Supercontinents, superplumes and continental growth: the Neoproterozoic record

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Abstract: Between 1300 and 500 Ma the Neoproterozoic supercontinent Rodinia aggregated (1300–950 Ma), broke up (850–600 Ma) and a new supercontinent, Pannotia–Gondwana, formed (680–550 Ma). Only *c*. 11% of the preserved continental crust was produced during this 800 Ma time interval and most of this crust formed as arcs, chiefly continental margin arcs. At least 50% of juvenile continental crust produced between 750 and 550 Ma is in the Arabian–Nubian Shield and in other terranes that formed along the northern border of Amazonia and West Africa. An additional 20% occurs in Pan-African orogens within Amazonia, and *c*. 16% in the Adamastor and West African orogens. The growth rate of continental crust between 1350 and 500 Ma was similar or less than the average rate of continental growth during the Phanerozoic of 1 km^3/a, and this low rate characterizes both formation and breakup stages of the supercontinents.

The low rates of continental growth during the Neoproterozoic may be due to the absence of a superplume event associated with either Rodinia or Pannotia–Gondwana. If supercontinent breakup is required to produce a superplume event, perhaps by initiating catastrophic collapse of lithospheric slabs at the 660 km seismic discontinuity, the absence of a Meso-proterozoic–Neoproterozoic superplume event may mean that a Palaeoproterozoic supercontinent did not fully breakup prior to aggregation of Rodinia.

Although it is now recognized that continental crust has grown rapidly at several periods in the geologic past, the relationship of production rate of continental crust to the supercontinent cycle is not well understood (Condie 1998, 2000). Is there a change in the growth rate of continental crust during the supercontinent cycle? For instance, do continents grow more rapidly during the formation stage of a supercontinent than they do during the breakup stage? During the aggregation of supercontinents, juvenile crust, such as arcs, oceanic plateaus and ophiolites, can be trapped in collisional orogens, thus adding mass to the continents. However, during the breakup stage of supercontinents, oceanic crust forms more rapidly in response to new ocean ridges, and increased mantle plume activity may also produce more oceanic plateaus and flood basalts. This leads to the question of where and in what tectonic setting does most continental crust originate. Although an unknown but probably small amount of continental crust is added by underplating of mafic magmas from plume sources, it would appear that the bulk of continental growth occurs when juvenile crust is trapped in collisional orogens between colliding cratons, or in peripheral orogens, as oceanic terranes collide with continental margins. Are periods of rapid continental growth associated with superplume events in the last 1 Ga as suggested by Condie (1998, 2000) for earlier periods, and if so, what, if any, relationship exists between superplume events and supercontinents?

Using the Neoproterozoic–Early Phanerozoic supercontinents Rodinia and Gondwana as examples, in this chapter questions related to the growth of continental crust and the relationship, if any, between crustal growth and the supercontinent cycle will be addressed. The relationship of continental growth to the early rifting and breakup stages of Rodinia (1100–750 Ma) and to the later collisional events that led to formation of the younger supercontinents, Gondwana and Pannotia (750–550 Ma), will also be examined. First of all, it is important to review the configuration of plates in Rodinia, discuss uncertainties in plate location and summarize the tectonic history of Neoproterozoic supercontinents. The Sr isotopic record of seawater during the Neoproterozoic as a proxy for supercontinent evolution will also be reviewed.

Rodinia configuration

Although the configuration of plates in Rodinia is becoming better constrained with more precise isotopic ages, the connection between East Gondwana (Antarctica, Australia, India) and Laurentia is still uncertain, as is the position of many of the smaller plates such as North China and Malaysia. The SW US–East Antarctic (SWEAT) reconstruction of East Gondwana proposed by Moores (1991) and Dalziel (1991) places Antarctica adjacent to southwestern Laurentia and Australia adjacent to western Canada. Karlstrom *et al*. (1999, 2001) and Burrett & Berry (2000) suggested that Australia satisfies more

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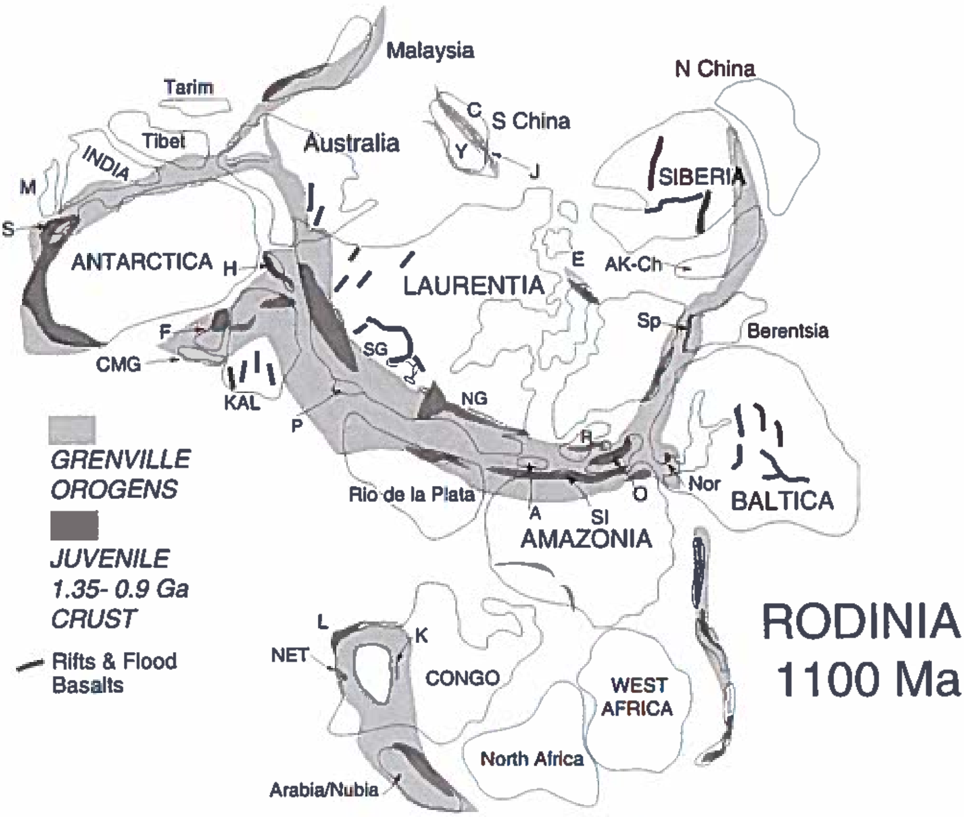

