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Assembly, configuration, and break-up history of Rodinia: A synthesis

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Abstract

This paper presents a brief synthesis of the current state of knowledge on the formation and break-up of the early-Neoproterozoic supercontinent Rodinia, and the subsequent assembly of Gondwanaland. Our discussions are based on both palaeomagnetic constraints and on geological correlations of basement provinces, orogenic histories, sedimentary provenance, the development of continental rifts and passive margins, and the record of mantle plume events.

Rodinia assembled through worldwide orogenic events between 1300 Ma and 900 Ma, with all, or virtually all, continental blocks known to exist at that time likely being involved. In our preferred Rodinia model, the assembly process features the accretion or collision of continental blocks around the margin of Laurentia. Like the supercontinent Pangaea, Rodinia lasted about 150 million years after complete assembly. Mantle avalanches, caused by the sinking of stagnated slabs accumulated at the mantle transition zone surrounding the supercontinent, plus thermal insulation by the supercontinent, led to the formation of a mantle superswell (or superplume) beneath Rodinia 40–60 million years after the completion of its assembly. As a result, widespread continental rifting occurred between ca. 825 Ma and 740 Ma, with episodic plume events at ca. 825 Ma, ca. 780 Ma and ca. 750 Ma.

Like its assembly, the break-up of Rodinia occurred diachronously. The first major break-up event occurred along the western margin of Laurentia (present coordinates), possibly as early as 750 Ma. Rifting between the Amazonia craton and the southeastern margin of Laurentia started at approximately the same time, but only led to break-up after ca. 600 Ma. By this time most of the western Gondwanan continents had joined together, although the formation of Gondwanaland was not complete until ca. 530 Ma.

Keywords: Rodinia; Supercontinent; Neoproterozoic; Mesoproterozoic; Grenvillian; Continental rift; Mantle plume; Gondwanaland

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1. Introduction

Valentine and Moores (1970) were probably the first to recognise that a supercontinent, comprising just about all continents on Earth, existed towards the end of the Precambrian. They suggested that the break-up of the supercontinent, which they called Pangaea I, by the Cambrian led to a divergence of environments and characteristics amongst the daughter continents, and a sudden prominence of shallow, nutrient-rich shelves and coastal areas, all factors conducive to the diversification of life forms on Earth. This late-Precambrian supercontinent was later renamed Rodinia (McMenamin and McMenamin, 1990) from the Russian word 'rodit' meaning 'to beget' or 'to give birth'. McMenamin and McMenamin (1990) considered Rodinia to have been the supercontinent that spawned all subsequent continents, while "the edges (continental shelves) of Rodinia were the cradle of the earliest animals" (McMenamin and McMenamin, 1990, p. 95).

The concept of the supercontinent Rodinia attracted much attention in 1991, when three researchers (Moores, 1991; Dalziel, 1991; Hoffman, 1991) published geological evidence for the assembly and break-up of Rodinia, with some of its daughter continents forming Gondwanaland. Common to their propositions is the connection between western Laurentia, Australia and East Antarctica (commonly known as the southwest U.S.–East Antarctic, or SWEAT, connection; Moores, 1991), which follows an earlier suggestion by Bell and Jefferson (1987) based on stratigraphic correlations and palaeomagnetic constraints across the Pacific, and by Eisbacher (1985) based on stratigraphic correlation alone.

An explosion of new data and ideas occurred in the following years, and, in order to better coordinate the global efforts in testing the Rodinia hypothesis, a UNESCO-IGCP project (No. 440, 1999–2004) was established to investigate the formation, configuration, and break-up of Rodinia, and to construct an interpretative Geodynamic Map of the Rodinia supercontinent (the Rodinia Map hereafter; Appendix I, see hard copy in the print version of this volume; also available online). Although few still doubt the existence of a late Precambrian supercontinent, there is still no consensus regarding the number of participating cratons, their relative configuration within the supercontinent and the chronology and mode of assembly and break-up of the supercontinent. In this paper we provide an overview of evidence for and against major Rodinian reconstructions, including those in the Rodinia Map (Appendix I). We discuss some current ideas regarding the formation of Rodinia and processes that led to its break-up, and present an animation for the evolution of Rodinia from 1100 Ma till the formation of Gondwanaland at 530 Ma (Appendix II, available online).

The challenges to reconstructing the history of Rodinia include inadequate high-quality geological, geochronological and palaeomagnetic data, multiple possible interpretations for each data set, and uncertainties in fundamental assumptions such as the application of modern-style plate tectonics to late-Precambrian time and that the geomagnetic field was a geocentric axial dipole field (an assumption that underlines interpretations of palaeomagnetic data). In this paper we assume that modern-style plate tectonics, with the possible complication of true polar wander (TPW), apply to late Precambrian (e.g., Stern, 2005), and that the geomagnetic field was dominantly a geocentric axial dipole field at that time (Evans, 2006). We emphasise the importance of considering multiple lines of evidence in testing any reconstruction, because only through such an approach can the potentially large number of solutions be reduced to the most likely scenarios (Fig. 1a). We strive to be as objective as possible when discussing alternative interpretations, at the same time trying to make a self-consistent synthesis. We emphasise that not every opinion expressed in this paper is agreed on by all co-authors.

2. Major inter-continental connections proposed for Rodinia using multidisciplinary evidence

Because Laurentia is flanked by Neoproterozoic passive margins, it is commonly regarded as being at the centre of Rodinia assembly and break-up (e.g., Hoffman, 1991). We will evaluate the various continental connections around the margin of Laurentia proposed for Rodinia time.

2.1. Continents facing the present western and northern margins of Laurentia: Australia–East Antarctica, South China, or Siberia?

2.1.1. The SWEAT hypothesis

The question of which continent(s) used to lie adjacent to the present western margin of Laurentia during the existence of Rodinia has been central to the Rodinia debate, and the topic remains controversial today. Among the competing models, the earliest and perhaps the best-known model is the SWEAT hypothesis (Fig. 2a). The model builds on the similar Neoproterozoic stratigraphy of the western margin of Laurentia and eastern Australia as recognised earlier by Eisbacher (1985) and Bell and Jefferson (1987), and elaborated later by Young (1992) and Rainbird et al. (1996) among others. Three landmark papers in 1991 (Moores, 1991; Dalziel, 1991; Hoffman, 1991) also used basement outcrops as piercing points that may correlate from one continent to the other, such as the correlation of Grenville-age (ca. 1300–1000 Ma) orogenic belts, to argue that the SWEAT connection probably existed as early as 1900 Ma ago and lasted until mid to late Neoproterozoic times. Moores (1991) implied that the united northern and western Australian craton might not have become part of the SWEAT connection until its collision with the Gawler craton along the late Mesoproterozoic Albany-Fraser-Musgrave belt. Powell et al. (1993) confirmed that a SWEAT connection was palaeomagnetically permissible for the time interval of ca. 1050–720 Ma based on the contemporary palaeomagnetic database, but would have to have broken apart by 580 Ma (or >650 Ma as pointed out earlier by Van der Voo et al., 1984, on palaeomagnetic grounds).

Closer scrutiny of the crustal provinces and basin provenance analysis, however, have shown that geological discontinuities exist across the SWEAT connection. In particular, isotopic and geochemical mapping along the Transantarctic Mountains has revealed that Mesoproterozoic basement provinces as young as ca. 1000 Ma along the continental margin truncate the Lau-

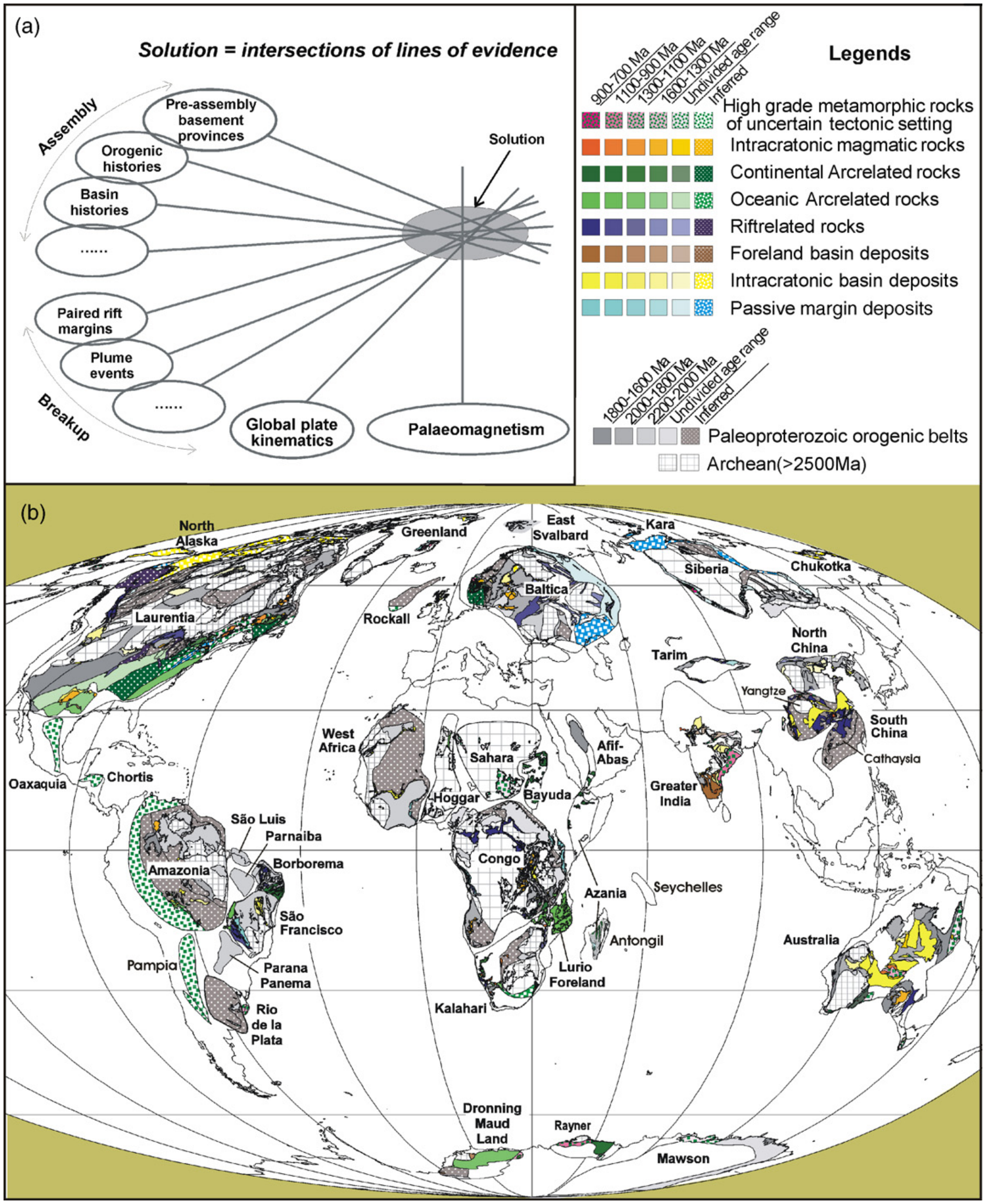

Fig. 1. (a) A schematic diagram illustrating the importance of taking a multidisciplinary and global approach in reconstructing palaeogeography in relation to the evolution of a Precambrian supercontinent (when biogeography is of limited value). The line from each research aspect symbolises an infinite number of possible solutions, with concentration of their intersections giving the most likely solution. (b) Present locations and major crustal elements of the Precambrian continental blocks.

rentian crustal provinces (Borg and DePaolo, 1994). To make the SWEAT connection viable, these authors suggested that those basement blocks were probably exotic in origin and were accreted to the East Antarctic margin after the opening of the proto-Pacific Ocean. Although this may be so for some of the terranes along the Cambrian Ross–Delamerian Orogen (Stump, 1995), there is no geological evidence for the docking of terranes like the Beardmore terrane, which would have to have occurred in a very narrow time window between the break-up of the SWEAT connection (ca. 750 Ma, see later discussions) and the development of the Beardmore Group passive-margin deposits no later than 668 ± 1 Ma (Goodge et al., 2002).

There are other lines of evidence that do not support the SWEAT connection. (1) There is a lack of continuation of the ca. 1400 Ma (1500–1350 Ma) transcontinental magmatic province of southern Laurentia (e.g., Nyman et al., 1994; Van Schmus et al., 1996) into the Transantarctic Mountains, although one could argue that they could be beneath the ice farther inland (Goodge et al., 2002). (2) There are mismatches in the Neoproterozoic mantle plume record across the SWEAT connection (Park et al., 1995; Wingate et al., 1998; Li et al., 1999; see more discussion in Section 2.1.2). (3) The Belt Basin of western North America and the possibly overlying Buffalo Hump Formation (the Deer Trail Group) require a western clastic source with rocks of 1786–1642 Ma, 1600–1590 Ma and 1244–1070 Ma, much of which cannot easily be identified in East Antarctica in the SWEAT configuration (Ross et al., 1992; Ross and Villeneuve, 2003). (4) The originally proposed continuation of

