectonothermal events are especially common around the eastern, northern and western Congo/Tanzania/Bangweulu at this time. The tectonic evolution of the eastern margin of the Congo/Tanzania/Bangweulu Block is described in the following section. The western and northern margins are outside the main remit of this paper, but because of their relevance to the location of the Congo/Tanzania/Bangweulu Block in this time period, a brief outline of the timing of Neoproterozoic metamorphism and deformation is discussed below.

Along the north of the Congo/Tanzania/Bangweulu Block, peak Neoproterozoic metamorphic conditions are reported to have occurred at ~630 Ma in Uganda (Leggo, 1974; Appel et al., 2004), ~630 Ma in the Oubanguides Belt of the Central African Republic (Pin and Poidevin, 1987), and ~640–600 Ma in Nigeria, the Cameroon and the Dahomeyide Belt (Fig. 10) (Bernard-Griffiths et al., 1991; Dada, 1998; Affaton et al., 2000; Ferré et al., 2002; Toteu et al., 2004). In the Dahomeyide Belt, between the collage of pre-Neoproterozoic terranes that make up Nigeria (Dada, 1998) and the West African Craton, ^{40}Ar–^{39}Ar hornblende ages of 590–580 Ma provide a young age constraint on amalgamation of this part of Gondwana (Attoh et al., 1997). The deformation and metamorphism along the northern Congo/Tanzania/Bangweulu Block is interpreted as dating the collision between this block and the enigmatic Saharan Metacraton (Abdelsalam et al., 2002), and, possibly, a southern extension to the LATEA microcontinent best exposed in the Hoggar and Aïr mountains (Fig. 10) (Liégeois et al., 2003).

The western Congo/Tanzania/Bangweulu Block is thought to include the Sâo Francisco Craton of Brazil

(Fig. 12) (Brito Neves and Cordani, 1991; Trompette, 1994). The Neoproterozoic Araçuaí Belt of Brazil and the West-Congo Belt of central west Africa form a large embayment into this combined continent, which has been interpreted to represent thrusting of a completely intra-continental basin (Trompette, 1994), or the closure of a restricted oceanic basin (Pedrosa-Soares et al., 1992, 2001). The interpretation of this belt is key to the location of the Congo/Tanzania/Bangweulu Block between 700 and 600 Ma because the Sâo Francisco Craton collided with Amazonia along the Brazília and Araguaia Belts (Fig. 12) at ~ 650–600 Ma (Pimentel et al., 1991, 2000; Moura and Gaudette, 1993; de Alvarenga et al., 2000; Valeriano et al., 2004), thereby fixing the location of the combined Congo/Tanzania/Bangweulu–Sâo Francisco continent with respect to the combined Amazonia/Laurentia continent (Figs. 3–5).