Fig. 1. Diagrammatic reconstruction of the Neoproterozoic supercontinent Rodinia showing the distribution of Grenvillian collisional orogens and 1350–900 Ma juvenile crust. Reconstruction after Hoffman (1991), Li & Powell (1995), Karlstrom *et al.* (1999) and Condie (2001*a*). Symbols: A, Arequipa; P, Precordillera; AK–Ch, Arctic Alaska–Chukotka; CMG, Coats Land–Maudheim–Grunehogna (East Antarctica); KAL, Kalahari Craton; M, Madagascar; S, Sri Lanka; Y, Yangtze Block; C, Cathaysia Block; Fk, Falkland Plateau; H, Haag Nunataks and parts of the Transantarctic Mountains; SG and NG, southern and northern Grenville Province, respectively; E, northern Ellesmere Island; R, Rockall; O, Oaxaquia; SI, San Ignacio Orogen; L, Lurio Belt; Sp, Svalbard; J, Jinning Orogen; K, Katangan Orogen; Nor, Norway; NET, NE Tanzania. Malaysia block includes Cambodia, Thailand and Vietnam.

tectonic constraints and palaeopole positions if it placed near southwestern Laurentia (AUSWUS), a fit which is tentatively adopted and illustrated in Figure 1 (all directions are relative to the present continental positions). AUSWUS matches the Grenvillian Orogen in southern Laurentia with the Arunta–Fraser–Albany Orogen of similar age in central and southwestern Australia. It also places the Yavapai–Mazatzal Palaeoproterozoic Provinces in southern Laurentia adjacent to the Broken Hill and Mount Isa terranes in eastern Australia, which have similar isotopic ages. AUSWUS also provides a somewhat better fit of palaeomagnetic data than SWEAT (Karlstrom *et al.* 2001). Palaeomagnetic data suggest that the South China block was located near Australia, with possible positions both NW and NE of Australia (Evans *et al.* 2000): the latter position is adopted in Figure 1.

Although most investigators agree that Siberia was connected to northern Laurentia during the Neoproterozoic, there is little agreement on the geometry of the fit. At least three configurations have been proposed using various piercing points, matching of alleged conjugate basins and palaeomagnetic results (Hoffman 1987; Condie & Rosen 1994; Pelechaty 1996; Frost *et al.* 1998). Still another interpretation places Siberia adjacent to the west coast of Laurentia (Sears & Price 2000). Although the problem of where Siberia and Laurentia were connected is far from being solved, the piercing points presented by Frost *et al.* (1998) are convincing, and their configuration is tentatively adopted in reconstructions given in this chapter. Contributing to the problem of the Siberia–Laurentia connection is the uncertainty of just when Siberia was rifted from Laurentia. Evidence has been cited to support both

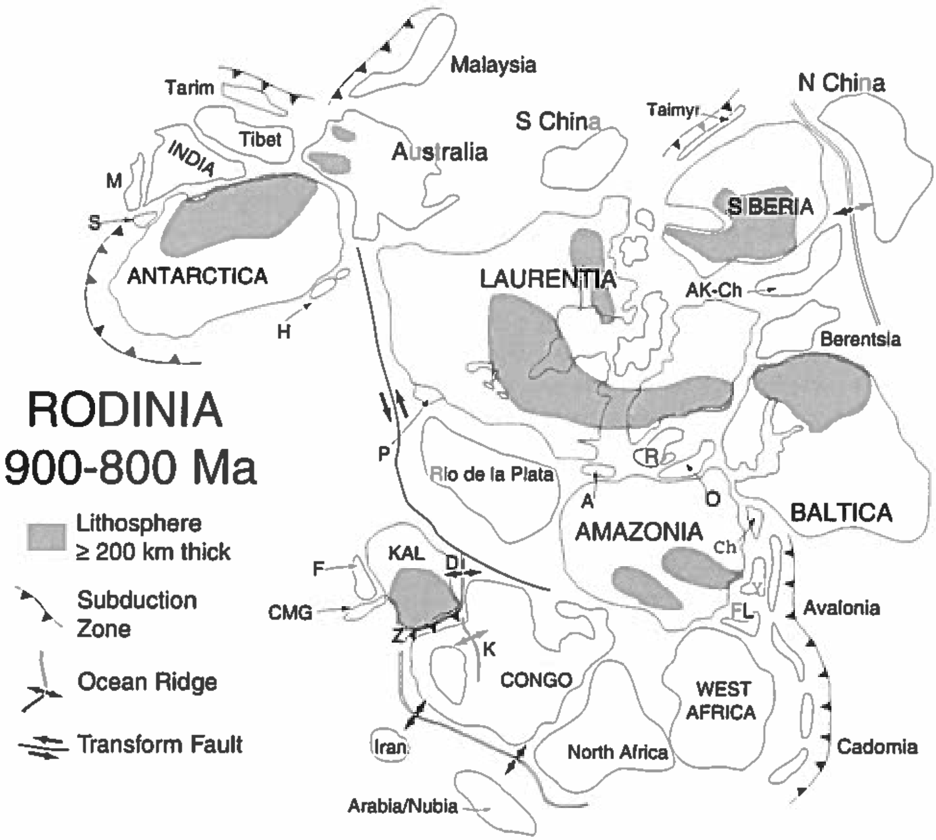

Fig. 2. Diagrammatic reconstruction of Rodinia at 900–800 Ma showing possible distribution of arc systems, rifts and flood basalts, ocean ridges, major transform faults, and thick Archean lithosphere. D, Damaran Orogen; Z, Zambezi Orogen; Ch, Chortis; Y, Yucatan. Other symbols defined in Figure 1.

Neoproterozoic and Early Cambrian rifting, but only Neoproterozoic rifting is acceptable in terms of palaeomagnetic data, the latter of which show that Laurentia and Siberia were not connected in the Early Cambrian (Kirschvink *et al.* 1997). If the Franklin Dyke swarms in northern Canada are related to similar aged dykes in northern Siberia, rifting of Siberia from Laurentia may have occurred at 720–700 Ma (Condie & Rosen 1994). Plate reconstructions in the Arctic region based on U–Pb zircon ages suggest that northern Alaska and Chukotka in eastern Siberia were part of the same plate during the Neoproterozoic. A striking similarity of pre-Caledonian magmatism in eastern Laurentia, Baltica and Svalbard to that in Arctic Alaska–Chukotka suggests that the latter plate was located between Siberia and Berentsia (Fig. 1; Patrick & McClellan 1995), although the exact location is unknown. Palaeomagnetic data suggest that Taimyr was located near Siberia in the Neoproterozoic; Neoproterozoic rock assemblages suggest arc affinities (Fig. 2; Vernikovskiy *et al.* 1998). The location of North China in Rodinia continues to be problematic. Similarities of Changcheng and Jixian Proterozoic sediments in North China to Riphean 1 and 2 sediments in Siberia are consistent with a close connection of Siberia and North China (Li & Powell 1995). Also, correlation of Late Neoproterozoic sediments and palaeomagnetic data support this interpretation (Figs 1 & 2; Halls *et al.* 2000).