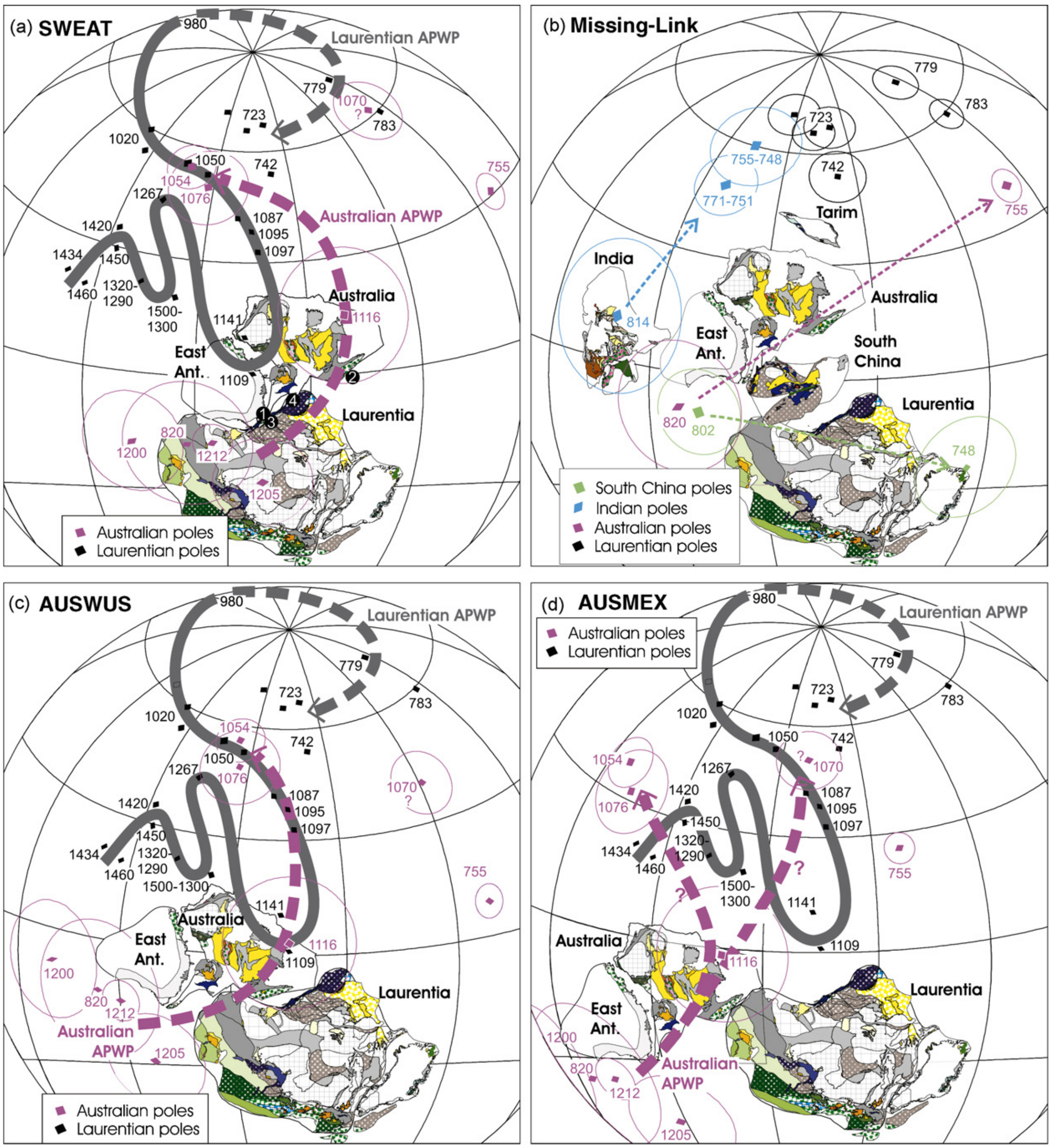

Fig. 2. Palaeomagnetic examinations of alternative configurations of continents west and north of Laurentia. Palaeopoles used are listed in Table 1, and Euler rotation parameters given in Appendix III (available online). (a) In the SWEAT fit (after Moores, 1991; Dalziel, 1991; Hoffman, 1991) palaeopoles from Laurentia (grey diamonds) and Australia (purple diamonds) do not merge until ca. 1050 Ma. (b) In the "Missing-Link" configuration (after Li et al., 1995) ca. 820–800 Ma poles from India, Australia and South China fall close to each other but the ca. 750 Ma poles are scattered. No dated ca. 820–800 Ma pole is available from Laurentia for a comparison. (c) In the AUSWUS configuration (Karlstrom et al., 1999; Burrett and Berry, 2000) ca. 1200 Ma poles from Laurentia and Australia fall >30° apart, whereas some of their 1100–1050 Ma poles follow the same apparent polar wander path (APWP). (d) In the AUSMEX configuration (Wingate et al., 2002) five of the seven ca. 1200–1050 Ma poles from Australia fall away from coeval Laurentian poles. (e) Siberia placed against Laurentia following Sears and Price (1978, 2000), in which the Siberian poles fall away from the Laurentia poles. (f) Siberia–Laurentia configuration following Pisarevsky et al. (2008), in which ca. 1050–970 Ma Siberian poles follow the same APWP as the Laurentian poles. In this figure and Figs. 5–7, all continents were rotated to Laurentia which is at its 900 Ma palaeolatitude. Latitude and longitude lines are shown in 30° intervals. Geotectonic polygon data are shown following the legend for the Rodinia Map (as in Figs. 1b and 8 and Appendix I).

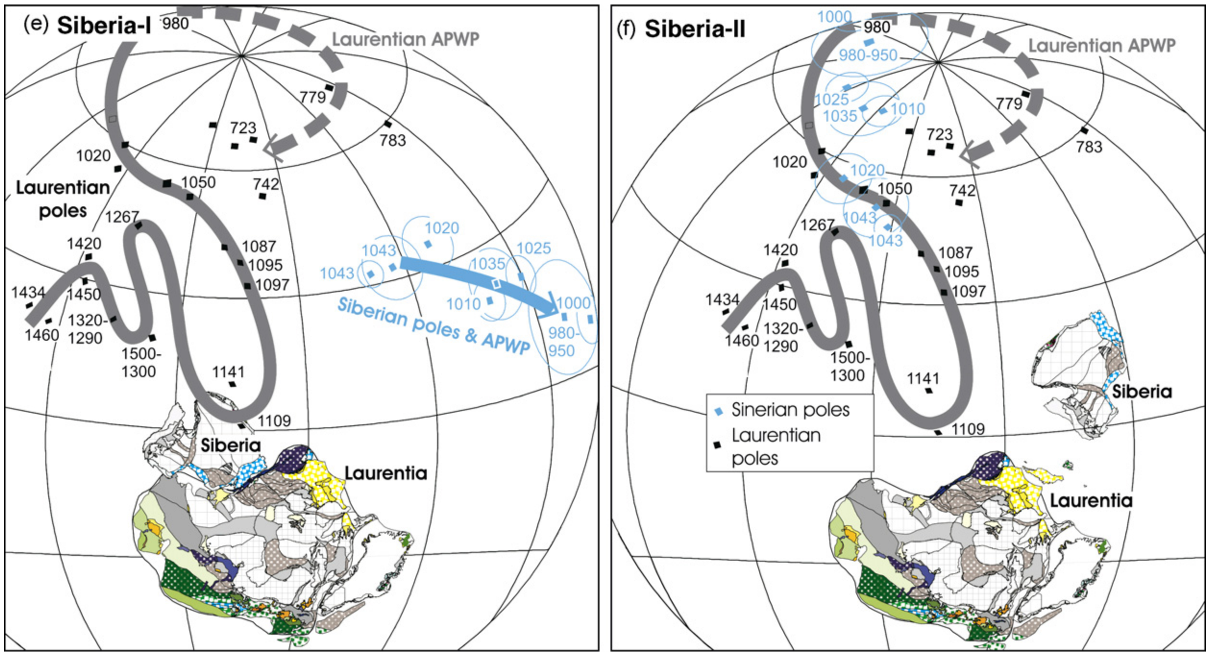

Fig. 2. (*Continued*).

the Grenville belt into the Coats Land and Dronning Maud Land of East Antarctica (Moores, 1991; Dalziel, 1991) was later proved untenable both palaeomagnetically (e.g., Gose et al., 1997) and geologically (e.g., Jacobs et al., 2003a,b). Coats Land was more likely part of Laurentia at 1110 Ma (e.g., Gose et al., 1997), but became connected with Kalahari when Kalahari collided with Laurentia at 1090–1060 Ma, and remained connected until the break-up of Gondwanaland (e.g., Jacobs et al., 2003a,b, 2008).

There is also a growing body of evidence suggesting the possible presence of Grenvillian orogenic belts along both the western margin of Laurentia and eastern Australia. On the Australia–East Antarctica side, apart from the possible presence of late Mesoproterozoic basement provinces along the Transantarctic Mountains (Borg and DePaolo, 1994), Berry et al. (2005) recently reported a 1287 ± 18 Ma metamorphic age (monazite U–Th–Pb) from King Island northwest of Tasmania (#1 in Fig. 2a), and Fioretti et al. (2005) published a 1119 ± 9 Ma age for quartz syenite at the South Tasman Rise. Like the basement for the Transantarctic Mountains, an allochthonous origin for northwestern Tasmania and King Island is possible, but the following observations argue against such a possibility: (a) the palaeomagnetic record shows that northwestern Tasmania was already part of cratonic Australia by the Late Cambrian (Li et al., 1997), and there is no prior compressional event recognised in the region that reflects terrane amalgamation between the time of Rodinia break-up (ca. 750 Ma, see Section 3.2) and the Late Cambrian; (b) the Neoproterozoic tectonostratigraphy of northwestern Tasmania and King Island can be well correlated with that of the Adelaide Fold Belt (Calver and Walter, 2000; Holm et al., 2003; Li, 2001; Z.X. Li et al., 2003b). A possible Grenvillian province has also been identified in northern Queensland (#2 in Fig. 2a; Blewett et al., 1998; Hutton et al., 1996), largely based on detrital zircon ages.

A number of studies reporting Grenvillian orogenic events along the western margin of Laurentia have yet to attract much attention. Perhaps most significant are the 1090–1030 Ma metamorphic ages (titanite U–Pb) from 1468 ± 2 Ma mafic sills in the Belt-Purcell Supergroup (#3 in Fig. 2a; Anderson and Davis, 1995). In the Mackenzie Mountains (#4 in Fig. 2a), there is evidence for an east-west compressional event (the Corn Creek Orogeny) that occurred after 1033 Ma but before 750 Ma (Thorkelson, 2000; Thorkelson et al., 2005). There is also a less-well defined 1175–1100 Ma recrystallisation age from a granite clast brought up from the basement in a diatreme (Jefferson and Parrish, 1989).

If there were indeed Grenvillian orogenic belts between Laurentia and Australia–East Antarctica, it would suggest that even if a SWEAT connection existed, it would have to have been Grenvillian or later.

Palaeomagnetic data permit a SWEAT connection after ca. 1050 Ma (Powell et al., 1993), but such a connection is not possible at ca. 1200 Ma (Fig. 2a; Pisarevsky et al., 2003a) although the reliability of the ca. 1200 Ma pole for Australia needs further verification. This scenario is consistent with pre-Grenvillian geological mismatches discussed above. However, we do not preclude the possible existence of a SWEAT-like link between Australia and Laurentia during late-Paleoproterozoic (ca. 1800–1600 Ma), as suggested by Idnurm and Giddings (1995) and Betts and Giles (2006).

2.1.2. The "Missing-Link" model

The "Missing-Link" model proposed by Li et al. (1995) has the South China Block sitting between Australia–East Antarctica and Laurentia in Rodinia, serving as the "missing-link" between the two continents (Fig. 2b). The model was initially developed in view of: (a) mismatches in the crustal provinces of Australia–East Antarctica and Laurentia (see discussions in

Table 1
Selected palaeomagnetic poles used for reconstructing the late-Mesoproterozoic to Neoproterozoic global palaeogeography

Rock unit	Age (Ma)	Pole		A_{95} (°)	Reference
		(°N)	(°E)		
Laurentia					
Long Range dykes[a]	620–610	19	355	18	Murthy et al. (1992); Kamo and Gower (1994)
Franklin dykes	723 + 4/−2	5	163	5	Heaman et al. (1992); Park (1994)
Natkusiak Formation	723 + 4/−2	6	159	6	Palmer et al. (1983); Heaman et al. (1992)
Kwagunt Formation	742 ± 6	18	166	7	Weil et al. (2004)
Tsezotene sills and dykes	779 ± 2	2	138	5	Park et al. (1989); LeCheminant and Heaman (1994)
Wyoming dykes	782 ± 8; 785 ± 8	13	131	4	Harlan et al. (1997)
Galeros Formation	780–820	−2	163	6	Weil et al. (2004)
Haliburton intrusions A	980 ± 10; 1000–1030	−36	143	10	Buchan and Dunlop (1976); Warnock et al. (2000)
Chequamegon sandstone	∼1020[b]	−12	178	5	McCabe and Van der Voo (1983)
Jacobsville sandstone J (A+B)	∼1020[b]	−9	183	4	Roy and Robertson (1978)
Freda sandstone	1050 ± 30	2	179	4	Henry et al. (1977); Wingate et al. (2002)
Nonesuch shale	1050 ± 30	8	178	4	Henry et al. (1977); Wingate et al. (2002)
Lake Shore Traps	1087 ± 2	22	181	5	Diehl and Haig (1994); Davis and Paces (1990)
Portage Lake volcanics	1095 ± 2	27	181	2	Halls and Pesonen (1982); Davis and Paces (1990)
Upper North Shore volcanics	1097 ± 2	32	184	5	Halls and Pesonen (1982); Davis and Green (1997)
Logan sills R	1109 + 4/−2	49	220	4	Halls and Pesonen (1982); Davis and Sutcliffe (1985)
Abitibi dykes	1141 ± 2	43	209	14	Ernst and Buchan (1993)
Wind River, Gr. B	1300–1500	22	209	9	Harlan et al. (2003a,b)
Mackenzie dolerite dykes	1267 ± 2	4.0	190.0	5.0	Buchan and Halls (1990)
Nain anorthosite	1320–1290	12.0	210.0	3.0	From Buchan et al. (2000)
Mistastin complex	∼1420	−1.0	201.0	8.0	From Buchan et al. (2000)
Laramie complex and Sherman granite	∼1434	−7.0	215.0	4.0	From Buchan et al. (2000)
Harp Lake complex	∼1450	2.0	206.0	4.0	From Buchan et al. (2000)
Michikamau anorthosite pluton	∼1460	−2.0	218.0	5.0	From Buchan et al. (2000)
Molson dykes, component A	∼1820–1720	15.4	263.5	4.0	From Buchan et al. (2000)
Baltica					
Hunnedalen dykes	>848	−41	222	10	Walderhaug et al. (1999)
Egersund-Ogna anorthosite	∼900	−42	200	9	Brown and McEnroe (2004)
Egersund anorthosite	929–932	−44	214	4	Stearn and Piper (1984); Torsvik and Eide (1998)
Hakefjorden	916 ± 11	5	249	4	Stearn and Piper (1984); Scherstén et al. (2000)
Göteborg-Slussen	935 ± 3	−7	242	12	Pisarevsky and Bylund (2006)
Dalarna dykes	946 ± 1	5	239	15	Bylund and Elming (1992), Söderlund et al. (2005)
Karlshamn-Fäjö dykes	946–954	2	242	30	Patchett and Bylund (1977), Söderlund et al. (2004)
Nilstorp dyke	966 ± 2	9	239	8	Patchett and Bylund (1977), Söderlund et al. (2004)
Pyätteryd amphibolite	933–945	−43	214	11	Pisarevsky and Bylund (1998); Wang et al. (1996); Wang and Lindh (1996)
Känna gneiss	948–974	−50	225	17	Pisarevsky and Bylund (1998); Wang et al. (1996); Wang and Lindh (1996)
Gällared amphibolite	956?	−46	214	19	Pisarevsky and Bylund (1998); Möller and Söderlund (1997)
Gällared granite-gneiss	980–990	−44	224	6	Pisarevsky and Bylund (1998); Möller and Söderlund (1997)
Bamble intrusions	1100–1040	3	217	15	Torsvik and Eide (1998); Brown and McEnroe (2004)
Laanila dolerite	1045 ± 50	−2	212	15	Mertanen et al. (1996)
Mean Jotnian dolerite intrusions	∼1265	4.0	158.0	4.0	From Buchan et al. (2000)
Åland quartz porphyry dykes	1571 ± 20; 1571 ± 9	12.0	182.0	7.0	From Buchan et al. (2000)
Åland dolerite dykes	1577 ± 12	28.0	188.0	9.0	From Buchan et al. (2000)
Subjotnian quartz porphyry dykes	∼1630	29.0	177.0	6.0	From Buchan et al. (2000)
Shoksha Formation, Vepsian Group	∼1780	39.7	221.1	4.0	From Buchan et al. (2000)

Table 1 (*Continued*)