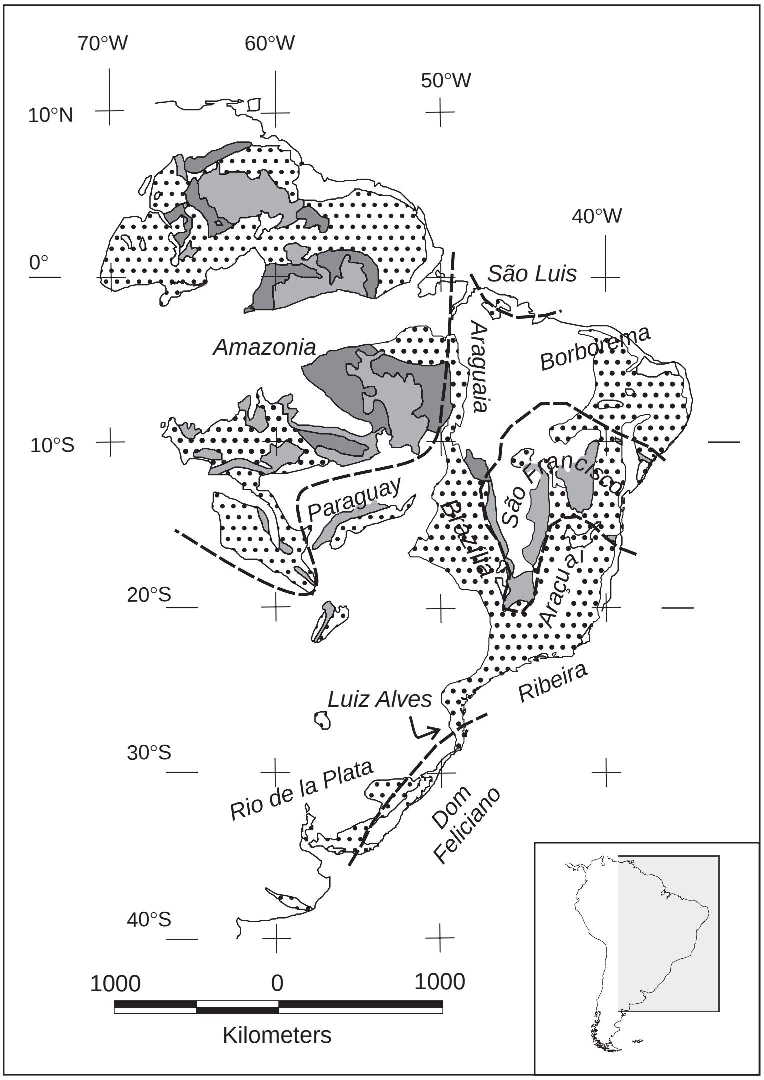

Fig. 12. Location and geological province map of central Africa (geology based on Commission for the Geological Map of the World, 2000). Archaean–Palaeoproteroozic cratonic blocks, Amazonia, Luiz Alves, Rio de la Plata, São Francisco, São Luis (part of the West African Block). Proterozoic orogenic belts = Araçuai, Araguaia, Dom Feliciano, Borborema, Brazília, Paraguay, Ribeira. Geological provinces after Cordani et al. (2000). Key as Fig. 7.

If the Sâo Francisco and the Congo/Tanzania/Bangweulu Block did not form the same continental mass, then the position of the Congo/Tanzania/Bangweulu Block is much more poorly constrained at this time.

6.4.1. Eastern Africa–Western India

High-grade ~655–610 Ma metamorphism in Tanzania, southern Madagascar and northern Mozambique (Fig. 11) was coeval with final suturing in the Arabian–Nubian Shield (Johnson and Woldehaimanot, 2003). The relatively high pressures (10–14 kbar) and high temperatures (800–850 °C) of peak metamorphic conditions (Appel et al., 1998; Sommer et al., 2003) in the Tanzania granulites are consistent with closure of a young, hot, marginal basin between the Congo/Tanzania/Bangweulu Block and Azania (Fig. 3). The slow cooling rate (2–5 °C/Ma) (Möller et al., 2000) suggests that post-orogenic collapse was not a major factor in exhuming the Tanzanian part of the orogen. In contrast, the chemistry and structural setting of ~630 Ma alkaline magmatism in Azania (central Madagascar) has been interpreted as indicating magmatism in an extensional post-collisional setting (Nédélec et al., 1994, 1995; Paquette and Nédélec, 1998). These tectonic models can be linked by having post-collisional extension localised in the east of the orogen due to renewed subduction and rollback along the site of the Betsimisaraka Suture.

The western Indian margin was stable during this time period, with only limited magmatic activity being reported from southern India (Ghosh et al., 2004; Rajesh, 2004).

In the model presented here, Azania collides with both the Congo/Tanzania/Bangewulu Block and the Saharan Metacraton at approximately the same time as the latter two cratons collided (see preceding section) (Fig. 3). This model links the tectonic evolution of the Tanzanian Mozambique Belt with the Arabian Nubian Shield in a similar manner to the original model of the East African Orogeny (Stern, 1994), and the more recent model of Meert (2003). Our model differs from these in that India has not collided with any African cratons by 600 Ma, and open ocean existed east of Azania until the latest Neoproterozoic (Figs. 3–5).

6.4.2. Eastern India–Western Australia/Mawson

Both the eastern Indian and Western Australian margins are interpreted as open, oceanic margins at this time (Fig. 3). Granitoid intrusion continued in the Meghalaya plateau at this time (van Breeman et al., 1989; Ghosh et al., 1991, 1994), suggesting a continuation of subduction in this region. Further south, granulite-facies metamorphism has been reported in the Chila Lake Domain of the Eastern Ghats, between 690 and 662 Ma, in a localised, intraplate setting (Dobmeier and Simmat, 2002).

On the Western Australian margin, ~615 Ma metamorphism had been reported from the Leeuwin Complex (Nelson, 1996). This was used by Powell and Pisarevsky (2002) as evidence for the amalgamation of Australia and India at this time. However, this timing for metamorphism and deformation was challenged by Collins (2003) who showed that metamorphism was ~80 Ma younger and occurred in latest Neoproterozoic/Cambrian times.