Most investigators agree that Baltica, Amazonia and Rio de la Plata surrounded the northeastern and eastern margins of Laurentia, with bits and pieces of small plates (such as Rockall and Arequipa) tucked in between (Fig. 1). Nd and Pb isotope data suggest that Oaxaquia, today comprising much of the basement beneath Mexico, was located in NW Amazonia until at least 1100 Ma (Lopez *et al.* 2001). Palaeomagnetic, palaeontologic, and U–Pb zircon ages indicate that numerous terranes now found in the Appalachian–Caledonian Orogen in eastern Laurentia (such as Avalonia, Florida and Carolina), and in the Variscan and Alpine Orogens in central and southern Europe (such as Cadomia, Amorica and Iberia), began life along the northern margins of Amazonia and West Africa (Liegeois *et al.* 1996; Erdtmann 1998; Murphy *et al.* 1999). Recent palaeomagnetic data suggest that the Kalahari Plate, rather than being connected to East Antarctica as preferred by some investigators (Hoffman 1987), was located south of southern Laurentia in the Neoproterozoic. Dalziel *et al.* (2000) have recently proposed that it collided with southern Laurentia at *c.* 1100 Ma, producing the Grenvillian deformation in this area.

Although the precise configurations are unknown, cratons in Congo, North Africa (Nile Craton) and West Africa appear to have been located east of Amazonia in the Neoproterozoic (Fig. 1). As with the India–Malaysia side of East Gondwana, subduction zones surrounded part of West Gondwana, and in particular were active in what is today the Arabian–Nubian Shield. By 800 Ma, subduction had also developed adjacent to West Africa and Amazonia, producing arc systems (Fig. 2), remnants of which are the terranes that became part of eastern Laurentia and southern Europe. The positioning of terranes such as Avalonia, Cadomia and Florida adjacent to Amazonia and West Africa is supported both by palaeomagnetic results and by U–Pb zircon ages (Opdyke *et al.* 1987; Nance *et al.* 1991; Keppie & Ramos 1999; Murphy *et al.* 1999).

Tectonic history of Rodinia

Early Rodinian rifting (1100–900 Ma)

The first rifting in Rodinia began some 300 Ma before the supercontinent actually began to fragment. This is recorded by widespread extensional deformation beginning at *c.* 1100 Ma, with examples such as the Mid-continent Rift System (1108–1086 Ma) in central Laurentia (Cannon 1994; Davis & Green 1997; Timmons *et al.* 2001) and the Umkondo rifts in Kalahari (1100 Ma) (Fig. 1; Hanson *et al.* 1998). Similar, but not as well-dated, rifting occurred in Siberia and Baltica. Timmons *et al.* (2001) have recently recognized two rifting events in the Neoproterozoic of southwestern Laurentia, the older of which at 1100 Ma has a northwesternly trend reflecting NE–SW extension, perhaps related to the Grenville collisions (Fig. 1). These structures may correlate with similar aged structures in central Australia, lending support to the AUSWUS Rodinia reconstruction. The 1100 Ma rifting was accompanied by eruptions of flood basalts, probably associated with mantle plumes beneath the growing supercontinent. An unknown but relatively small amount of mafic juvenile crust may have underplated the continents at this time. It is noteworthy that this early rifting followed Grenvillian collisional events, which occurred between 1250 and 1190 Ma along the margin of eastern Laurentia (Rivers 1997), at 1150–1120 Ma (continuing to 980 Ma) in southern Laurentia and in the Namaqua Orogen in Kalahari (Figs 1 & 2; Dalziel *et al.* 2000; Condie 2001*a*; Knoper *et al.* 2001). Also recording extension are dyke swarms intruded at 1100 Ma and again at 800–750 Ma in southwestern Laurentia and in Kalahari at 1100 Ma. The 800–750 Ma dykes are also recognized in Australia (Park *et al.* 1995). The 20–100 Ma time interval between the 1240–1120 Ma Rodinia collisions and the first rifting may have been the time needed for a large mantle upwelling to develop beneath the new supercontinent.

Renewed collision along the Grenvillian orogens occurred between eastern Laurentia and Amazonia–Rio de la Plata at 1080–1020 Ma, and between Kalahari and southern Laurentia at 1060–980 Ma (Rivers 1997; Dalziel *et al.* 2000). During these collisions, the Mid-continent and related Laurentian rifts were partly closed by thrust faulting related to the collisions (Cannon 1994). Following the Grenvillian collisions in Rodinia, intracratonic rifting continued between 900 and 750 Ma, when rift basins in southwestern Laurentia (Uinta, Chuar, Pahrump) and in Siberia opened. Unlike the earlier rifting event in southwestern Laurentia, the rifts in this event (800–740 Ma) have a dominantly northern trend reflecting E–W extension, consistent with the rifting of East Gondwana from the west coast of Laurentia at this time (Timmons *et al.* 2001). The Uinta Rift in northern Utah, with an E–W trend, is a notable exception to the N–S trends. Aborted rifting also occurred in southern Australia betweeen 830 and 800 Ma as various rifts in the Adelaide Basin opened (Preiss 2000). In what is now NE Africa and Arabia, between 870 and 700 Ma oceanic arcs and back-arc basins collided to form superterranes, and these later collided to form the Arabian–Nubian Shield (Abdelsalam & Stern 1996). Studies of orogenic terranes in East Antarctica suggest that an arc system existed in this area from *c.* 1150 to 1070 Ma, perhaps to 900 Ma (Jacobs *et al.* 1998) (Fig. 2).

It is important to note that none of the intracratonic rifting before *c.* 900 Ma actually fragmented the supercontinent. Dalziel *et al.* (2000) suggest that the pre-900 Ma rifts in Laurentia and Kalahari may have been impactogens related to the Grenville collisions, and as such they would not be expected to fragment the supercontinent. Arguing against an impactogen origin, however, is the fact that most of the rifting (at 1100 Ma) did not coincide with Grenvillian collisions, which occurred before 1120 Ma or after 1100 Ma. Even more important is the large amount of basaltic magma associated with the Mid-continent rift, as well as other 1100 Ma rifts,

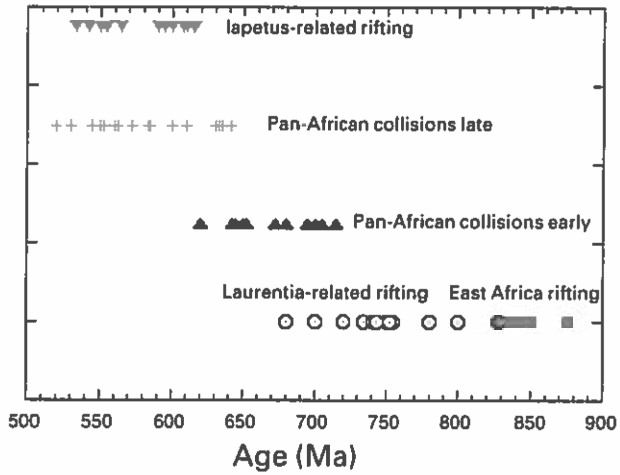

Fig. 3. Summary of the timing of rifting and collisional events in Rodinia between 900 and 500 Ma; references given in Table 1.

which favours a mantle-plume origin rather than an impactogen origin for these rifts.