Rock unit	Age (Ma)	Pole (°N)	Pole (°E)	A_{95} (°)	Reference
India					
Mahe dykes, Seychelles[c]	748–755	80	79	11	Torsvik et al. (2001b)
Malani Igneous Suite	751–771	68	88	8	Torsvik et al. (2001a)
Harohalli dykes	814 ± 34	27	79	18	Radhakrishna and Mathew (1996)
Wajrakarur kimberlites	1079?	45	59	11	Miller and Hargraves (1994)
Australia					
Elatina Formation	600–620	51	157	2	Embleton and Williams (1986)
Elatina Formation	600–620	54	147	1	Schmidt et al. (1991)
Elatina Formation	600–620	52	167	11	Schmidt and Williams (1995)
Elatina Formation	600–620	39	186	9	Sohl et al. (1999)
Yaltipena Formation	620–630	44	173	8	Sohl et al. (1999)
Mundine Well dykes	755 ± 3	45	135	4	Wingate and Giddings (2000)
Walsh Tillite	750–770	22	102	14	Li (2000)
Hussar Formation	800–760	62	86	10	Pisarevsky et al. (2007)
Browne Formation	830–800	45	142	7	Pisarevsky et al. (2007)
Wooltana Formation	∼820	62	322	17	McWilliams and McElhinny (1980)
Kulgera dykes	1054 ± 13	−8	75	6	Schmidt et al. (2005)
Bangemall Basin sills	1070 ± 6	34	95	8	Wingate et al. (2002)
Stuart dykes	1076 ± 33	10	262	10	Idnurm and Giddings (1988)
Lakeview dolerite	1116 ± 12	10	311	17	Tanaka and Idnurm (1994)
Bremer Bay & Whalebone Pt	∼1200	74	304	13	Pisarevsky et al. (2003a)
Mt. Barren Group	1205 ± 40	47	347	13	Pisarevsky et al. (2003a), Dawson et al. (2003)
Fraser dyke	1212 ± 10	56	3260	5	Pisarevsky et al. (2003a)
Congo					
Mbozi Complex, Tanzania	755 ± 25	46	325	9	Meert et al. (1995); Evans (2000)
Gagwe lavas, Tanzania	795 ± 7	25	93	10	Meert et al. (1995); Deblond et al. (2001)
São Francisco					
Ilheus dykes	1011 ± 24	30	100	4	D'Agrella-Filho et al. (1990); Renne et al. (1990)
Olivença dykes, normal	∼1035[b]	16	107	8	D'Agrella-Filho et al. (1990); Renne et al. (1990)
Itaju de Colonia	∼1055[b]	8	111	10	D'Agrella-Filho et al. (1990); Renne et al. (1990)
Olivenca dykes (rever.)	1078 ± 18	−10	100	9	D'Agrella-Filho et al. (1990); Renne et al. (1990)
Kalahari[d]					
Ritscherflya Supergroup (rotated to Kalahari)	1130 ± 12	61	29	4	Powell et al. (2001)
Umkondo Igneous Province	1105 ± 5	66	37	3	Powell et al. (2001); Wingate (2001)
Kalkpunt Formation,	∼1065?	57	3	7	Briden et al. (1979); Powell et al. (2001)
Central Namaqua	∼1030–1000	8	330	10	Onstott et al. (1986); Robb et al. (1999)
Siberia					
Ust-Kirba sediments and sills	∼980	−8	183	10	Pavlov et al. (2002)
Kandyk Formation	∼1000	−3	177	4	Pavlov et al. (2002); Rainbird et al. (1998)
Ignikan Formation	∼1010	−16	201	4	Pavlov et al. (2000)
Nelkan Formation	∼1020	−14	219	6	Pavlov et al. (2000)
Milkon Formation	∼1025	−6	196	4	Pavlov et al. (2000)
Kumahinsk Formation	∼1030	−14	201	7	Pavlov et al. (2000)
Malgina Formation	1043 ± 14	−22	226	7	Osipova in Smethurst et al. (1998); Ovchinnikova et al. (2001)
Malgina Formation	1043 ± 14	−25	231	3	Gallet et al. (2000); Ovchinnikova et al. (2001)
Chieress dyke	1384 ± 2	4	258	6.6	Ernst et al. (2000)
Kuonamka dykes	1503 ± 5	6	234	22	Ernst et al. (2000)
North China					
Dongjia Formation, Lushan	∼650	−60.8	97.4	6.7	From Zhang et al. (2006)
Mean poles	∼700	−42.9	107.0	5.7	From Zhang et al. (2006)
Nanfen Formation	800–780	−16.5	121.1	11.1	From Zhang et al. (2006)
Cuizhuang/Sanjiaotang Formations	950	−41.0	44.8	11.3	From Zhang et al. (2006)
Tieling Formation	1100	2.2	163.6	25.3	From Zhang et al. (2006)
Baicaoping Formation	1200	−43.0	143.8	11.1	Zhang et al. (2006)

Table 1 (*Continued*)

Rock unit	Age (Ma)	Pole		A_{95} (°)	Reference
		(°N)	(°E)		
Yunmengshan Formation	1260	−60.6	87.0	3.7	Zhang et al. (2006)
Yangzhuang/Wumishan Formations	1300	−17.2	45.0	5.5	From Zhang et al. (2006)
Yangzhuang Formation	∼1350	−17.3	3.5	5.7	Wu et al. (2005)
Taihang dyke swarm	1769.1 ± 2.5	−36.0	67.0	2.8	Halls et al. (2000)
Tarim					
Aksu dykes	807 ± 12	19	128	6	Chen et al. (2004)
Baiyixi Formation	∼740	17	194	4	Huang et al. (2005)
South China					
Doushantuo carbonates	584 ± 26	6	197	6	Macouin et al. (2004)
Liantuo Formation	748 ± 12	4	161	13	Evans et al. (2000)
Xiaofeng dykes	802 ± 10	14	91	11	Li et al. (2004)
Amazonia					
Nova Floresta	∼1200	25	165	6	Tohver et al. (2002)
Oaxaquia					
Oaxaca anorthosite	∼950	47	267	23	Ballard et al. (1989)

Abbreviations: SA, South Australia; NT, Northern Territory; WA, Western Australia.
[a] Recalculated by Hodych et al. (2004).
[b] Age based on APWP interpolation.
[c] Rotated to India 28° counterclockwise around the pole of 25.8°N, 330°E (Torsvik et al., 2001b).
[d] For ages see Powell et al. (2001) and references therein.

the previous section); (b) similarities in the Neoproterozoic stratigraphy of South China, southeastern Australia and western Laurentia as recognised earlier by Eisbacher (1985); (c) similarities between crustal provinces in the Cathaysia Block of southeastern South China and southern Laurentia; (d) the need for a western source region to provide the late Mesoproterozoic detrital grains in the Belt Basin. The modified reconstruction (Fig. 2b) and the schematic history for the formation and break-up of such a configuration were adapted for constructing the Geodynamic Map of Rodinia (Appendix I) and the related palaeogeographic time slices and animation. Although not a unique solution, it negates the need for matching basement geology between Australia–East Antarctica and western Laurentia prior to 1000–900 Ma (see discussions in Section 2.1.1), and is plausible in view of the currently known Neoproterozoic rift record, early Neoproterozoic mantle plume record, and palaeomagnetic constraints. Key evidence for the model is discussed below.

2.1.2.1. Cathaysia as an extension of Laurentia from ca. 1800 Ma until Rodinia time, and no connection between Australia–East Antarctica, Yangtze craton and Laurentia–Cathaysia until 1100–900 Ma. The Cathaysia Block, although poorly exposed, has a crustal composition similar to what may be expected for the western source region of the Belt Basin in southwestern Laurentia (Ross et al., 1992). It has a ca. 1830–1430 Ma basement that recorded 1300–1000 Ma metamorphism (Z.X. Li et al., 2002). Hainan Island in this block is particularly similar to the Mojave Province in that it has ca. 1430 Ma granitic intrusions, as well as synchronous intracratonic sedimentary and volcanic successions (Z.X. Li et al., 2002, and unpublished data), comparable to the 1500–1350 Ma transcontinental granite–rhyolite province of southern Laurentia (e.g., Nyman et al., 1994). The sedimentary provenance of quartzite in the overlying post-1200 Ma succession (Li et al., unpublished data), assumed to have sourced from older Cathaysian rocks, contains populations comparable to the 1610–1490 Ma, westerly sourced non-Laurentian detrital grains reported in the Belt Basin (Ross and Villeneuve, 2003). It is thus plausible to have Cathaysia as a part of Laurentia for at least the latter part of the 1830–1000 Ma interval (Figs. 2b and 3).

According to the "Missing-Link" model, the collision between Laurentia–Cathaysia and Yangtze did not start until ca. 1140 Ma or younger at one end of the Sibao Orogen (Greentree et al., 2006), and lasted until ca. 900 Ma at the other end in South China (Z.X. Li et al., 2003a; Ling et al., 2003). This late formation of the Laurentia–South China–Australia–East Antarctica connection provides an explanation for the occurrence of Grenvillian orogenic events along western Laurentia, northern Queensland, King Island-Transantarctic Mountains, and the Albany-Fraser-Musgrave belt.

2.1.2.2. Neoproterozoic plume record. Large features related to mantle plume events, such as radiating dyke swarms or large igneous provinces, can be used for reconstructing past continental connections (e.g., Ernst et al., 1995, 2008). Park et al. (1995) were the first to apply this technique to test the SWEAT connection, but precise dating of the relevant dyke swarms revealed that their ages are too different to represent the same event: ca. 825 Ma for the Gairdner-Amata dyke swarm in central and southeastern Australia (Sun and Sheraton, 1996; Wingate et al., 1998), and ca. 780 Ma for the radiating Gunbarrel dyke swarms in western Laurentia (e.g., Harlan et al., 2003a,b). No counterpart of the Australian 825 Ma event has been reported in western Laurentia as one would expect in the SWEAT configuration. The radiating 780 Ma Gunbarrel dyke swarm in western Laurentia

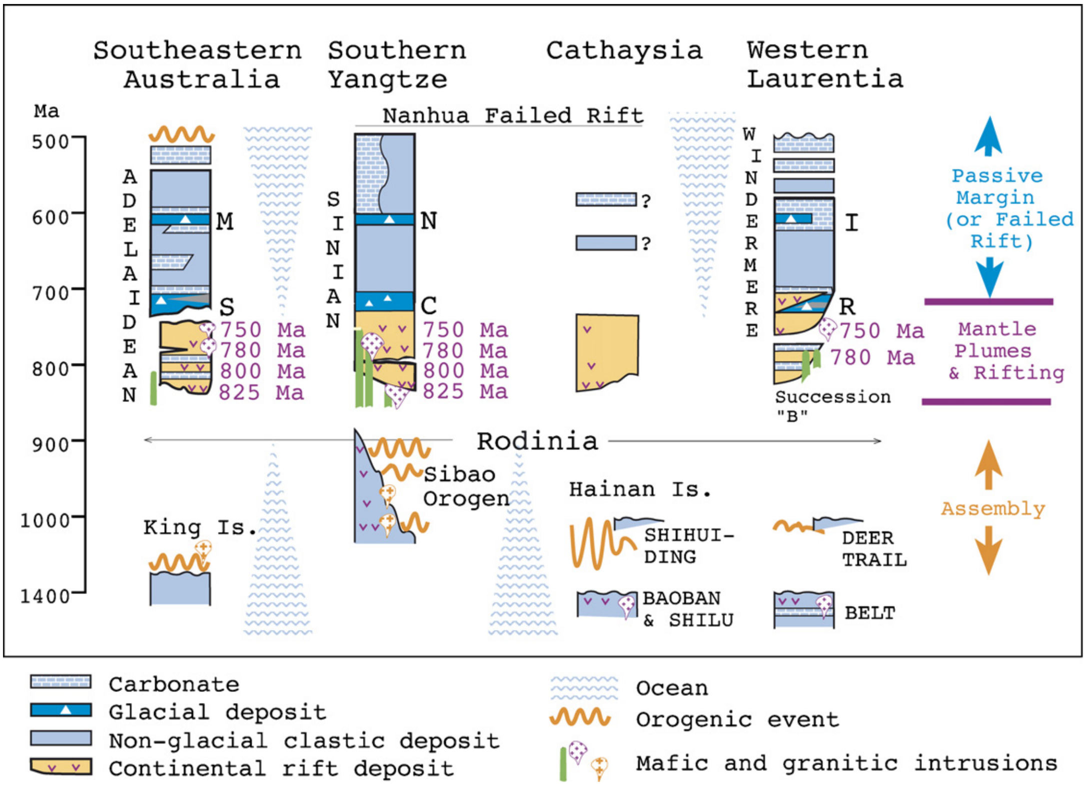

Fig. 3. Tectonostratigraphic correlations between Australia, Yangtze craton, Cathaysia and Laurentia. Major glaciations: M = Marinoan; S = Sturtian; N = Nantuo; C = Changan; I = Ice Brook; R = Rapitan.

points to a plume-centre to its west (present coordinates), but in eastern Australia there is no evidence for a plume centre apart from limited volcanism in the Adelaide Rift Complex (Preiss, 2000) and minor mafic igneous rocks in northwestern Tasmania (Holm et al., 2003).

Li et al. (1999) and Z.X. Li et al. (2003b) presented evidence for South China being above a plume centre at both ca. 825 Ma and 780 Ma, which is consistent with its position in the "Missing-Link" configuration (Fig. 4). Key evidence for a ca. 825 Ma plume-head beneath South China includes the widespread occurrence of magmatism with rock types ranging from granite to mafic–ultramafic dykes and sills (implying mafic underplating and a large heat source), large-scale syn-magmatic doming, and the development of ca. 820 Ma continental rift systems (Li et al., 1999; X.H. Li et al., 2003a,b; Ling et al., 2003; Wang and Li, 2003). A similar large-scale magmatic breakout occurred at ca. 780 Ma, with some mafic dykes showing geochemical characteristics of continental flood basalts (Z.X. Li et al., 2003b; Lin et al., 2007). It should be noted that alternative geochemical interpretations exist for the 825–740 Ma magmatism in South China (e.g., arc magmatism, Zhou et al., 2002; magmatism caused by post-orogenic slab break-off, Wang et al., 2006).

2.1.2.3. Neoproterozoic rift records and glacial events. Rifting records in South China show remarkable similarities to that of eastern Australia, featuring four major episodes of magmatism and rifting in the ca. 830–700 Ma interval: ca. 820 Ma, ca. 800 Ma, ca. 780 Ma, and ca. 750–720 Ma (Powell et al., 1994; Li et al., 1995, 1999; X.H. Li et al., 2002; Z.X. Li et al., 2003b; Preiss, 2000; Wang and Li, 2003) (Fig. 3). A rift–drift transition has been identified in the Adelaidean stratigraphy to be between the ca. 720 Ma Sturtian glacial deposits and the overlying Tapley Hill Formation (Powell et al., 1994).

There is no record of rifting along western Laurentia until after the 780 Ma Gunbarrel dyke swarm (e.g., Harlan et al., 2003a,b). Rift magmatism mainly occurred between 750 Ma and 720 Ma (e.g., Heaman et al., 1992; Ross and Villeneuve, 1997; Karlstrom et al., 2000), although evidence exists for events of younger ages (Lund et al., 2003; Fanning and Link, 2004) as in South China (e.g., Zhou et al., 2004; Zhang et al., 2005). Along with the glacial record, the post-780 Ma tectonostratigraphy of western Laurentia, eastern Australia (e.g., Young, 1992; Rainbird et al., 1996) and South China (e.g., Wang and Li, 2003) are widely regarded as correlative (Fig. 3), although such a correlation is not unique across the globe.

2.1.2.4. Palaeomagnetism. Another advantage of the "Missing-Link" model is that it does not require a common apparent polar wander path (APWP) between the relevant continents until amalgamation sometime between 1000 Ma and 900 Ma. On the other hand, the currently available palaeomagnetic data for ca. 820–800 Ma permit such an Australia–South China–Laurentia configuration at that time, although it must have broken up by ca. 750 Ma (Li et al., 2004; note the overlap between the 820 Ma and 800 Ma palaeopoles for India, Australia and South China, and the scatter in the ca. 750 Ma poles in Fig. 2b).