6.5. 600 – 530 Ma palaeogeography

India collided obliquely with both Australia/Mawson and the previously amalgamated Azania–Congo/Tanzania/Bangweulu–Amazonia–Rio de la Plata–West Africa continent at this time. In our reconstruction (Figs. 4–6), we follow recent geochronological work and interpret Kalahari to have docked with both the Congo/Tanzania/Bangweulu Block along the Zambezi (John et al., 2003; Johnson and Oliver, 2004) and Damara Belts (Jung et al., 2000; Jung and Mezger, 2003a,b), and the Rio de la Plata Craton along the Gariep Belt (Fig. 1) (Frimmel and Frank, 1998; Frimmel and Fölling, 2004) in latest Neoproterozoic/Cambrian times. This late amalgamation of Kalahari into Gondwana is also interpreted to have caused the latest Neoproterozoic to Cambrian Rio Doce Orogeny (or Búzios Orogeny) in the Ribeira Belt of southern Brazil (Fig. 12) (Campos Neto and Figueiredo, 1995; Schmitt et al., 2004).

6.5.1. Indian margins

India collided with Azania–Congo/Tanzania/Bangweulu along the Betsimisaraka Suture in the Early Cambrian (Collins et al., 2003c), causing widespread orogenesis in Azania. The suture was interpreted by Collins and Windley (2002) to pass north into the Horn of Africa where it outcrops in northern Somalia as the Maydh (or Mait) Complex (Sassi et al., 1993).

Central Madagascar was thrust east over the margin of India (represented there by the Antongil Block) (Collins et al., 2003a) causing high-grade metamorphism of the Neoproterozoic metasediments caught in the Betsimisaraka Suture at ~520 Ma (Collins et al., 2003c; Fitzsimons et al., 2004a). Granulite-facies metamorphism also took place in central and western Madagascar at ~560–550 Ma (Kröner et al., 2000; Cox et al., 2004). In the latter region metamorphism is associated with E-directed thrusting of the 620–560 Ma Molo Group metasediments (Cox et al., 2004) that were deposited in an intracontinental setting after the ~630 Ma amalgamation of Azania with Congo/Tanzania/Bangweulu. Ultra-high temperature metamorphism also peaked at this time in southern India (Braun et al., 1998; Braun and Bröcker, 2004).

On the other side of India, high-grade metamorphism and contractional deformation occurred at this time all along the eastern and south eastern margin of Neoproterozoic India (now preserved in Antarctica and south Western Australia) (Fitzsimons, 2000b).

6.6. Comparisons with other models

In this contribution, we have reviewed the geochronology and tectonothermal history of orogens that marked the western, eastern and southern margins of a Neoproterozoic India and were responsible for the assembly of eastern Gondwana. We have correlated these geological data with palaeogeographic information provided by the available palaeomagnetic solutions to develop a model for the Neoproterozoic geography of eastern Gondwana. A number of previous works have also addressed this issue. Below we outline the main similarities and differences between these models and the model presented here.

Meert and co-workers (Meert and Van der Voo, 1997; Meert, 2001, 2003) were amongst the first to recognise a bi-modality in geochronological data from eastern Gondwana and suggest that the region assembled during two orogenies. They suggested the presence of an earlier East African Orogeny between ~750 and 620 Ma and a later orogeny at ~570 to 530 Ma, termed the Kuunga Orogeny by Meert et al. (1995). This later orogeny was interpreted to mark the collision of Australia and Antarctica with the rest of Gondwana and was subsequently correlated with a broad belt of orogenesis from the Damara Orogen in the west to the Pinjarra Orogen in the east, with a southern spur to Dronning Maud Land (Fig. 1) (Meert, 2003). The palaeogeographic model proposed by these workers involved a Neoproterozoic continent consisting of Sri Lanka, Madagascar and India (SLAMIN) colliding with a combined Congo/Kalahari continent at ~750–620 Ma, followed by Australia/East Antarctica colliding with the bulk of Gondwana at ~570–530 Ma (Meert, 2003; Meert and Torsvik, 2003).

Boger and Miller (2004) pointed out that India did not collide with western Gondwana until latest Neoproterozoic/Early Cambrian times and proposed a variation on the SLAMIN model of Meert (2003) where the East African Orogen evolved as an accretionary orogen and was partially superimposed by a 590–560 Ma orogen created by the collision of a combined India, Madagascar and part of Antarctica with eastern Africa. They termed this orogen the Mozambique suture. In their model, Australia–Antarctica (with an enlarged Antarctic component including the Ruker Terrane) collided with India along the 'Kuunga suture' at 535–520 Ma.

In the model we present here, we follow Meert and co-workers by sub-dividing Neoproterozoic to Cambrian 'Pan-African' events into pre-600 Ma orogenesis and later ~570–500 Ma orogenesis. However, we point out that both of these time frames encompass many terrane amalgamation events (both accretionary and collisional). In the next couple of paragraphs we discuss the rationale for our subdivision of orogenies in eastern Gondwana.