A puzzling question is why none of the 1100 Ma rifts fragmented the supercontinent Rodinia, but rather they became inactive prior to major supercontinent breakup at 850–750 Ma. As described below, the major fragmentation of Rodinia occurred along peripheral rift systems around to the margins. Perhaps the thick lithosphere beneath Archean cratons in the central part of Rodinia gave additional strength to the lithosphere (Fig. 2). Thus, the rifts that eventually broke up the supercontinent were those that developed in thinner lithosphere closer to the margins of the supercontinent. Also, as pointed out by Krabbendam & Barr (2000), in large orogens such as the Grenville, collision may result in partial delamination and removal of the mantle lithosphere, thus weakening the lithosphere and making it more susceptible to later rifting.

Breakup of Rodinia (850–600 Ma)

The earliest fragmentation of Rodinia began in the East African and Katangan Orogens 900–850 Ma (Figs 2 & 3) (Stern 1994). If Kalahari was part of Laurentia at 1000 Ma (Dalziel *et al.* 2000), rifting of Kalahari from Laurentia must have begun at *c.* 900 Ma and involved largely transform faulting (Fig. 2). As recorded in the Zambezi Orogen, Kalahari collided with the Congo Craton at *c.* 820 Ma, which can be considered as the first major collision in the formation of Gondwana (Hanson *et al.* 1994). At 842 Ma, part of the Arabian–Nubian Shield was rifted from the North African Craton, perhaps in response to a mantle plume that produced an oceanic plateau, a remnant of which is preserved as the Baish Group in the western Arabian Shield (Kroner *et al.* 1992; Stein & Goldstein 1996). Rodinia continued to fragment between 850 and 750 Ma as East Gondwana and South China were rifted away from western Laurentia (Figs 3 & 4). Very little if any juvenile crust survives that accompanied the breakup along the west coast of Laurentia. Although continental blocks appear to have been rifted from the eastern margin of the Congo Craton between 900 and 850 Ma (Stern 1994), the location of these blocks today is unknown; perhaps one of the blocks is the Iran Plate (Fig. 2). Because arc-related igneous rocks are not found in the Katangan Orogen, the ocean basin that opened must have been small (Red Sea size), such that the basin margins collided before subducted slabs reached melting depths (*c.* 70 km) (Kampunzu *et al.* 1991). Extensive flood basalts in the Katangan Orogen at *c.* 850 Ma probably record a mantle plume centered near the triple junction of the Katangan, Damaran and Zambezi Orogens (Fig. 2; Kampunzu *et al.* 2000). Although the Katangan Rift continued to open between 850 and 775 Ma, it appears that only a small volume of oceanic crust was formed before it rapidly closed at 750 Ma. The Katangan Collision was later than the Zambezi Closure at 820 Ma but considerably earlier than the Damaran Closure at 600–550 Ma. Rifting continued in and south of the Arabian–Nubian Shield during the interval of 870–840 Ma, as recorded by zircon ages from the Namaqua, Gariep, Damara, Zambezi and Katangan Orogens (Table 1; Frimmel *et al.* 2001). In the East African Orogen, this rifting led to the development of a passive continental margin,

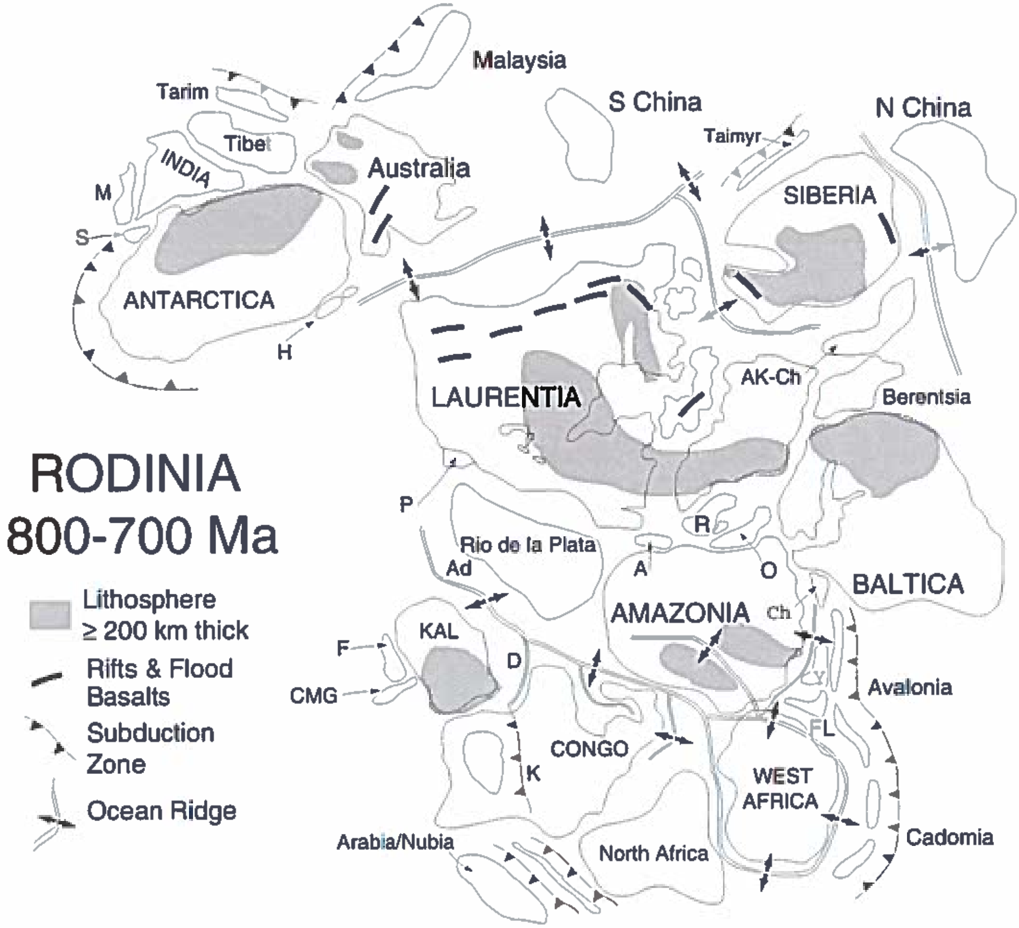

Fig. 4. Diagrammatic reconstruction of Rodinia at 800–700 Ma showing possible distribution of arc systems, rifts and flood basalts, ocean ridges, and thick Archean lithosphere. Ad, Adamastor Ocean; other symbols defined in Figures 1 & 2.

now recorded by remnants of sediments in Kenya (Stern 1994).