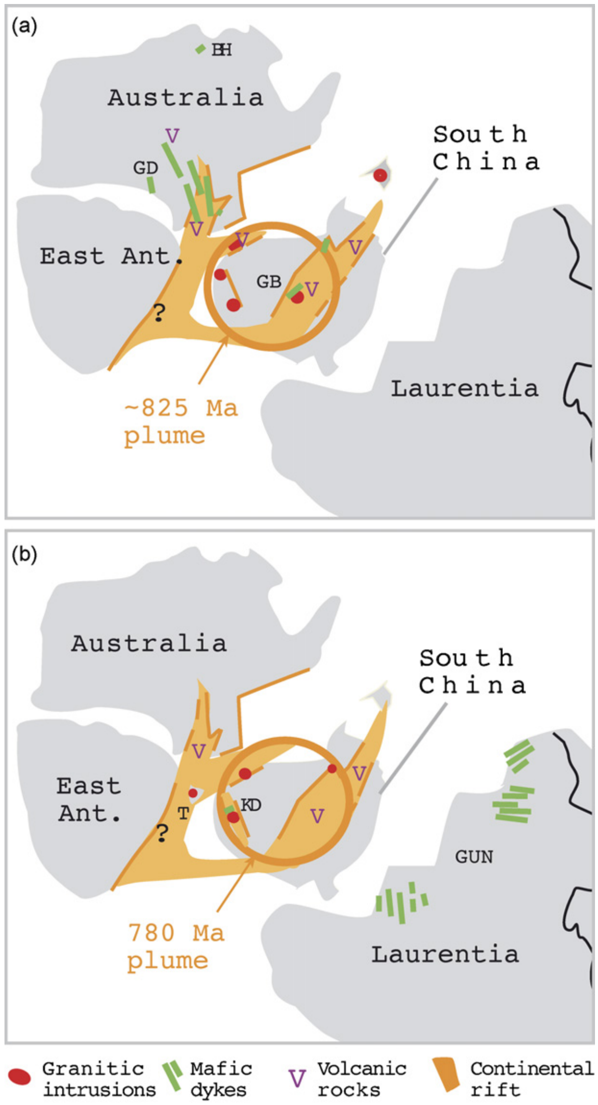

Fig. 4. Evidence for plume activities in the Missing-Link model: GD, Gairdner-Amata dyke swarm (Sun and Sheraton, 1996; Wingate et al., 1998); BH, Bow Hill lamprophyre dykes (Pidgeon et al., 1989); GB, Guibei mafic–ultramafic intrusions (Li et al., 1999); GUN, Gunbarrel dyke swarms (Harlan et al., 2003a,b); KD, Kangding dykes and mafic intrusions (Z.X. Li et al., 2003b; Lin et al., 2007); T, Mafic and granitic intrusions in Tasmania (Holm et al., 2003).

However, the configuration in Fig. 2b is not unique either geologically or palaeomagnetically. An alternative position for South China northeast of Australia satisfies the ca. 750 Ma poles (Evans et al., 2000), but contradicts the ca. 800 Ma poles (Li et al., 2004). Positions to the northwest or west of Australia, and adjacent to the northern margin of Greater India, have also been proposed (Evans et al., 2000; Jiang et al., 2003; Li et al., 2004; Yang et al., 2004), but the geological matches in these configurations are not as strong as in the “Missing-Link” configuration discussed above. Also, by removing South China from the “Missing-Link” configuration, one would have to find other explanations for the geological and palaeomagnetic mismatches between Australia–East Antarctica and Laurentia as discussed in Section 2.1.1.

2.1.3. The AUSWUS and AUSMEX connections

AUSWUS (Australia–Southwest US)-like fit (Fig. 2c) was first suggested by Brookfield (1993) based on matching linear fractures supposedly formed during the break-up of Rodinia along the margins of eastern Australian craton and western Laurentia. However, the lineaments along the eastern margin of the Australian craton were later shown to be no older than ca. 600 Ma (Direen and Crawford, 2003). The AUSWUS connection was later revived by Karlstrom et al. (1999) and Burrett and Berry (2000), mostly through matching basement provinces and sedimentary provenances between Australia and southwestern Laurentia (also see Blewett et al., 1998; Berry et al., 2001; Ross and Villeneuve, 2003; Wade et al., 2006). Despite its merits as stated by its proponents, there are noticeable difficulties in this model (Fig. 2c), namely:

(1) The truncation of the Albany-Fraser-Musgrave belt against cratonic western Laurentia, and difficulties in explaining the scattered evidence of Grenvillian metamorphism in western Laurentia, northern Queensland, Tasmania and the Ross Orogen (see Section 2.1.1).
(2) The lack of a prominent ca. 1400 Ma granite–rhyolite province in Australia as is present in southern Laurentia.
(3) The lack of a ca. 825 Ma plume record in Laurentia as a counterpart to the Gairdner-Amata dyke swarm in Australia (Wingate et al., 1998), and a lack of ca. 780 Ma plume record in northern Australia, where one would expect to find a plume-head for the 780 Ma Gunbarrel radiating dyke swarm in western Laurentia in such a configuration.
(4) The significantly younger starting age of continental rifting in western Laurentia (<750 Ma) compared with eastern Australia (ca. 825 Ma).
(5) The ca. 1200 Ma palaeomagnetic misfit (Pisarevsky et al., 2003a), although the ca. 1100–1050 Ma palaeopoles mostly agree rather well (Fig. 2c).

Based on a new 1070 Ma palaeomagnetic pole from the Bangemall Basin sills in the Edmund Fold Belt of Western Australia, Wingate et al. (2002) argued that previous poles of the same age from central Australia (poles marked as 1076 Ma and 1054 Ma in Fig. 2b and c) are unreliable. Using only their own pole, they argued that if Australia was connected to Laurentia at ca. 1070 Ma, the connection could only have been between northern Queensland of Australia and Mexico of southern Laurentia (the AUSMEX fit; Fig. 2d). However, subsequent work on the 1070 Ma rocks in central Australia reaffirmed the reliability of the central Australian poles (Schmidt et al., 2005). Therefore it appears probable that either the 1070 Ma Bangemall Basin sills, or the central Australian data, have suffered vertical-axis rotations, during either the ≥900 Ma (^{40}Ar–^{39}Ar cooling ages) Edmundian Orogeny in the southern Capricorn Orogen (Occhipinti and Reddy, 2005) in the case of the Bangemall sills, or the late Neoproterozoic Petermann or Phanerozoic Alice Springs orogenies in central Australia. The geological merit of the AUSMEX fit is yet to be demonstrated. Nonetheless, a new palaeomagnetic pole from the 800 Ma to 760 Ma oriented cores from the Lancer-1 drill hole in the Officer Basin of south Central Australia is more compatible with the ∼780 Ma Laurentian poles in the AUSMEX configuration than in any of the abovementioned configurations (Pisarevsky et al., 2007).

As mentioned earlier, ca. 1200 Ma poles do not permit either an AUSWUS or an AUSMEX fit at that time (Pisarevsky et al., 2003a; Fig. 2c and d).

2.1.4. Siberia–Laurentia connection

Although there is a consensus that Siberia was likely connected to Laurentia for much of the Proterozoic, opinions differ regarding how they were connected. The dominant view is that Siberia was adjacent to northern Laurentia, but different models show its position in a range of orientations (Hoffman, 1991; Condie and Rosen, 1994; Pelechaty, 1996; Frost et al., 1998; Rainbird et al., 1998; Ernst et al., 2000; Vernikovsky and Vernikovskaya, 2001; Metelkin et al., 2005a; see review by Pisarevsky et al., 2003b). A vastly different position for Siberia was proposed by Sears and Price (1978, 2000) with Siberia placed west of the present western margin of Laurentia. However, this latter fit is not supported by palaeomagnetic data (Fig. 2e) which show that Siberian poles fall far away from coeval palaeopoles for Laurentia in such a configuration. The configuration we adopted for the Rodinia Map follows that of Pisarevsky et al. (2008). It is largely based on existing palaeomagnetic constraints for the ∼1050–970 Ma interval between Siberia and Laurentia (Fig. 2f, which shows that Siberian poles generally follow the APWP of Laurentia for the 1043–980 Ma interval). Detailed discussions for the various models are given in Pisarevsky et al. (2008).

Palaeomagnetic data place Siberia at some distance from the northern Laurentian margin, allowing space for other Precambrian blocks (e.g., Arctida of Zonenshain et al., 1990; Vernikovsky, 1997; Vernikovsky et al., 2003). The relative distance between Siberia and Laurentia explains the lack of any counterpart for the Mackenzie large igneous event in Siberia. Sklyarov et al. (2003), Metelkin et al. (2005c) and Gladkochub et al. (2006) reported mafic intrusions in southern Siberia with ^{40}Ar–^{39}Ar ages around ∼740 Ma that may be correlated with the Franklin igneous event in northern Laurentia associated with the Neoproterozoic opening of the Palaeo-Asian Ocean. This event broadly coincides with the accumulation of the Karagas Group in south Siberia, which has been interpreted as a passive margin succession (Pisarevsky and Natapov, 2003; Vernikovsky et al., 2003; Gladkochub et al., 2006). This earlier break-up model contrasts with the Early Cambrian break-up suggested by Pelechaty (1996) based on tectonostratigraphic analyses, with the latter model being challenged by both Rainbird and de Freitas (1997) and Khudoley (1997).

2.2. Continents along the present eastern margin of Laurentia: Baltica, Amazonia and West Africa cratons

The relative position of Baltica to Laurentia in Rodinia is probably one of the least controversial. Although geological correlations suggest that the two blocks could have been together as long ago as ca. 1800 Ma (e.g., Gorbatschev and Bogdanova, 1993; Karlstrom et al., 2001; see also Zhao et al., 2002 for a review) or even before (e.g., Heaman, 1997; Bleeker and Ernst, 2006), palaeomagnetic data suggest a rather complex history between the two cratons during the Mesoproterozoic (e.g., Elming et al., 1993, 2001; Buchan et al., 2000, 2001; Pesonen et al., 2003). The configuration we adapted in the Rodinia Map (Appendix I; Fig. 5) follows that of Hoffman (1991; see also Bogdanova et al., 2008; Pease et al., 2008), which is a geologically based reconstruction but is also supported by a common APWP between the two cratons for the ∼1000–900 Ma interval (Elming et al., 1993; Hyodo et al., 1993; Weil et al., 1998; Pisarevsky et al., 2003b; Fig. 5). It is noted that, although pre-1000 Ma palaeomagnetic data do not permit exactly the same configuration, they still allow the two adjacent cratons to develop correlative Proterozoic belts including the Sveconorwegian and Grenville orogens (Winchester, 1988; Gower et al., 1990; Gorbatschev and Bogdanova, 1993), and similar ca. 1500–1350 Ma intracratonic magmatism (Åhäll and Connely, 1998; Karlstrom et al., 2001; Bogdanova et al., 2001, 2008; Čečys et al., 2002; Söderlund et al., 2002, 2005; Čečys and Benn, 2007).

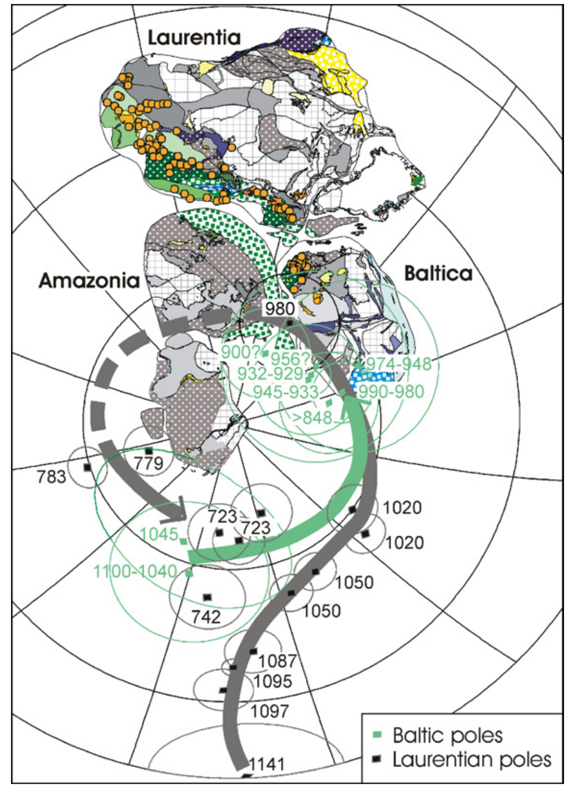

Fig. 5. Positions of Baltica, Amazonia and West Africa in Rodinia, with palaeomagnetic South poles showing data from Baltica (green poles) merge with that of Laurentia (black poles) by ca. 1000 Ma. Orange dots represent ca. 1500–1350 Ma intracratonic magmatism.

The position of the Amazonia craton roughly follows that of Hoffman (1991, a geologically based reconstruction) and Weil et al. (1998), (a palaeomagnetic reconstruction). Amazonia was one of the cratons that was separated from Laurentia by a Mesoproterozoic ocean, the closure of which made it part of Rodinia during the Grenvillian Orogeny (e.g., Davidson, 1995, 2008; Rivers, 1997; Loewy et al., 2003). According to this interpretation, the Rondonia-Sunsas belt in southwestern Amazonia resulted from continental collision with Laurentia between 1080 Ma and 970 Ma (e.g., Sadowski and Bettencourt, 1996; Tassinari et al., 2000 and references therein). A recent ∼1200 Ma Amazonia palaeomagnetic pole reported by Tohver et al. (2002) led these authors to suggest a possible juxtaposition of western Amazonia with the Llano segment of Laurentia's Grenville orogen. This is broadly in accord with the proposed history of the Llano orogen (Mosher, 1998). Tohver et al. (2004, 2005) more recently suggested that the initial docking of Amazonia with southern Laurentia was followed by strike–slip transport of Amazonia to the northeast (present coordinates). However, such palaeomagnetic reconstructions, utilizing a single pole from each continent rather than matching relatively long segments of APWPs, have inherent longitudinal uncertainties.

Continental rifting along the eastern and southern margin of Laurentia may have started as early as ca. 750 Ma (Su et al., 1994; Aleinikoff et al., 1995; Fetter and Goldberg, 1995), although a later pulse of magmatism at ∼615–570 Ma is commonly interpreted as representing the break-up and opening of the Iapetus Ocean, first between Laurentia and Baltica, and then between Laurentia and Amazonia (Cawood et al., 2001; Cawood and Pisarevsky, 2006 and references therein).

There are few constraints on the position of the West Africa craton in Rodinia. No orogen of Grenvillian age has been found in or around the craton; therefore, if it was part of Rodinia, it would likely have been part of a larger craton (e.g., together with Amazonia, similar to their connection in Gondwanaland). No palaeomagnetic data are available to verify this position. Trompette (1994, 1997) proposed the existence of a single West Africa–Amazonia–Rio de La Plata mega-craton in the Meso- and Neoproterozoic, although their relative positions cannot be defined precisely. Onstott and Hargraves (1981), on the basis of a comparison of Paleo- to Mesoproterozoic palaeomagnetic data from these two blocks, also suggested that the two cratons were together during the Proterozoic but that large shear movements occurred between them. We have used an Amazonia–West Africa fit similar to their Gondwana fit.