The areal extent of a collisional orogen is limited by the geometry of the colliding blocks. Hence, an orogeny caused by the collision of Azania with the Congo/Tanzania/Bangweulu Block cannot extend south of the southern limit of Azania. We, therefore, propose that the ~650–630 Ma East African Orogeny (as opposed to the East African Orogen) be restricted to the event in the Arabian–Nubian Shield and the Mozambique Belt caused by the collision of Azania with the Congo/Tanzania/Bangweulu Block. There are many pre-650 Ma terrane accretion events in the East African Orogen, but we argue that these events are spatially and temporally distinct and should not be lumped together.

The Kuunga Orogeny was originally defined on geochronological data and interpreted to be related to the collision between Australia/Antarctica and an already combined India/East Africa (Meert et al., 1995). In recent years, a number of workers have demonstrated that India did not collide with the Congo/Tanzania/Bangweulu Block until latest Neoproterozoic/Early Cambrian times (Collins et al., 2003c; Boger and Miller, 2004), and the Kuunga Orogeny, or suture has become associated with the India–Australia/Mawson collision (Meert, 2003; Boger and Miller, 2004). We follow this usage and suggest the Kuunga Orogeny be restricted to ~570–500 Ma (note the slight change in age range from Meert, 2003) orogenesis related to this India–Australia/Mawson collision.

The Betsimisaraka Suture in east Madagascar (Collins et al., 2000) is interpreted as the site of the India–Azania collision and a provenance boundary between East African-derived terranes and Indian terranes (Collins and Windley, 2002; Collins et al., 2003c). Deformation and metamorphism in east Madagascar is Early Cambrian in age and we interpret collision here to be coeval with the Kuunga Orogeny the other side of India (Boger and Miller, 2004). We suggest that this orogenic event be called the Malagasy Orogeny after the place where it is best demonstrated.

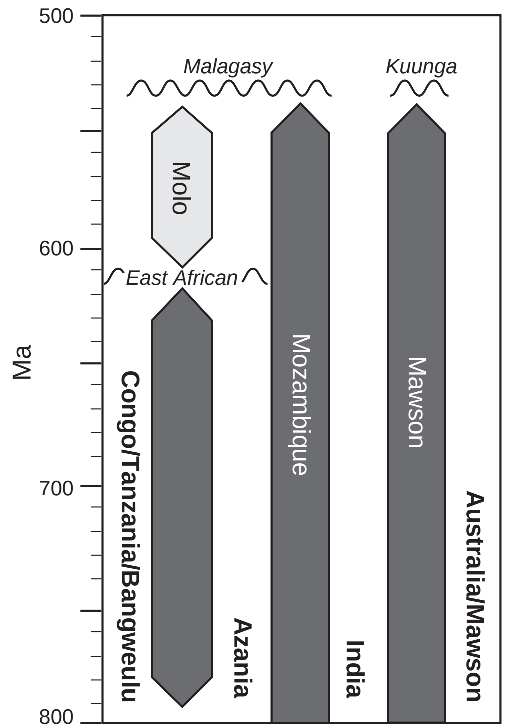

Fig. 13. Schematic summary representation of the main Neoproterozoic continental blocks around India (Congo/Tanzania/Bangweulu, Azania, India and Australia/Mawson), the timing of closure of the Mozambique Ocean, the Mawson Sea and the unnamed ocean between Azania and the Congo/Tanzania/Bangweulu Block (dark grey blocks) and the intracontinental Molo basin between Azania and the Congo/Tanzania/Bangweulu Block (light grey block). Also shown are the orogenies that developed because of these collisions (the East African, Malagasy and Kuunga).

7. Conclusions

The tectonothermal evolution of the orogenic belts that delineate Neoproterozoic India has revealed:

(1) Large tracts of pre-Neoproterozoic crust reworked within Neoproterozoic/Cambrian orogens can be used to better delineate the geometry of the major Neoproterozoic continental blocks and allow the recognition of previously unidentified microcontinents.
(2) The recognition that one of these microcontinents (Azania) collided with the Congo/Tanzania/Bangweulu Block at ~630 Ma (Figs. 3 and 13), causing granulite-facies metamorphism in east Tanzania and terminating accretion and arc–arc collision in the Arabian/Nubian Shield.
(3) 550–520 Ma granulite-facies metamorphism and contractional deformation occur to both the west and east of Neoproterozoic India, suggesting final collision between India and the Congo/Tanzania/Bangweulu Block occurred coeval with the collision between India and the Australia/Mawson Block (Figs. 4–6 and 13), and that both these collisions occurred in the latest Neoproterozoic/Cambrian (cf. Meert, 2003).

These geological data are combined with the available palaeomagnetic data to produce a non-unique, but permissive model for the incorporation of India into Gondwana and the final amalgamation of eastern Gondwana from ~750 to 530 Ma.

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