Most of the rifting in Avalonia occurred between 800 and 750 Ma (Table 1). Rifting at *c.* 760 Ma is recorded in Newfoundland, in the Maritimes in eastern Canada, in New England and in the Carolina Province in North Carolina (Harris & Glover 1988; Nance *et al.* 1991; Murphy *et al.* 1999). Arc magmatism occurred in Avalonia between *c.* 670 and 580 Ma, followed by collisions and accretion to Laurentia between 640 and 560 Ma (Fig. 4). Oblique shear zones with associated rift basins and post-tectonic intrusions formed after accretion at *c.* 540 Ma. Similar ages are reported in terranes that comprise Cadomia, many of which are between 620 and 600 Ma (Samson & D'Lemos 1998). Final accretion of Cadomian and Avalonian Terranes to Baltica or Laurentia occurred chiefly between 580 and 540 Ma (Table 1). Although the positions of Tarim and Malaysia are not well known, rock assemblages suggest that arc systems also existed in these areas during the Neoproterozoic (Condie 2001*a*).

Pan-African orogenic system (750–550 Ma)

A Pan-African orogenic belt appears to have completely surrounded the West African Craton during the Neoproterozoic (Fig. 5). This orogen is exposed in NW Africa as the Anti Atlas, Ougarta, Pharusian–Tuareg, Gourma and Dahomeyan Belts on the east, and the Mauretanian, Bassaride and Rokelide Belts on the west (Caby 1987; Dostal *et al.* 1994; Attoh *et al.* 1997; Hefferan *et al.* 2000). All of these deformation belts developed during Pan-African collisions, chiefly between 650 and 550 Ma (Table 1; Fig. 3),

Table 1. *Distribution of Neoproterozoic juvenile crust in Rodinia*

Location	Crustal age (Ma)	Rifting age (Ma)	Accretion age (Ma)	Juvenile crust (%)	Scaled area ($\times 10^6$ km^2)*	Arc (%)	Ocean ridge (%)	WPB+ (%)	References
Arabian–Nubian Shield	900–850	900–850	700–650	90	1.02 [0.92]	50	5	45	Duyvermna *et al.* 1982; Reischmann *et al.* 1984; Kroner *et al.* 1987;
	700–650		760–650	95	1.02 [0.92]	94	5	1	Stern & Kroner 1993; Stern 1994; Stein & Goldstein 1996
Brazil	900–760	800–750	650–550	85	0.98 [0.83]	90	10	0	Pimentel & Fuck 1992; Babinski *et al.* 1996; Machado *et al.* 1996; Pimentel *et al.* 1996; de Almeida *et al.* 2000
Africa Adamastor					0.97 [0.32]				Porada 1989; Kukla & Stanistreet 1991; Frimmel *et al.* 1996*a*, *b*
Damara	750–650	750	600–550	20		20	70	10	
Kaoko–W Congo	750–650	800–750	600–570	10		0	100	0	
Gariep–Saldanian	740–720	750	600–580	50		10	60	30	
Kibaran–Lufilian	750–650	900–760	570–530	50	0.25 [0.12]	50	0	50	Kampunzu *et al.* 2000
West African Craton					0.642 [0.32]				Black *et al.* 1979; Chikaoui *et al.* 1980; Caby 1987;
Pharusian	750–600	850–800	645–560	70		90	10	0	Villeneuve & Dallmeyer 1987;
Mauritanide–Bassaride	700–650	850–700	660–640	40		100	0	0	Dallmeyer & Villeneuve 1987;
Dahomeyide	700–600	900–800	640–560	15		80	20	0	Saquaque *et al.* 1989; Dostal *et al.* 1994; Hefferan *et al.* 2000
Anti Atlas	780–600	800–780	615–565	80		95	5	0	
Rokelide	750–600	850–700	550	50		80	20	0	
East Africa	900–850	900–850	600–550	20	0.3 [0.06]	80	20	0	Maboko *et al.* 1985; Maboko 1995; Kroner *et al.* 1999; Moller *et al.* 2000

Table 1. (*cont.*)

Location	Crustal age (Ma)	Rifting age (Ma)	Accretion age (Ma)	Juvenile crust (%)	Scaled area ($\times 10^6$ km^2)*	Arc (%)	Ocean ridge (%)	WPB+ (%)	References
Cadomia-Dalradian					0.45 [0.36]				Cabanis *et al.* 1987; Pharaoh *et al.* 1987; Winchester *et al.* 1987; Murphy *et al.* 1991; Nance *et al.* 1991; Liegeois *et al.* 1996; Hefferan *et al.* 2000
Carpathians	800–780			100		100	0	0	
British Isles	700–600	800–750	580–540	80		75	15	10	
America–Normandy	700–550	800–750	580–540	80		80	0	20	
Dalradian	680–550			50		50	0	50	
Avalonia–Florida					0.43 [0.39]				Feiss 1982; Opdyke *et al.* 1987; Harris & Glover 1988; Nance *et al.* 1991; Nance & Murphy 1994; Barr & White 1996; Murphy *et al.* 1999; Hefferan *et al.* 2000
Maritimes	700–600	800–750	580–570	80		75	15	10	
Carolina	700–600	800–750	600–540	80		80	10	10	
New England	675–550	800–750		100		80	10	10	
Florida	650–550	800–750		100		100	0	0	
Siberia					0.05 [0.04]				Rosen *et al.* 1994; Vernikovskiy *et al.* 1998
Taimyr	800–750		600	80		80	20	0	
Yenisey	800–600		600	80		75	15	10	

* Values in [ ] corrected for reworked component with Nd isotopes using the values given for the percentages of juvenile crust.
+ WPB, Within-plate basalts, chiefly oceanic plateau and flood basalts.

and all record remarkably similar tectonic histories, with initial rifting occurring between 850 and 700 Ma (Hefferan *et al.* 2000). Although pre-collisional subduction appears to have been directed chiefly outward from the West African Craton, little agreement exists as to the size of the ocean basins that opened. Indeed, if closure occurred during a 100 Ma interval on all sides of the West African Craton, it is unlikely that the intervening ocean basins were as large as originally suggested by Caby (1987). Most of the Pan-African juvenile crust trapped in the circum-West African orogens formed as continental margin arcs (Chikhaoui *et al.* 1980; Caby 1987; Saquaque *et al.* 1989; Dostal *et al.* 1994).