2.3. Congo-São Francisco, Rio de la Plata and Kalahari cratons

The positions of Congo-São Francisco and Kalahari cratons relative to Laurentia in Rodinia (Fig. 6) broadly follow that proposed by Hoffman (1991) on geological grounds, but with some additional justifications according to more recent palaeomagnetic results. The Rio de la Plata terrane is placed between the

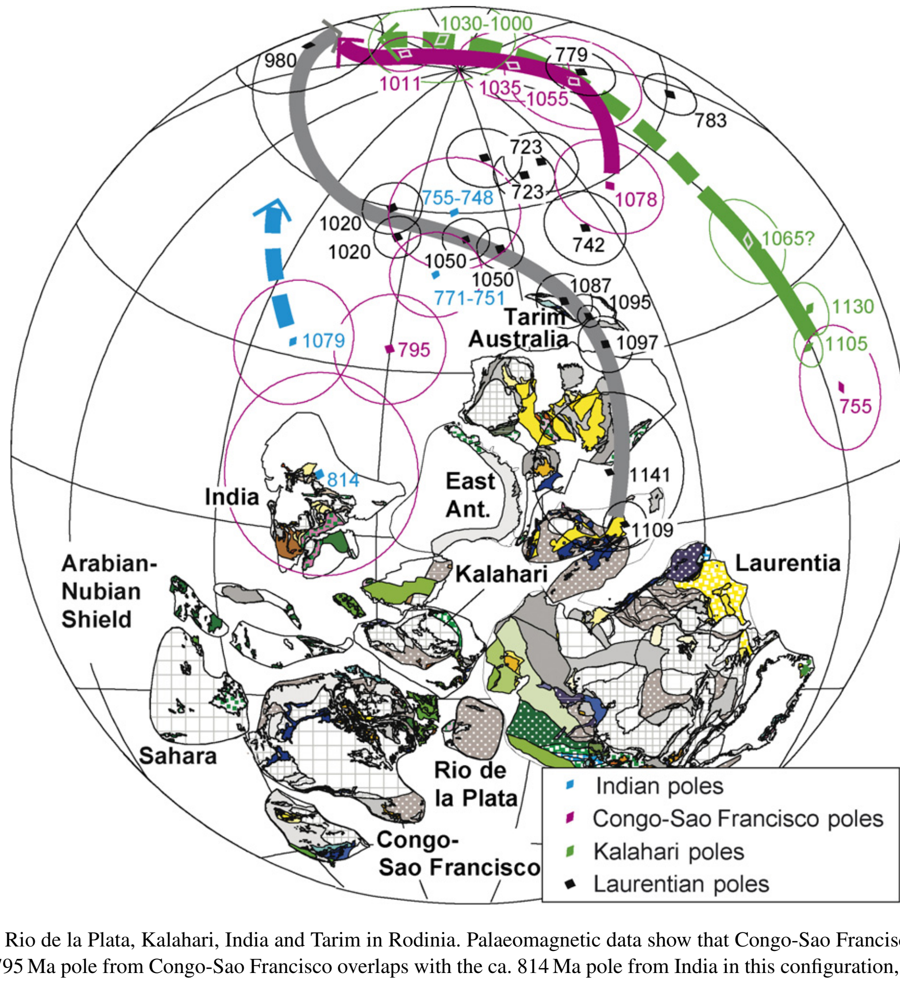

Fig. 6. Positions of Congo, Rio de la Plata, Kalahari, India and Tarim in Rodinia. Palaeomagnetic data show that Congo-Sao Francisco did not join Laurentia until after ca. 1010 Ma. The ca. 795 Ma pole from Congo-Sao Francisco overlaps with the ca. 814 Ma pole from India in this configuration, but their ca. 755 Ma poles are widely separated (as the other ca. 755 Ma poles in Fig. 2b), suggesting this configuration was already disintegrating by that time.

Congo-São Francisco craton and Laurentia in Rodinia (e.g., Weil et al., 1998; D'Agrella-Filho et al., 2004).

The preservation of late Mesoproterozoic reworked and juvenile crust and orogenic events along the Irumide belt of the Congo-São Francisco craton (e.g., De Waele et al., 2003) and along the margin of the Rio de la Plata terrane (Fuck et al., 2008) permits them to be the continents that joined Laurentia during the Grenvillian Orogeny, although there are opinions that the Congo-São Francisco craton may not have been part of Rodinia at all (e.g., Kröner and Cordani, 2003; Pisarevsky et al., 2003b). Palaeomagnetic data indicate the presence of a large latitudinal gap (an ocean?) between the Congo-São Francisco craton and Laurentia at ca. 1100–1050 Ma (D'Agrella-Filho et al., 2004); the $\geq$1000 Ma orogenic events in the Irumide belt may indicate closure of this ocean by ca. 1000 Ma (e.g., De Waele et al., 2003, 2006a,b, 2008; De Waele, 2005; Johnson et al., 2006, 2007a).

The Kalahari craton, together with the Grunehogna terrane (e.g., Groenewald et al., 1995; Jones et al., 2003) and perhaps part of the Maud Province of East Antarctica (e.g., Fitzsimons, 2000), are placed against the Grenville belt in southern Laurentia following Hoffman (1991) and Hanson et al. (1998). Palaeomagnetic data from Kalahari and Laurentia at ca. 1110 Ma suggest that there was a latitudinal difference of ca. $30 \pm 14^\circ$ between their facing margins at that time (Powell et al., 2001; Hanson et al., 2004), but by ca. 1000 Ma the APWPs of the two continents merged together in this configuration (Fig. 6). This is consistent with metamorphic ages of as young as ca. 1000 Ma along both the Namaqua-Natal belt in southern Kalahari (e.g., Eglington and Armstrong, 2003) and the Grenville belt in Laurentia (e.g., Davidson, 1995; Rivers, 1997; Keppie et al., 2003; Weber and Hecht, 2003). However, other possible positions of Kalahari in Rodinia have also been proposed on geological and palaeomagnetic grounds (e.g., Dalziel, 1997; Pisarevsky et al., 2003b).

Like the Amazonia craton, the Congo-São Francisco, Rio de la Plata and the Kalahari cratons are believed to have broken away from Laurentia between 750 Ma (see evidence below) and the final assembling of Gondwanaland by ca. 550–520 Ma (e.g., Meert, 2003; Collins and Pisarevsky, 2005). Rift magmatism at ca. 750 Ma is present along the Grenvillian margin of Laurentia (e.g., Su et al., 1994; Fetter and Goldberg, 1995; Aleinikoff et al., 1995), the western margin of Kalahari (e.g., Frimmel et al., 1996, 2001; Hoffman et al., 1996), and the Damaran margin of the Congo-São Francisco craton (e.g., Hoffmann et al., 2004; Halverson et al., 2005). The stratigraphy of the Kalahari and Congo-São Francisco continental margins indicates rift–drift transitions immediately after ca. 750 Ma (Frimmel et al., 1996; Hoffman et al., 1996; Hoffmann et al., 2004; Halverson et al., 2005; Johnson et al., 2007b). As shown in Fig. 6, ca. 750 Ma palaeomagnetic poles from the Congo-São Francisco craton fall $>30^\circ$ away from coeval poles of Laurentia.

The position of the Congo-São Francisco and Kalahari cratons in Rodinia as shown in Fig. 6 is also consistent with the common occurrence of ca. 830–745 Ma plume-induced intraplate magmatism in these two cratons as well as coeval magmatism in adjacent India, Australia, South China and Tarim cratons (see Section 3.2). However, no pre-750 Ma Neoproterozoic magmatism has yet been identified in southern Laurentia.

2.4. India, Madagascar, Seychelles, and the Arabian–Nubian Shield

Geological data demonstrate that northeastern Madagascar (the Antongil block) was part of the Indian craton before and during Rodinia time (e.g., Collins and Windley, 2002; Collins et al., 2003; Collins, 2006)—here we collectively call them the Indian craton. The Seychelles appears to have formed along the west margin of India during the Neoproterozoic (Torsvik et al., 2001a,b).

The position of the Indian craton in Rodinia is controversial. Earlier workers attached it to Australia and East Antarctica as in the Gondwanan configuration (e.g., Dalziel, 1991; Hoffman, 1991; Moores, 1991; Li et al., 1996; Torsvik et al., 1996; Weil et al., 1998). However, subsequent geological (Fitzsimons, 2000) and palaeomagnetic (see below) investigations have challenged such a configuration. Based on their palaeomagnetic results from the 771 Ma to 751 Ma Malani igneous suite of northwestern India, Torsvik et al. (2001a,b) suggested that India was either not part of Rodinia, or was adjacent to the northwestern margin of Australia at Rodinia time and underwent a left-lateral movement relative to Australia–East Antarctica before 535 Ma to reach its Gondwanan position. Powell and Pisarevsky (2002) supported the idea that India was probably never part of Rodinia because a ca. 810 Ma pole (Radhakrishna and Mathew, 1996) places India at a polar position when Rodinia was supposed to occupy a largely low palaeolatitude position.

In the Rodinia Map (Appendix I), the position of India between 900 Ma and 800 Ma is made following both geological evidence (e.g., plume record as described by Z.X. Li et al., 2003b) and palaeomagnetic interpretations as described by Li et al. (2004). Both India and Seychelles have Neoproterozoic bimodal magmatism similar to that in Australia and South China in age distributions. Many interpreted this to represent continental arc magmatism, partly because of the perceived "continental margin" positions of these continental blocks (e.g., Torsvik et al., 2001a,b; Tucker et al., 2001; Ashwal et al., 2002). However, their bimodal nature, the intraplate characteristics of the Malani igneous suite (e.g., Roy, 2001; Singh and Vallinayagam, 2004) and other Neoproterozoic intrusions elsewhere in India (e.g., Santosh et al., 1989), and their similar age distributions to plume-related rocks in Australia and South China, make them more likely the products of melting above Neoproterozoic plume-heads during Rodinia time (e.g., Frimmel et al., 2001; Li et al., 2001; Z.X. Li et al., 2003b).

Palaeomagnetically, a poorly dated ca. 1079 Ma pole from India (Miller and Hargraves, 1994) does not agree with coeval poles from Australia in the preferred Rodinia configuration, whereas the ca. 810 Ma pole from India (Radhakrishna and Mathew, 1996) agrees with similar-aged poles from South China (Li et al., 2004), South Australia (McWilliams and McElhinny, 1980; however, the reliability of this pole has not been demonstrated with confidence) (Fig. 2b), and the Congo-São Francisco craton (Meert et al., 1995; Deblond et al., 2001) (Fig. 6).

We thus envisage that India became part of Rodinia by ca. 900 Ma through continental collision along the ca. 990–900 Ma high-grade metamorphic Eastern Ghats belt of India and the corresponding Rayner Province in East Antarctica (e.g., Mezger and Cosca, 1999; Boger et al., 2000, 2001; Fitzsimons, 2000; Kelly et al., 2002; Fig. 6).

The ca. 770–750 Ma palaeopoles from the India craton are far removed from the ca. 755 Ma pole of the Mundine Well dykes in Australia (Wingate and Giddings, 2000) in this configuration (Fig. 2b), indicating that if the configuration is correct, India would have rifted away from Australia–East Antarctica by ca. 755 Ma.

There are few constraints on the positions of the numerous terranes constituting the East African Orogen (Stern, 1994), which includes the Arabian–Nubian Shield and the ∼3000 km long Azania continent that stretches from Arabia to central Madagascar (Collins and Windley, 2002; Collins and Pisarevsky, 2005; Collins, 2006). Cox et al. (2004) and Fitzsimons and Hulscher (2005) suggest that central Madagascar was part of the Congo-São Francisco craton at ca. 1800 Ma but became an independent terrane during Rodinia time and joined India by ca. 700 Ma. Most of the terranes in the Arabian–Nubian Shield are believed to represent juvenile Neoproterozoic arcs and micro-continental fragments that accreted to western Gondwana during the Neoproterozoic (e.g., Stern, 1994; Whitehouse et al., 2001; Meert, 2003; Collins and Pisarevsky, 2005). However, the nature of widespread 850–750 Ma igneous rocks in these terranes remains controversial. Some believe they represent arc volcanism (e.g., Stern, 1994; Handke et al., 1999; Tucker et al., 2001). Others suggest that those bimodal rocks were likely formed (1) during continental extension (e.g., Kröner et al., 2000; Loizenbauer et al., 2001) with possible underplating of plume magmatism, (2) with contributions of oceanic plateaux (Stein and Goldstein, 1996), or (3) as arcs with a plume input (Teklay et al., 2002). The Neoproterozoic Arabian–Nubian Shield is thought to represent one of the most rapid episodes of predominantly juvenile crustal generation in Earth's history (Stein and Goldstein, 1996; Stein, 2003). As the age spectrum of these rocks is almost identical to those in South China, Australia and India, there could indeed have been a significant plume input as part of the Neoproterozoic Rodinia superplume activity (Li et al., 2003; Hargrove et al., 2004; see further discussions in Section 3.2).

In previous Rodinia reconstructions the Arabian–Nubian Shield was either placed adjacent to northeast Madagascar and India (e.g., Li and Powell, 2001), or adjacent, and outboard of the Congo-São Francisco and Saharan cratons (Collins and Pisarevsky, 2005). In the Rodinia Map (Appendix I and Figs. 6 and 8), these terranes are placed between India and Sahara, satisfying both the common plume records in Australia, India, the Arabian Shield, and the African cratons, and the records of arc accretion during the late Neoproterozoic assembly of Gondwanaland.

2.5. Tarim

We have positioned the Tarim craton of northwest China adjacent to northwestern Australia (Fig. 6) following the suggestion of Li et al. (1996) that was based on tectonostratigraphic correlations, with some palaeomagnetic justifications. Minor terranes (such as the Cimmerian terranes, Metcalfe, 1996) may have existed between Tarim and Australia (Li and Powell, 2001). The geological arguments for such a configuration are: (1) the presence of late Mesoproterozoic-earliest Neoproterozoic active margins along both the present northwestern and southern margin of this Archaean–Paleoproterozoic craton (Zhang et al., 2003; Lu et al., 2008a) suggests that it did not join Australia until the beginning of the Neoproterozoic; (2) 820–750 Ma bimodal (plume-induced?) intrusions in Tarim (Chen et al., 2004; Guo et al., 2005; C.L. Zhang et al., 2006; Lu et al., 2008b) can be correlated with the ca. 820–800 Ma lamprophyre dykes and kimberlite pipes in the Kimberley craton of western Australia (Pidgeon et al., 1989), and the 755 Ma Mundine Well dyke swarm in the northwestern Pilbara craton (Wingate and Giddings, 2000) and A-type granitic magmatism in the Leeuwin Block of southwestern Australia (Collins, 2003). A noticeable difference is the lack of ≤750 Ma rift magmatism in western Australia, which is present in Tarim (Xu et al., 2005; Z.X. Li et al., unpublished data). This could be due to the rift margin being farther off shore in Australia, or to it being totally reworked by the Pinjarra Orogeny (Fitzsimons, 2003); (3) both the Kimberley region of northwestern Australia and Tarim had up to three Neoproterozoic glacial intervals, although precise ages of the glacial events and their lateral correlations are still debated (Li et al., 1996; Grey and Corkeron, 1998; Xiao et al., 2004); (4) Cambrian mafic volcanic units in northeastern Tarim may correlate with the Antrim Plateau Volcanics in the Kimberley (Li et al., 1996), which have been interpreted as part of the 513 Ma Kalkarindji Large Igneous Province in central and northern Australia (Hanley and Wingate, 2000; Glass and Phillips, 2006).

Two Neoproterozoic palaeomagnetic poles have recently been reported from the Tarim craton: one from ca. 800 Ma mafic dykes (Chen et al., 2004), and the other from a ca. 750–730 Ma volcanic unit (Huang et al., 2005). The ca. 800 Ma pole puts Tarim at a palaeolatitude comparable to the position as in the Rodinia Map, although the relative orientation is somewhat different. The younger pole places Tarim at a significantly lower palaeolatitude than that adopted in the Rodinia Map. As no fold test has been reported for either of the two poles, and both study regions are inside younger fold-and-thrust belts, we cannot rule out vertical-axis rotation which can only be identified through comparative study of similar-age rocks in other parts of Tarim.

2.6. North China

Based mainly on tectonostratigraphic correlations, Li et al. (1996) suggested that the North China craton was adjacent to Siberia in Rodinia. The connection may have started as early as ca. 1800 Ma, and did not break-up until after ca. 600 Ma. Reliable palaeomagnetic data at both ca. 1770 Ma (Halls et al., 2000) and ca. 1350 Ma (Wu et al., 2005) from North China lend support to such a configuration (Fig. 7b). However, data from younger Proterozoic rocks suggest a slightly different configuration, involving a ca. 90° vertical-axis rotation of the North China

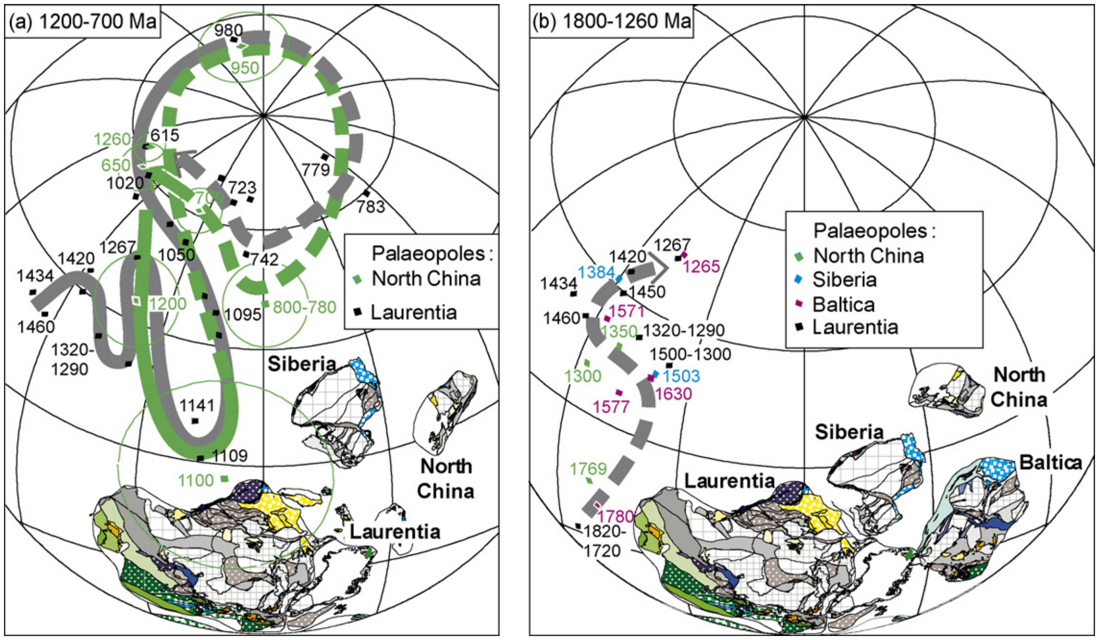

Fig. 7. Position of the North China craton (a) in Rodinia (after Zhang et al., 2006), and (b) during Paleo- to Mesoproterozoic (after Wu et al., 2005). Note that palaeomagnetic data in (b) suggest that Laurentia, Siberia, Baltica and North China could have been together for the ca. 1800–1300 Ma interval (Wu et al., 2005).

craton in relation to Laurentia between ca. 1350 Ma and 1200 Ma (S. Zhang et al., 2006) (Fig. 7a). According to S. Zhang et al. (2006), North China and Laurentia could have shared a common path between ca. 1200 Ma and ca. 700 Ma, but split up by ca. 615 Ma. It is unclear whether it was Siberia or other continental blocks that filled the gap between the two cratons for either the 1800–1350 Ma or the 1200–700 Ma intervals. We have adopted the rotation parameters for the North China craton provided by S. Zhang et al. (2006) for the Rodinia Map (Figs. 7a and 8). The lack of Neoproterozoic plume record in North China (e.g., Lu et al., 2008b) is consistent with it being at distances from continents like Australia and South China in Rodinia.

2.7. Minor terranes

The fragmented nature of many minor continental terranes, as well as their usual lack of any palaeomagnetic constraint, made it more difficult to place these terranes in Rodinia with much confidence. However, it is important to consider these terranes because they could account for spaces and geological links between larger cratons in Rodinia.

Keppie and Ramos (1999) proposed that two Central American terranes – Oaxaquia (Mexico) and Chortis (Honduras and Guatemala) – were situated along the northern boundary of South America in their reconstruction for the Ediacaran (Vendian)–Cambrian boundary. Keppie and Ortega-Gutierrez (1999) suggested that these blocks originated as arcs in a Grenvillian ocean between Laurentia, Baltica, and Amazonia, and were caught in-between the colliding cratons. These blocks experienced high grade, collisional-style metamorphism during their terminal collisions among Amazonia, Laurentia, and Baltica by ca. 1000 Ma. We place the Oaxaquia and Chortis blocks along the northern margin of Amazonia, within the zone of its collision with Baltica at ca.1000 Ma. Our model is also constrained by palaeomagnetic data from Oaxaquia (Ballard et al., 1989).

The space between Siberia and Laurentia in Rodinia is filled with several continental blocks that rifted away from northern Laurentia during the Cretaceous opening of the Canadian basin. These include terranes in northern Alaska, northern Chukchi Peninsula, Wrangel Island, the New Siberian Islands, Severnaya Zemlya, northern Taimyr and the Chukchi Plateau (e.g., Zonenshain et al., 1990; Vernikovsky, 1997; Bogdanov et al., 1998; Embry, 1998; Nokleberg et al., 2000; Vernikovsky and Vernikovskaya, 2001; Natal'in, 2004; Drachev, 2004). Our poor knowledge of these terranes makes geological correlation between northern Laurentia and Siberia more difficult. The position of the Kara Plate (exposed on Severnaya Zemlya in northern Taimyr) is debatable (Metelkin et al., 2005b). Although detrital zircon grains from the base of the Kara stratigraphy indicate a Baltican affinity (Pease et al., 2005, 2006), there are some geological arguments for a north Laurentian affinity (Zonenshain et al., 1990; Natal'in et al., 1999). Exact positions of these minor blocks in Rodinia are unclear due to the lack of palaeomagnetic data.

Previous reconstructions placed the Barentsia plate (exposed in Eastern Svalbard) against the southern part of east Greenland using their similarities in Neoproterozoic sedimentary successions (e.g., Winchester, 1988; Fairchild and Hambrey, 1995; Harland, 1997; Andresen, 2004). However, we choose to place it close to the northern part of East Greenland after Gee and Teben'kov (2004) who correlated the geochronological, structural, and stratigraphical data of Svalbard to northern Greenland. We have also reduced the size of the Barentsia plate in accordance with the recent discovery of a Caledonian arm in the Barents Sea (Breivik et al.,2002).

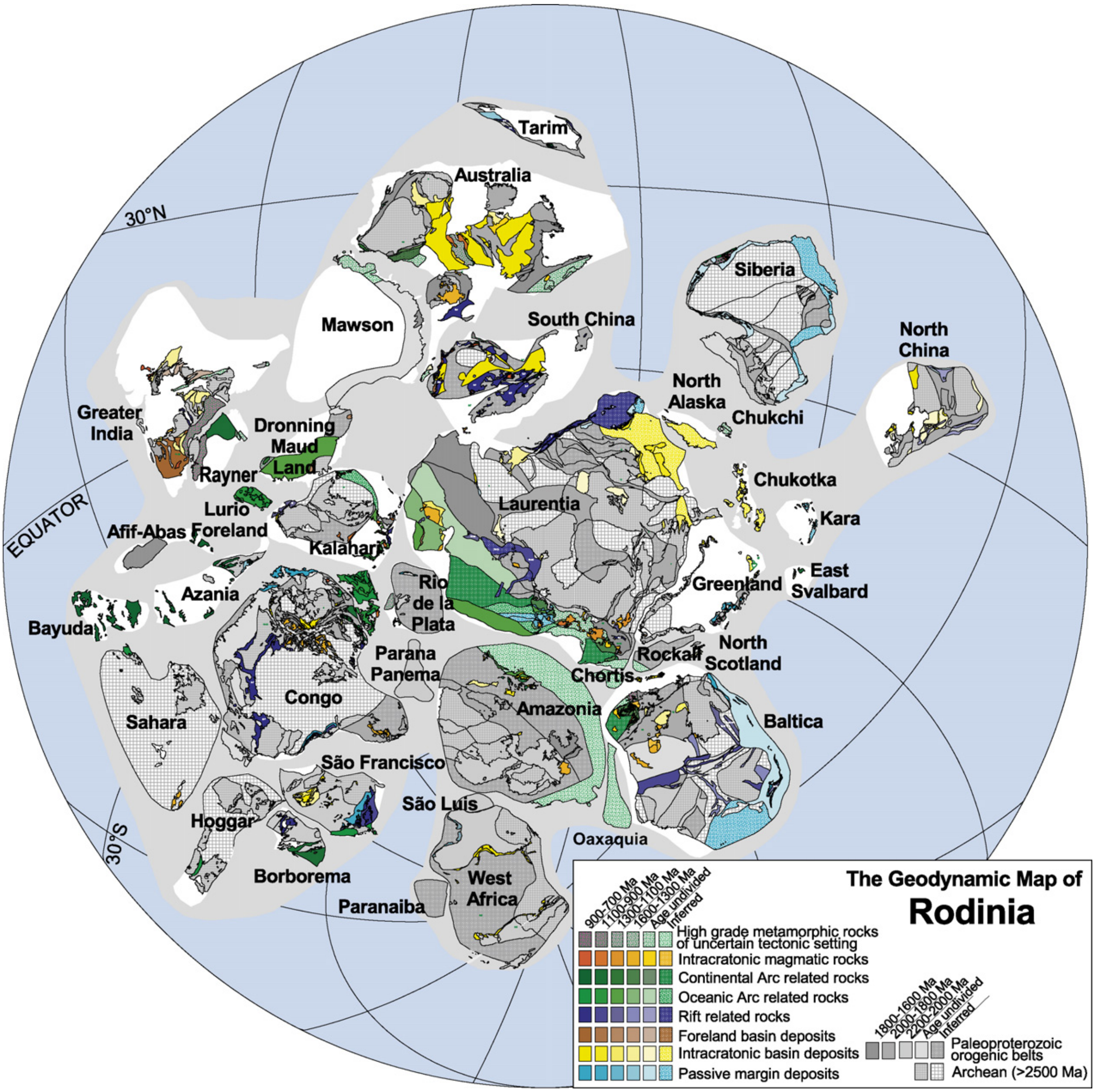

Fig. 8. A simplified (and reduced) Rodinia Map with legends.

3. Synthesis and animation: assembly and break-up of Rodinia, and formation of Gondwanaland

Through a series of cartoons, we illustrate a feasible scenario for the formation of Rodinia as shown in Fig. 8 and Appendix I, its break-up, and the eventual formation of Gondwanaland by the Early Cambrian. The rotation parameters for the major continental blocks are given in Appendix III, available online. A digital animation of the palaeogeographic evolution between 1100 Ma and 530 Ma, with geological features shown for selected time windows, is also given (Appendix II). We recognise that this is just one of a number of feasible scenarios for the evolution of Rodinia, and there are time intervals in the animation (e.g., the ca. 1000–820 Ma interval) for which we have very little palaeomagnetic constraints. Such reconstructions emphasise potential geodynamic linkages and provide a testable hypothesis predicting continental positions and plate interactions, and enable palaeoclimatic modellers to simulate continent-ocean-atmosphere interactions.

3.1. The formation of Rodinia (ca. 1100–900 Ma)

At 1100 Ma (Fig. 9a), Laurentia, Siberia, North China, Cathaysia (part of present day South China) and perhaps Rio de la Plata were already together, and the Yangtze craton had begun its oblique collision with Laurentia (at southern Cathaysia; Greentree et al., 2006). However, all other continental blocks were still separated from Laurentia by oceans. The Australian craton, including the East Antarctica part of the Mawson craton, had amalgamated by then. King Island, where a 1287 ± 18 Ma metamorphic age was recently reported (Berry et al., 2005), Tasmania, and the South Tasman Rise (1119 ± 9 Ma quartz syenite, Fioretti et al., 2005), could have been close to where Yangtze and Laurentia were colliding, thus receiving Laurentia-sourced sediments (Berry et al., 2001).

By ca. 1050 Ma (Fig. 9b), Kalahari had probably collided with southern Laurentia (see Section 2.3). Continued collision of the Yangtze craton with western Laurentia may have caused the 1090–1030 Ma metamor-

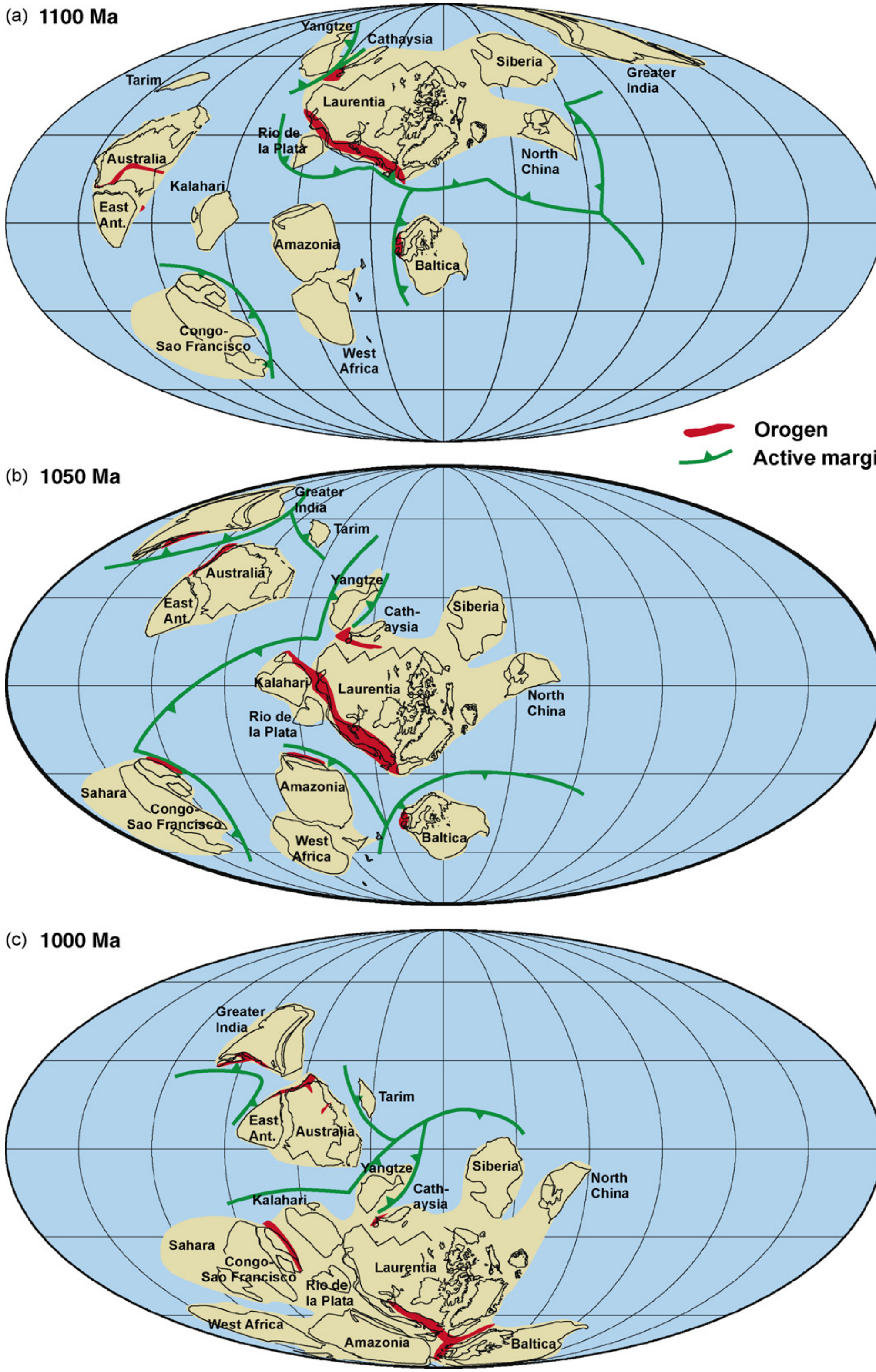

Fig. 9. Cartoons showing the assembly and break-up of Rodinia, and the formation of Gondwanaland. (a) 1100 Ma; (b) 1050 Ma; (c) 1000 Ma; (d) 900 Ma; (e) 825 Ma; (f) 780 Ma; (g) 750 Ma; (h) 720 Ma; (i) 630 Ma; (j) 600 Ma; (k) 550 Ma; (l) 530 Ma.