Rifting contemporary with that in West Africa occurred in Amazonia and probably also in Rio de la Plata (Castaing *et al.* 1994; de Almeida *et al.* 2000). Numerous blocks of Archean and Palaeoproterozoic crust were rifted apart in eastern Amazonia (Brazil) as small ocean basins formed between blocks. Continental margin arcs formed around the margins of many of these blocks and are the principal type of Neoproterozoic juvenile crust (Table 1). Small volumes of ophiolite were also trapped in the sutures. Closure of these basins occurred chiefly between 650 and 550 Ma, similar to basin closures around the West Africa Craton (Fig. 5). During collisions of continental blocks, which caused the Braziliano Orogeny, large volumes of pre-1 Ga continental crust were reworked and partially melted (de Almeida *et al.* 2000). In most cases, the final stages of collision are characterized by transcurrent shear zones.

Numerous continental margin arcs were accreted to the Arabian–Nubian Shield during the Pan-African collisions (Klemenic *et al.* 1985; Furnes *et al.* 1996; Alene *et al.* 2000). The sutures between oceanic terranes often contain well-preserved ophiolites (Zimmer *et al.* 1995). The closure of the Mozambique Ocean, in what is now Saudi Arabia, between 750 and 650 Ma resulted in collision of rifted blocks (such as the Afif Terrane) and juvenile arcs located between East and West Gondwana along north-trending sutures (Fig. 5). Continued convergence between 650 and 540 Ma led to the formation of NW-trending sinistral and NE-trending dextral transcurrent fault systems.

Mafic dyke swarms probably associated with mantle plumes indicate diachronous opening of the Iapetus Ocean, beginning at 620–600 Ma between Baltica and Greenland and spreading southward between the Appalachian Orogen and Amazonia–Rio de la Plata by 600–550 Ma (Figs 3 & 4; Bingen *et al.* 1998). As indicated by tectonic subsidence models (Bond & Kominz 1984), renewed rifting occurred in western Laurentia (Canadian Cordillera) at 600–550 Ma. Geologic evidence for rifting in the Death Valley area of California also supports renewed rifting at this time (Prave 1999). Although terranes were undoubtedly rifted from what is now western Canada, it is not clear where these terranes are located at present. The fragmenting pieces associated with Iapetus Ocean opening eventually collided, becoming part of Gondwana, with major collisions occurring dominantly in two stages (Fig. 5). The oldest collision occurred when part of East Gondwana (India?) collided with West Gondwana, producing deformation in the Mozambique Orogen at 725–650 Ma. Although some juvenile crust was trapped in the central and southern parts of the Mozambique Orogen, most of the crust in this part of the orogen is reworked Palaeoproterozoic and Archean crust (Maboko 1995). This early Mozambiquian collision was followed by the widespread Pan-African event at 650–550 Ma (Fig. 3). At this time the interconnecting array of rifted small ocean basins in western Gondwana closed, producing the Pan-African and Braziliano Orogenies. Most of the Neoproterozoic juvenile crust preserved today was 'captured' between colliding continental blocks during one of these collisional events.

An important question is why none of the Pan-African ocean basins in West Gondwana developed into large oceans before convergence began. Grunow *et al.* (1996) have suggested that it was the synchronous opening of the Iapetus Ocean and the Pan-African collisions at 650–550 Ma that was responsible for premature closure of the Pan-African basins (Figs 3 & 4). In effect, West Gondwana was caught between compressive forces on both sides, which closed the small ocean basins leading to the widespread Pan-African and Braziliano Deformation.

Rifting of Rio de la Plata from the Congo and Kalahari Plates occurred chiefly between 800 and 750 Ma as the Adamastor and Khomas Oceans opened (Fig. 4; Porada 1989; Kukla & Stanistreet 1991; Frimmel *et al.* 1996*a*, *b*). By 717 Ma, subduction began in the Gariep Belt in South Africa and the Adamastor Ocean began to close, with a continent–continent collision occurring at *c.* 545 Ma. Closure of the Adamastor Ocean and related ocean basins, and accretion of juvenile Neoproterozoic crust, occurred chiefly between 600 and 550 Ma (Table 1). Sedimentological studies in the Damara Orogen in Namibia indicate that the Congo–Rio de la Plata suturing predated the Congo–Kalahari suturing (Prave 1996).

Sr isotopic record of Rodinia

As shown by Veizer (1989), the Sr isotope record in marine carbonates can be useful in tracking the relative inputs of continental and mantle Sr into seawater. When land areas are extensive and elevations relatively high, weathering and erosion transport

Fig. 5. Diagrammatic reconstruction of Rodinia at 700–550 Ma showing possible distribution of arc systems, rifts and flood basalts, and ocean ridges. Symbols defined in Figures 1 & 2.

large amounts of continental Sr into the oceans, which has relatively high $^{87}Sr/^{86}Sr$ ratios. In contrast, when the sea level is high and ocean ridges or/and mantle plumes are widespread and active, the input of mantle Sr into seawater is enhanced. Many of the long-term changes in the Sr isotopic composition of seawater may be related to the supercontinent cycle and perhaps also to superplume events (Condie *et al.* 2000). Although the Sr isotope record of seawater for the Phanerozoic is relatively well known (Veizer *et al.* 1999), except for the Neoproterozoic, the record for Precambrian seawater is poorly known (Asmerom *et al.* 1991; Jacobsen & Kaufman 1999).

In principle, it should be possible to track the formation and destruction of a supercontinent with the Sr isotopic record in seawater. During supercontinent formation, when the land area is increasing, the $^{87}Sr/^{86}Sr$ ratio of seawater should increase; during fragmentation, when ocean ridges and mantle plume activity increase, the ratio should decrease. It is not yet possible to completely track the history of Rodinia with Sr isotopes, since the seawater Sr isotope record is only well known for the last *c.* 900 Ma onwards (Fig. 6). There are few Sr isotope data from well-dated sediments corresponding to the formation of Rodinia between *c.* 1300 and 900 Ma. Sr isotope ratios from well-dated limestones of the Belt Supergroup at 1470 Ma are *c.* 0.7062 (Veizer & Compston 1976; Evans *et al.* 2000*b*). Carbonates from China, with less well-constrained ages in the 1000–1300 Ma range, have low Sr isotope ratios of 0.704–0.705 (Jahn & Cuvellier 1994). Grenvillian marbles from the Bancroft Terrane in eastern Canada, deposited at 1250 Ma (Rivers 1997), have a $^{87}Sr/^{86}Sr$ ratio of *c.* 0.7055 and marine carbonates from the 1200 Ma Society Cliffs Formation in the Arctic region (Baffin Island) have a minimum ratio of *c.* 0.706 (Veizer & Compston 1976; Kah *et al.* 2001). Although not as well dated, recent results from shallow-marine successions in Siberia and the Urals suggest that seawater Sr isotope ratios were in the range of 0.7060–0.7065 at *c.* 1300 Ma and then fell to *c.* 0.7050 by 1030 Ma (Bartley *et al.* 2001). If the Adrar carbonates in Mauritania are representative of seawater at 900 Ma (*c.* 0.707; Veizer *et al.* 1983), then an increase in the Sr isotopic ratio of marine carbonates between 1030 and 900 Ma may record the last stages in the formation of Rodinia. Why the earlier stages (1300–1000 Ma) in the formation of this supercontinent are not reflected in the Sr isotopes is not clear – perhaps pre-1000 Ma collisions were relatively minor.