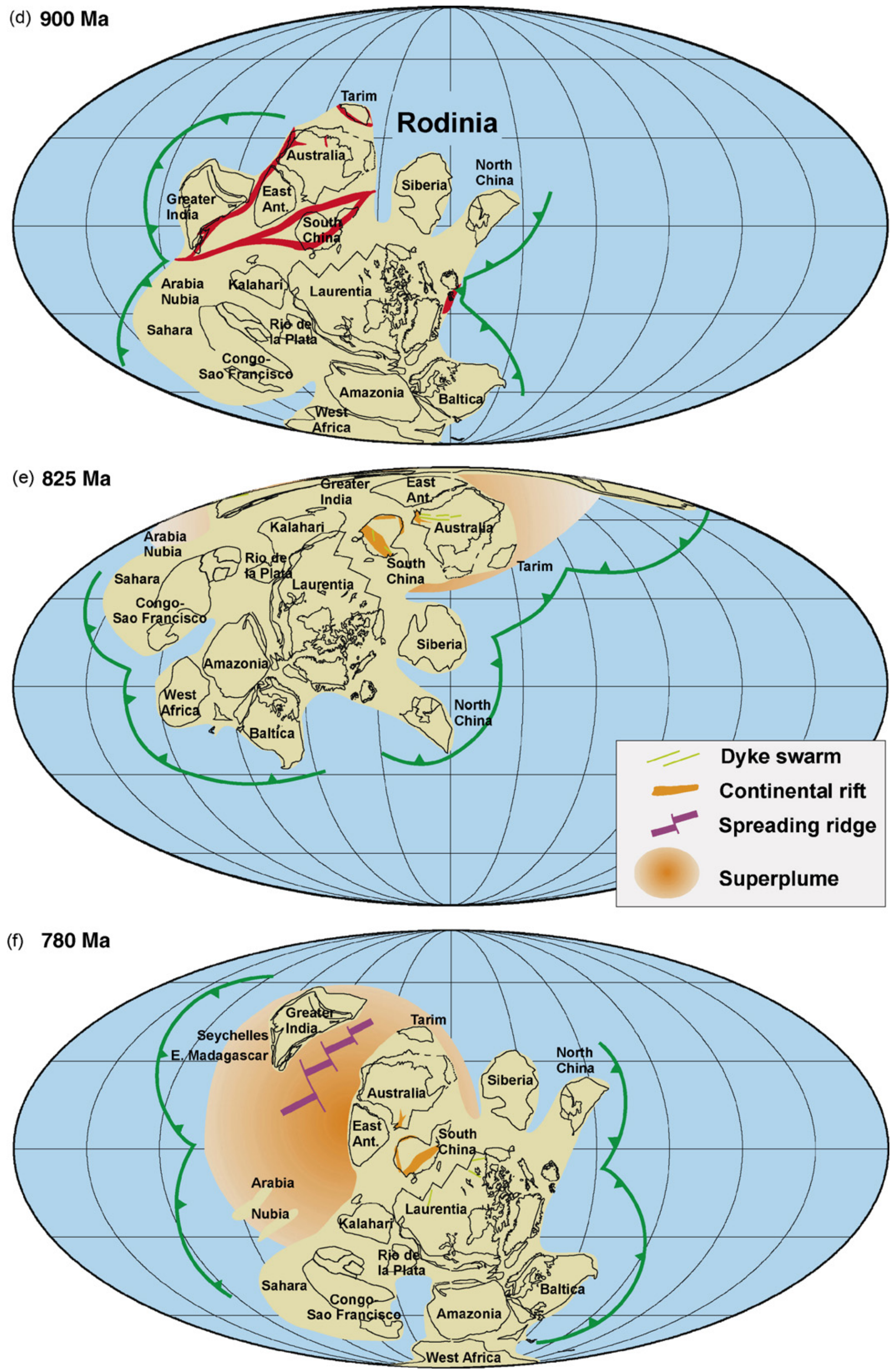

Fig. 9. (*Continued*)

phism of the mafic sills in the Belt-Purcell Supergroup (Anderson and Davis, 1995). Convergent margins were developed between most continents, as the oceanic lithosphere between them was consumed during the assembly of Rodinia.

At ca. 1000 Ma (Fig. 9c), all but India, Australia–East Antarctica and Tarim had assembled to be joined with Laurentia, whereas the Yangtze craton was still suturing to Cathaysia (part of Laurentia). The transpressional movement between Greater India and Western Australia may explain the 1100–1000 Ma

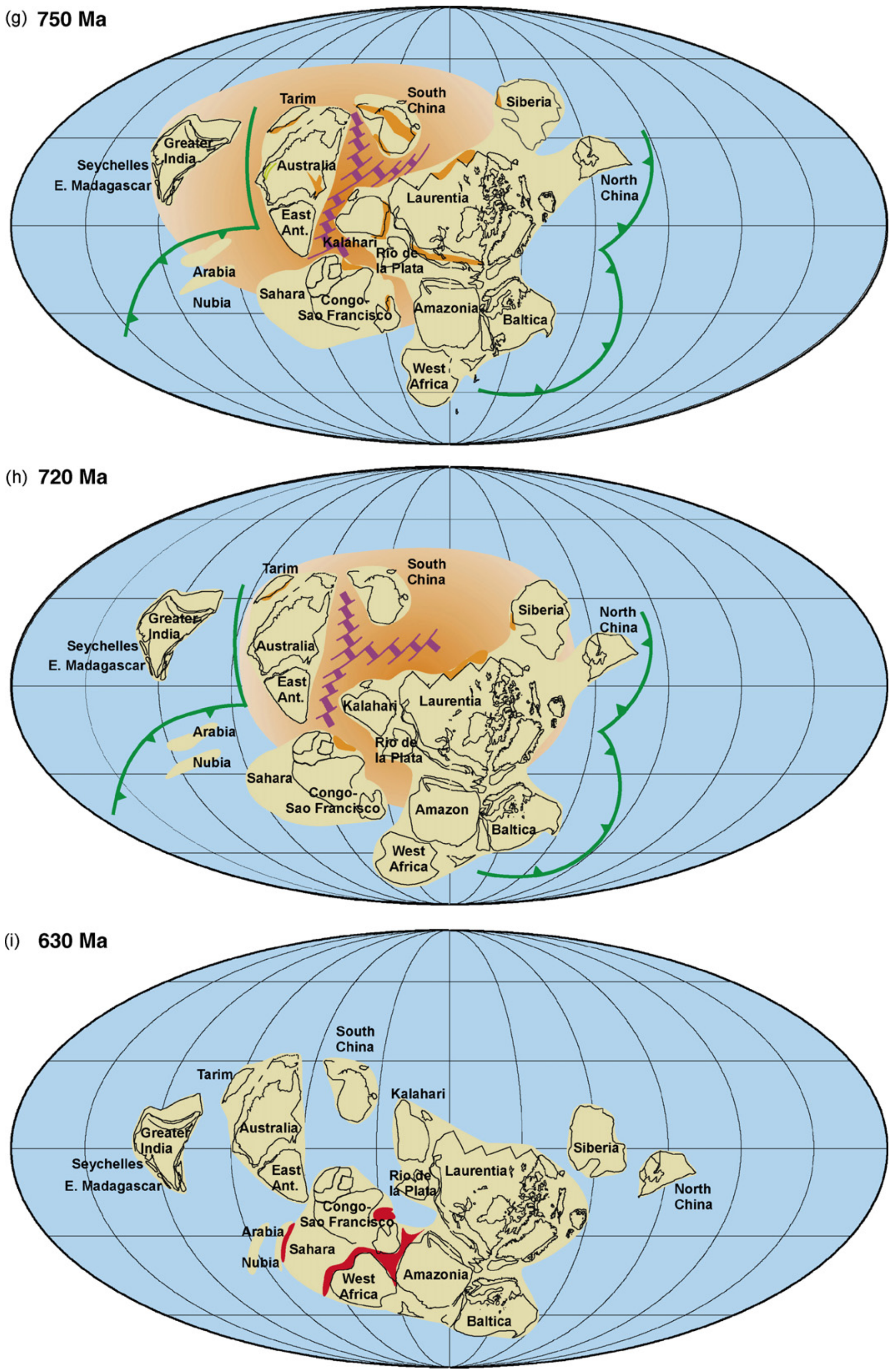

Fig. 9. (*Continued*)

metamorphic ages reported from the Pinjarra Orogen (Bruguier et al., 1999; Fitzsimons, 2003).

By ca. 900 Ma all major known continental blocks had aggregated to form the Rodinia supercontinent (Figs. 8 and 9d, and Appendix I). Evidence for ca. 900 Ma orogenic events include the ca. 920–880 Ma arc volcanics and ophiolite obduction in the eastern Sibao Orogen of South China (Li et al., 2005), 950–900 Ma arc volcanics along the northern margin of the Yangtze craton (Ling et al., 2003), and the 990–900 Ma high-grade metamorphic events in both the Eastern Ghats Belt of India

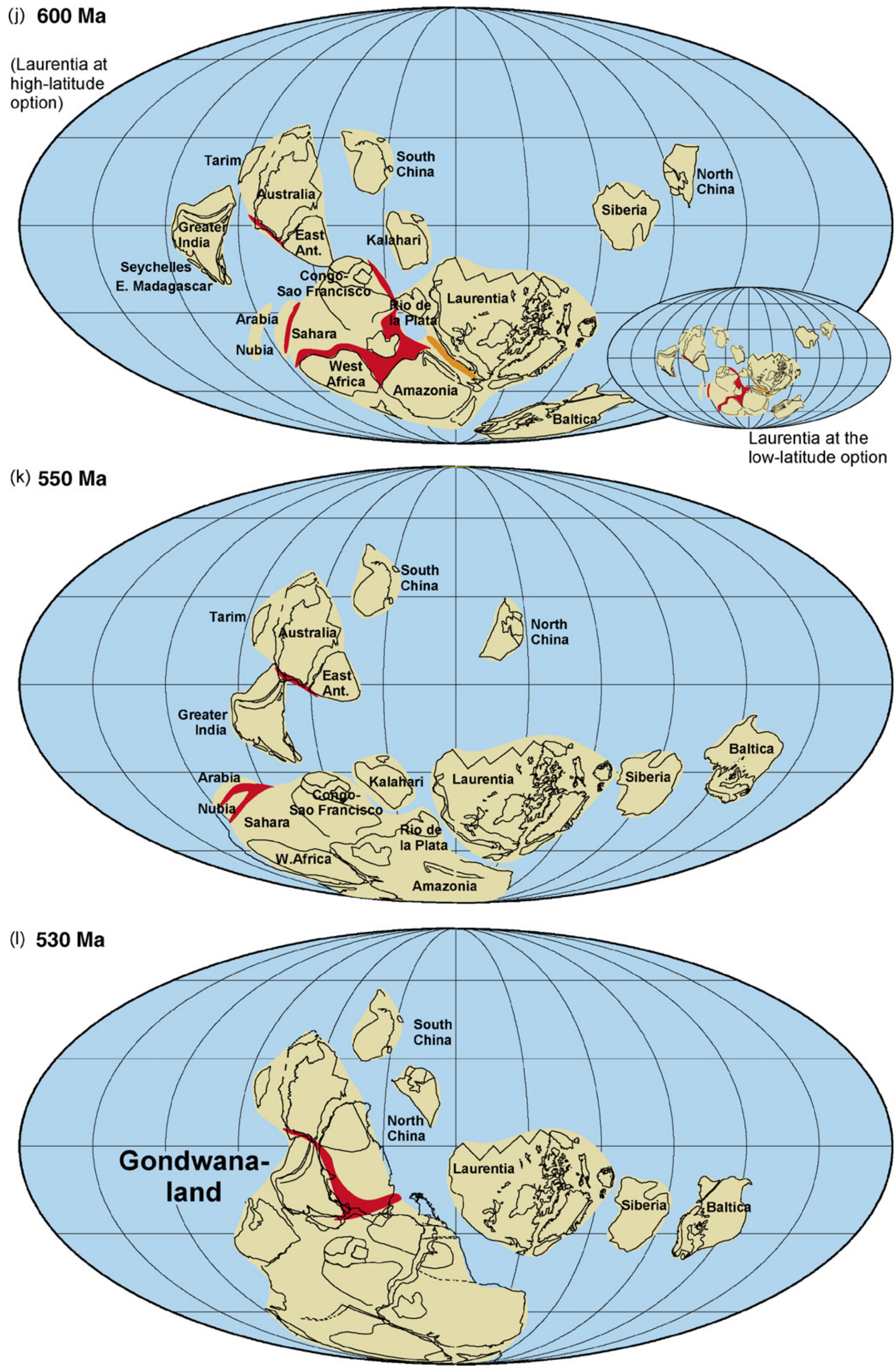

Fig. 9. (*Continued*).

and the corresponding Rayner Province in East Antarctica (e.g., Mezger and Cosca, 1999; Boger et al., 2000; Fitzsimons, 2000; Kelly et al., 2002). This may also be the time when the Tarim craton joined Australia, as indicated by the development of the Aksu blueschist which predates ca. 800 Ma mafic dyke intrusions and have ^{40}Ar–^{39}Ar cooling ages of 872–862 Ma (Zhang L.F., unpublished data, as quoted in Chen et al., 2004).

Stresses induced by the ca. 900 Ma event probably caused reactivation of older orogens within Rodinia. In the Mackenzie Mountains region of northwestern Laurentia, there is evidence for an east-west compressional event (the Corn Creek Orogeny) sometime between 1033 Ma and 750 Ma (Thorkelson, 2000; Thorkelson et al., 2005). In the southern Capricorn Orogen inside the West Australian craton, the Edmundian Orogeny continued until ca. 900 Ma as shown by ^{40}Ar–^{39}Ar ages (e.g., Occhipinti and Reddy, 2005). In the King Leopold Orogen of the North Australian craton, there was also the Yampi orogenic event (Tyler et al., 1998; Tyler and Griffin, 1990) with a possible ^{40}Ar–^{39}Ar age of ca. 900 Ma (Bodorkos and Reddy, 2005).

3.2. Superplume events, continental rifting, and the prolonged break-up process of Rodinia (ca. 860–570 Ma)

Palaeomagnetic constraints for the time interval between 900 Ma and 830 Ma are very poor. Little geological record is available within Rodinia from this time interval, apart from a small number of 870–850 Ma intrusions such as those in South China and Africa (X.H. Li et al., 2003b; Johnson et al., 2005). Ca. 845 Ma and 870 Ma bimodal intrusions have also been reported from the Scandinavian Caledonides (Paulsson and Andreasson, 2002) and the Scottish promontory of Laurentia (Dalziel and Soper, 2001), both interpreted as representing the beginning of Rodinia break-up. Z.X. Li et al. (2003b) suggested that these intrusions could have been the first sign of a Rodinia superplume, i.e. local anatectic melts due to enhanced thermal gradients above ascending plume-heads.

Widespread plume activity did not occur until ca. 825 Ma, as shown by mafic dyke swarms, intra-continental mafic–ultramafic intrusions, and felsic intrusions (resulting from crustal melt or magma differentiation). However, such magmatism is commonly found in the polar end of Rodinia only (Li et al., 2004), including Australia (Zhao et al., 1994; Wingate et al., 1998), South China (Li et al., 1999; X.H. Li et al., 2003a; Z.X. Li et al., 2003b), Tarim (Zhang et al., 2006), India (Radhakrishna and Mathew, 1996), Kalahari (Frimmel et al., 2001), and the Arabian–Nubian terranes (Stein and Goldstein, 1996; Teklay et al., 2002). In places these intrusive rocks are unconformably overlain by similar-aged rift-related volcaniclastic successions, suggesting syn-magmatic doming (e.g., Li et al., 1999). Li et al. (1999) and Z.X. Li et al. (2003b) interpreted this widespread, largely synchronous magmatic event as the first major episode of superplume events which eventually led to the break-up of Rodinia (Fig. 9e). They regard the globally common sedimentary hiatus between ca. 900–880 Ma and 820 Ma as partly due to plume-induced crustal unroofing.

Away from the suggested polar superplume, similar-aged gabbro and monzonite intrusions are reported from the Scandinavian Caledonides, which are interpreted as indicative of continental rifting (Reginiussen et al., 1995). However, it is unclear whether this event was related to the high-latitude superplume event as no such rocks have been reported in-between the two regions. Nonetheless, such a connection would be possible given the mechanism for the generation of the superplume (see discussions below).

We illustrate the possible formation and tomography of a superplume (Fig. 10). Although thermal insulation of a supercontinent may elevate the temperature of the upper mantle (e.g., Anderson, 1982; Gurnis, 1988), Li et al. (2004) suggested that the dominant driving force for the generation of mantle superplumes (or superswells) is the double (circum-Rodinia) push-up effects of mantle avalanches (e.g., Kellogg et al., 1999; Moores et al., 2000; Fig. 10). Such avalanches could occur once oceanic slabs subducting below margins of the supercontinent become too dense to be supported at the mantle transitional zone (e.g., Tackley et al., 1993). On the surface the superplume would appear as a “plume cluster” (Ernst and Buchan, 2002; Schubert et al., 2004) consisting of “secondary” plumes originating above the top of the superplume (Courtillot et al., 2003).

Such a mechanism would explain the bipolar nature of superplume occurrences, as illustrated by the present-day Africa and Pacific superplumes (e.g., Zhao, 2001; Courtillot et al., 2003), which are likely the residuals of the Pangaean superplume (e.g., Anderson, 1982; Burke and Torsvik, 2004) and the Palaeo-Pacific superplume (e.g., Larson, 1991a). The model would thus predict the occurrence of numerous oceanic plateaux and seamounts in the oceans on the opposite side of the Earth from Rodinia, traces of which would be found in the “Pan-African”-age orogens (e.g., parts of the Arabian–Nubian Shield; Stein and Goldstein, 1996). This model agrees with that of Condie (1998, 2000) in as much as superplumes are results of slab avalanches. It differs from the Condie (1998, 2000) model in that the superplume events in our model occurred after the formation of the supercontinent, not during the formation of the supercontinent. In our model the superplume events eventually led to the break-up of the supercontinent.

The ca. 825 Ma superplume event, followed by continental rifting, lingered on for ca. 25 million years, with another weaker magmatic peak occurring at ca. 800 Ma (Z.X. Li et al., 2003b; Ernst et al., 2008).

There appears to have been a global hiatus in plume/rift magmatism at around 790 ± 5 Ma (Z.X. Li et al., 2003b), but another superplume breakout occurred at ca. 780 Ma, with the best examples being the Gunbarrel event in Western Laurentia (Harlan et al., 2003b) and the Kangding event in South China (Z.X. Li et al., 2003b; Lin et al., 2007). By that time Rodinia had moved away from the northern polar region (Fig. 9f). India may have already rifted away from Rodinia if we accept that the palaeomagnetic pole from the Malani igneous suite (Torsvik et al., 2001b) is reliable and applicable for the entire 770–750 Ma interval, although no record of continental rifting and break-up has been reported for that time interval. The rapid rotation of Rodinia between ca. 820–800 Ma and ca. 780–750 Ma is interpreted by Li et al. (2004) as a true polar wander (TPW) event related to the occurrence of a high-latitude superplume (Evans, 1998, 2003b; Z.X. Li et al., 2003b). It is interesting to note that the rotation axis for the suggested TPW event is just east of Greenland (Li et al., 2004, Fig. 5), close to the Scandinavian Caledonides and the Scottish promontory of Laurentia where ca. 845 Ma and 870 Ma bimodal intrusions are present and are interpreted to represent the beginning of Rodinia break-up (Dalziel and Soper, 2001; Paulsson and Andreasson, 2002). Perhaps these magmatic events indicate the arrival of a plume-head which acted as the equatorial minimum-inertial axis for multiple TPW event(s) (Evans, 2003b; Li et al., 2004; Maloof et al.,2006).

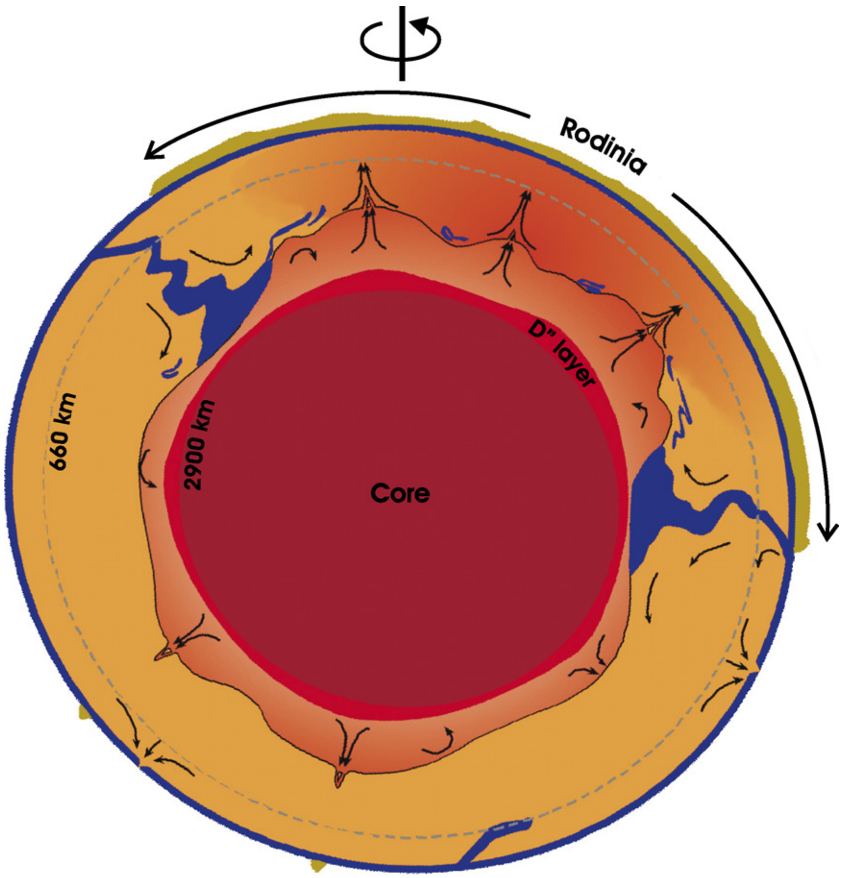

Fig. 10. A cartoon showing the formation of a mantle "superplume" beneath supercontinent Rodinia (inspired by Kellogg et al. (1999) and Moores et al. (2000)).

Our proposed mechanism for the formation of the Rodinia superplume (i.e. due to mantle avalanches surrounding the supercontinent, possibly enhanced by the thermal insulation effect of the supercontinent; Fig. 10), and the palaeomagnetic implication that both the superplume and Rodinia above it may have moved from a high-latitude position to an equatorial position between ca. 820–800 Ma and ca. 780–750 Ma (Li et al., 2004), imply that whole-mantle convection is primarily driven by subduction processes (a kind of "top-down tectonics", but differs from that of Anderson (2001) and others (e.g., Foulger et al., 2005) in that we have whole-mantle convection here), not by stationary heat sources at the core–mantle boundary.

By ca. 750 Ma (Fig. 9g), the western half of Rodinia may have started to break apart above an equatorial superplume. Bimodal magmatism at ca. 755–750 Ma was the last such major event seen across Rodinia (or a breaking-apart Rodinia). Like the ca. 825 Ma episode, syn-magmatic continental unroofing, documented by erosional contacts between ca. 770–750 Ma intrusive rocks and ca. 750–740 Ma rift successions, are common features (e.g., in western Kalahari, South China, and western and southeastern margins of Laurentia; see Z.X. Li et al., 2003b and references therein). By ca. 720 Ma (Fig. 9h), Australia–East Antarctica and South China are probably separated from each other by wide oceans. Even Kalahari and Siberia may have started to break away from Laurentia by this time. The ca. 750–700 Ma interval is also when the first major global Sturtian glaciation occurred and when most continents were located at low- to moderate-latitude positions (the first snowball-Earth? Kirschvink, 1992; Hoffman et al., 1998). At ca. 650–630 Ma, when the dispersing continental blocks became even further aligned along the palaeo-equator (Fig. 9i), the second widespread low-latitude glaciation (the Marinoan glaciation, and another snowball-Earth event? Hoffman and Schrag, 2002) occurred. Numerous potential causes for these enigmatic global climatic events have been proposed and all are highly controversial. We refer readers to a dedicated web site (http://snowballearth.org) on current understanding of the subject, and the bibliography listed therein.

By ca. 600 Ma, the Amazonia, West Africa and Congo-São Francisco cratons had largely come together during the Brasiliano Orogeny (e.g., Trompette, 1997) (but note the alternative suggestion of Trindade et al., 2006 that these did not collide until Cambrian times by a Pampean-Araguaia orogeny), while Amazonia and Rio de la Plata were probably still attached to Laurentia (Fig. 9j). Siberia (Pisarevsky et al., 2008), North China (Zhang et al., 2006) and Baltica (Cawood and Pisarevsky, 2006; Pease et al., 2005, 2006) were separated from Laurentia by ca. 600 Ma. However, equivocal 580–560 Ma palaeomagnetic data put Laurentia at either a high-latitude position, or a low-latitude position (Cawood and Pisarevsky, 2006 and references therein). In our model we show the high-latitude option as the main figure, but have the low-latitude alternative shown as an insert (Fig. 9j). The high-latitude option makes it easier to explain the ca. 580 Ma Gaskiers glaciation in Baltica (e.g., Meert and van der

Voo, 1994), but this does not apply to the likely post-Marinoan glaciation in Tarim (Xiao et al., 2004) which lacks precise palaeomagnetic constraints. Amazonia was likely separated from Laurentia by ca. 570 Ma (e.g., Cawood and Pisarevsky, 2006 and references therein). The rifting away of continental blocks (e.g., Siberia, Kalahari, Rio de la Plata, Amazonia and Baltica) from Laurentia between ca. 630 Ma and 550 Ma may have been responsible for the widespread subsidence recorded at the margins of many of these cratons (e.g., Bond et al., 1984).

3.3. The birth of Gondwanaland (600–530 Ma)

Hoffman (1991) first suggested that the break-up of the Rodinia supercontinent involved fragmentation around Laurentia with continental pieces moving away from Laurentia and colliding on the other side of the Earth to form Gondwanaland. As mentioned in Section 3.2, with the exception of Kalahari and perhaps some minor terranes, West Gondwana was largely together by ca. 600 Ma. However, oceans still existed between Australia–East Antarctica, India, eastern Africa and Kalahari at that time (e.g., Meert, 2003; Jacobs and Thomas, 2004; Collins and Pisarevsky, 2005). By ca. 550 Ma (Fig. 9k), India had moved closer to its Gondwanaland position along the western margin of Australia, as recorded by the sinistral strike–slip movement along the Pinjarra Orogen (Fitzsimons, 2003). Kalahari started to collide with Congo and Rio de la Plata, thus closing the Neoproterozoic Adamastor Ocean between them (e.g., Prave, 1996). North China, separating from Laurentia–Siberia after ca. 650 Ma (S. Zhang et al., 2006), was drifting across the palaeo-Pacific Ocean toward Australia at that time.

Gondwanaland finally amalgamated by ca. 540–530 Ma (Fig. 9l) through the closure of both the so-called “Mozambique Ocean”, causing the Malagasy Orogeny in the East African Orogen (Meert, 2003; Jacobs and Thomas, 2004; Collins and Pisarevsky, 2005), and the final docking of India to Australia–East Antarctica along the Pinjarra Orogen (Fitzsimons, 2003; Boger and Miller, 2004; Collins and Pisarevsky, 2005; also known as the Kuunga Orogen, Meert and Van der Voo, 1996; Meert, 2003). The formation of Gondwanaland by ca. 530 Ma is supported by palaeomagnetic analyses (e.g., Meert and Van der Voo, 1996; Li and Powell, 2001). Both the South China and North China blocks were located close to Australia, as indicated by episodic bioprovince connections with east Gondwana during the Early to mid-Palaeozoic (e.g., Burrett and Richardson, 1980).

4. Concluding remarks

The late Mesoproterozoic and Neoproterozoic period is one of the most remarkable time intervals in Earth’s history. During this time we see the assembly and break-up of the supercontinent Rodinia that was ancestral to the long-lived supercontinent of Gondwanaland, possible global superplume events and rapid true polar wander event(s), repeated low-latitude glaciations, and finally the explosion of multicellular life (McMenamin and McMenamin, 1990) and the emergence of a plate dynamic and climatic system very similar to what we have today (Moores, 2002; Evans, 2003a). The occurrence of some of these events, and possible genetic links between them, are still highly controversial. Nonetheless, these conceptual ideas enable us to think beyond the Phanerozoic “comfort zone” that persisted until recently. They have also forced geoscientists to think beyond their specific disciplines, bringing together tectonicists, structural geologists, geochemists, stratigraphers, petrologists, geophysicists, climatologists, and palaeontologists alike to work together toward a better understanding of Earth system science at the dawn of the Phanerozoic. With this background in mind, we present this overview of the formation and break-up of Rodinia. The scenarios presented here are not necessarily correct and there are many other scenarios which may explain some of the observations as well. It nonetheless provides a basis for future research on related topics.

Rodinia and Pangaea are the two supercontinents that we know to have included almost all the continents on Earth. Despite the secular nature of Earth’s evolution (and thus its tectonic processes), there are remarkable similarities between them: (1) They both had a lifespan of ca. 150 Ma: ca. 900 Ma to ≥750 Ma for Rodinia (this work), and ca. 320 Ma to 180–160 Ma for Pangaea (e.g., Li and Powell, 2001; Veevers, 2004); (2) The break-up of both supercontinents started with broad mantle upwellings (or superplumes) beneath them, resulting in widespread bimodal magmatism (including plume magmatism) and continental rifting (see Section 3.2); (3) The formation and break-up of Pangea also coincided with the occurrence of geomagnetic superchrons (Larson, 1991b; Eide and Torsvik, 1996), although there are still inadequate data for assessing the presence of similar events during the history of Rodinia. The Rodinia and Pangea supercontinental episodes thus provide us rare opportunities to gain insights into the 4D geodynamic system featuring close interactions between the thermal dynamics of the Earth’s outer core, mantle convections, and lithospheric plate tectonics.

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Not all co-authors agree on everything written in this paper, and the map of Rodinia presented in this paper reflects only one of the plausible models. We are grateful for support from UNESCO/IUGS/IGCP and their national committees, the Australian Research Council (through the Tectonics Special Research Centre and Discovery Project grant DP0450020), the National Natural Science Foundation of China and the Swedish Research Council. This work would not have been possible without the inspiration and leaderships of the late professors Chris Powell, Raphael Unrug, and Henri Kampunzu. We also acknowledge the contributions made by all members of IGCP 440, particularly J. Myers, M.K. Pandit, J.P. Milesi, Guojie Hao and Guanghua Niu who helped to compile the individual map sheets, and other contributors as shown on the Rodinia Map (Appendix I). The palaeogeographic reconstructions were made using PLATES developed at the University of Texas at Austin. Tectonics Special Research Centre publication #387, and a contribution to IGCP 440.

Appendix I

The Geodynamic Map of Rodinia – see printed version inside the back cover of this volume; an electronic version is available at doi:10.1016/j.precamres.2007.04.021.

Appendix II

From Rodinia to Gondwanaland – An animated history, 1100-530 Ma – PowerPoint animation available at (link to the PowerPoint file at the Elsevier online version) doi:10.1016/j.precamres.2007.04.021.

Appendix III

Rotation parameters for selected time slices during the assembly and break-up of Rodinia and the assembly of Gondwanaland – available at doi:10.1016/j.precamres.2007.04.021.

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