The $^{87}Sr/^{86}Sr$ ratio decreases in seawater from *c.* 0.7074 at 900 Ma to a minimum of 0.706 at 850–775 Ma (Fig. 6; Jacobsen & Kaufman 1999; Walter *et al.* 2000). This dramatic decrease probably records the initial breakup of Rodinia (Fig. 3), with increased input of mantle Sr accompanying the breakup. This broad minimum is followed by a small but sharp increase in radiogenic Sr, levelling off between *c.* 700 and 600 Ma. This increase in the $^{87}Sr/^{86}Sr$ ratio may reflect some of the early plate collisions in the Arabian–Nubian Shield and elsewhere. The most significant change in the Sr isotopic ratio of Neoproterozoic seawater occurs between 600 and 500 Ma, when the $^{87}Sr/^{86}Sr$ ratio rises to nearly 0.7095 in only 100 Ma (Fig. 6). Although there is scatter in the data at this time, any fit of a curve through the data results in a rapid increase in the $^{87}Sr/^{86}Sr$ ratio. This rapid increase corresponds to the Pan-African collisions leading to the formation of the Early Palaeozoic supercontinent Gondwana. As collisions occurred, land areas were elevated and a greater proportion of continental Sr was transported into the oceans. The levelling or slight decrease in seawater Sr isotopic composition between 700 and 600 Ma is problematic, but could reflect enhanced mantle Sr input from the rifting of Siberia and Alaska–Chukotka from Laurentia (Figs 3 & 5; Kirschvink *et al.* 1997). Neither of the Neoprotero-

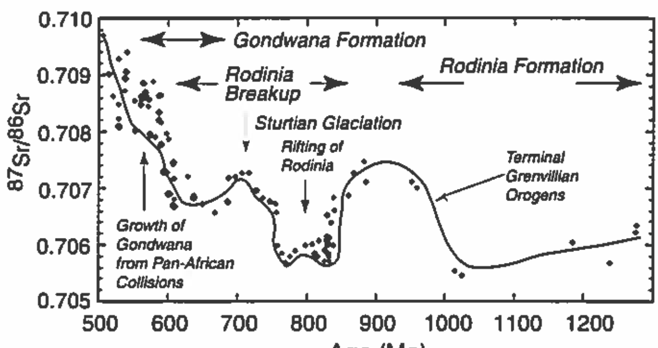

Fig. 6. Distribution of the $^{87}Sr/^{86}Sr$ ratio in seawater from 1000 to 400 Ma. Points represent published data from the least diagenetically altered marine limestones; the curve is a visual fit of the data. Principal references: Veizer & Compston (1976), Jacobsen & Kaufmann (1999) and Walter *et al.* (2000); other references given in text.

zoic glaciations are well represented in the Sr isotope record, a feature that may be due inadequate resolution of ages and lack of data for these glaciations.

A well-established minimum in the $^{87}Sr/^{86}Sr$ ratio at *c*. 470 Ma (Veizer *et al.* 1999) may reflect a superplume event at this time. Numerous other features, such as a period of no reversals in the Earth's magnetic field, a peak in black shale abundance and a maximum in sea level at this time, are also consistent with such a superplume event (Condie 2001*b*).

Juvenile continental crust (1350–500 Ma)

Condie (2001*a*) has shown that only 7–13% of the present continental crust formed during the 1350–900 Ma time window. As shown in Figure 1, most of this crust occurs in the long collisional Grenvillian Orogen, extending from Siberia along the eastern and southern margins of Laurentia into Australia and Antarctica. The largest volume, which is not well constrained in age or geographic distribution, is that in East Antarctica. A small but significant volume of juvenile crust also probably occurs in the triple-junction area of the Kalahari–Australia–Laurentia Plates. Almost all of the juvenile crust in the Grenville collisional belt appears to have formed between 1350 and 1200 Ma. The San Ignacio Orogen in South America, including small occurrences in Oaxaquia, Rio de la Plata and Rockall, contains up to *c*. 2% of the total juvenile crust, formed between 1350 and 1100 Ma. Only very small amounts of Grenvillian crust are preserved in other areas such as East Africa, the Arctic region, Malaysia, South China, Amazonia and Australia. Most of this crust also formed between 1300 and 1100 Ma, and only in the Arctic region (East Greenland, Svalbard, Ellesmere Island) is 1100–1000 Ma juvenile crust possibly present. In addition, a small volume of largely arc crust produced between 900 and 850 Ma occurs in the Arabian–Nubian Shield (Table 1).

A summary of juvenile crust produced in orogens between *c*. 800 and 500 Ma is included in Table 1 and shown on a Pannotia–Gondwana reconstruction in Figure 7. Estimates of the volume of juvenile crust are based on published U-Pb zircon ages and Nd isotopic data, as described in Condie (2001*a*). The amount of plume-related magma that may underplate the continental crust is not included in these estimates. However, because the Grenvillian and Pan-African Orogens expose varying crustal levels, from a few kilometres to nearly 30 km deep in a few cases, the estimates are thought to represent juvenile components in the upper and middle part of the continental crust in Neoproterozoic Orogens. At least 50% of the known juvenile crust of this age occurs in the Arabian–Nubian Shield, and in Avalonia–Cadomia and related terranes that formed along the border of Amazonia and West Africa (Fig. 7; Abdelsalam & Stern 1996). Another 20% is found in Pan-African orogens in Brazil, and 8% each in the Adamastor Orogen (bordering the South Atlantic today) and the circum-West Africa Orogens. The remaining 25% occurs as tectonic slices in such Neoproterozoic orogens as Taimyr, western Siberia and East Africa, and in Phanerozoic collisional orogens.

Tectonic settings of juvenile crust can be constrained using lithologic assemblages and geochemical characteristics of basalts (Condie 1997). From available data, as summarized in Table 1, it would appear that in most instances arc assemblages greatly dominate the exposed Neoproterozoic juvenile crust. The large volume of granitoids associated with most of these assemblages further suggests that continental margin arcs are most important. Data from Grenvillian orogens (1350–900 Ma) result in a similar conclusion (Condie 2001*a*). Although ophiolites are generally minor in Neoproterozoic orogens, a notable exception is the Adamastor Orogen and associated orogens (Table 1, row 3) in southern Africa, where mid-ocean ridge basalt (MORB) may comprise a significant proportion of the juvenile crustal assemblage. These orogens appear to have formed as continental rifts, which opened just enough to form small ocean basins before closing, trapping some of the oceanic crust.

Within-plate basalts (WPB) are also generally a minor component in most Neoproterozoic crust, averaging a few per cent of the total juvenile component (Table 1). The Katanga–Lufilian, the Arabian–Nubian Shield (900–850 Ma additions only) and Dalradian orogens are unique in that they contain *c*. 50% of WPB, chiefly oceanic plateau or continental flood basalts (Klemenic *et al.* 1985; Berhe 1990; Abdelsalam & Stern 1996; Alene *et al.* 2000). In the Arabian–Nubian Shield, an oceanic plateau, now represented by remnants as the Baish Group (*c*. 850 Ma) in Saudi Arabia, may have been the starting point for continental crust in this area (Stein & Goldstein 1996). The only ocean basin closures that captured significant volumes of oceanic crust are those of the Adamastor and Khomas Basins, now preserved in the Kaoko, West Congo, Damara and Gariep–Saldanian Orogens in southwestern Africa (Table 1). Geochemical data from basalts in the Gariep Orogen in Namibia indicate a within-plate tectonic setting for all of the mafic igneous units (Frimmel *et al.* 1996*b*): this includes bimodal volcanism during continental rifting, and seamounts and volcanic islands associated with a mantle plume in the Adamastor Ocean.

Cadomia, Avalonia and related terranes scattered along the eastern margin of Laurentia and through-

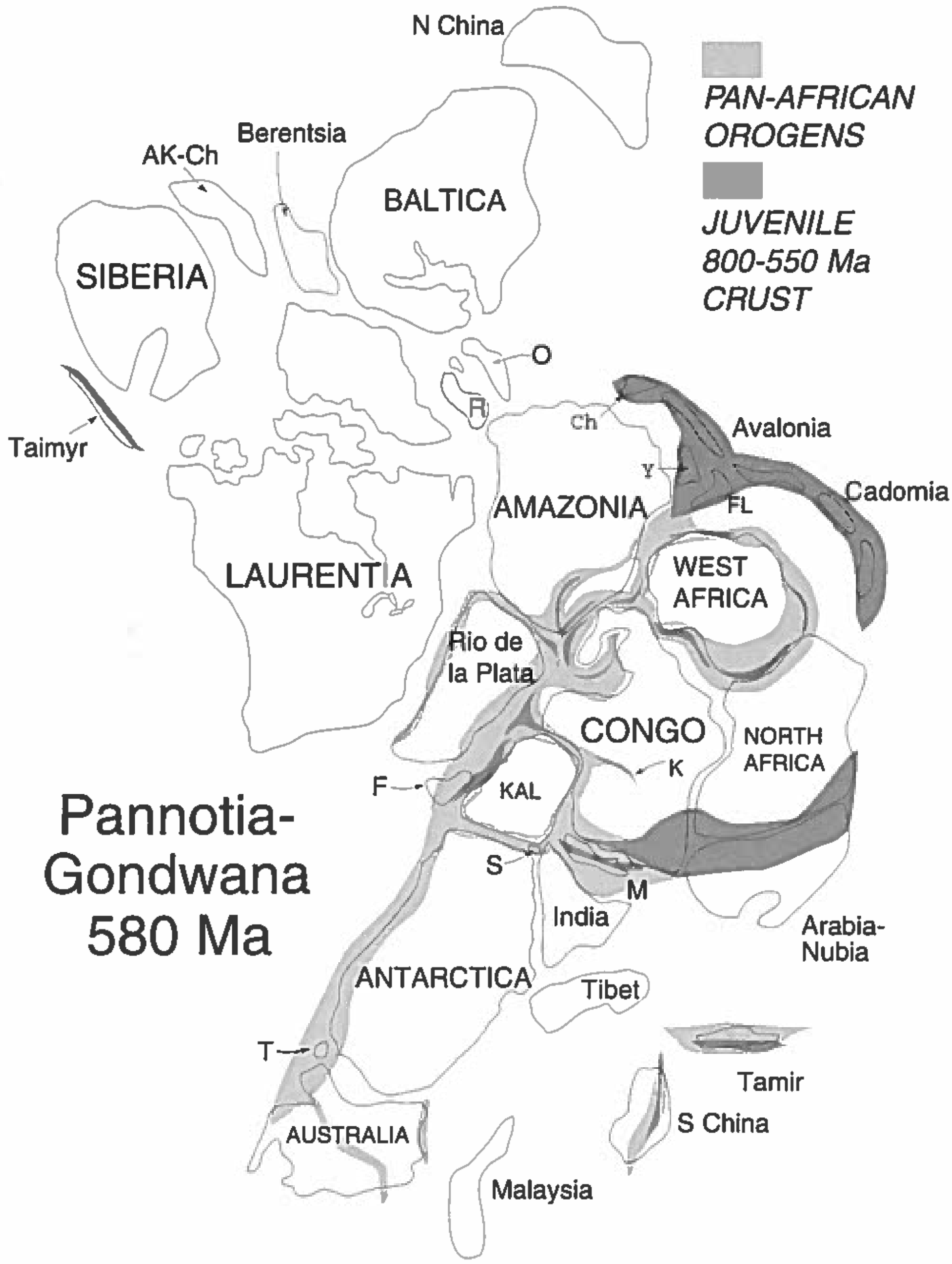

Fig. 7. Diagrammatic reconstruction of Pannotia–Gondwana at 580 Ma showing possible distribution of known juvenile crust formed between 800 and 550 Ma in relation to Pan-African orogens. T, Tasmania; other symbols defined in Figures 1 & 2.

out central Europe are comprised chiefly of juvenile continental margin arcs, and contain up to 10–20% of WPB, chiefly continental flood basalts (Dostal *et al.* 1990; Nance *et al.* 1991; Barr & White 1996; Liegeois *et al.* 1996; Murphy *et al.* 1999). Because of possible crustal contamination, the WPB component in these terranes may not represent entirely juvenile additions to the crust. Three main phases are recognized in the evolution of Avalonia (Murphy *et al.* 1999): (1) an early arc phase (780–660 Ma) during which continental margin arcs developed; (2) the main arc phase (630–590 Ma) when large volumes of juvenile arc crust formed; and (3) a transition stage during which compressive deformation changed to transcurrent deformation as large transform faults developed. Most of the juvenile crust preserved in Cadomia and Avalonia formed between 700 and 600 Ma (Table 1).

Rodinia and crustal growth rate

The minimal areal distribution of juvenile continental crust produced during the lifetime of Rodinia is shown in Figure 8, on an equal-area projection of the continents. Estimates of the area of this crust were scaled from maps in published papers and corrected for reworked components using Nd isotopic data (Table 2, column 4). Results are extrapolated to geographic regions where data are not available. The estimated percentages of juvenile crust relative to the present volume of continental crust are given in column 6 of Table 2. In terms of understanding crustal growth during supercontinent evolution, it is useful to divide juvenile crust into two types, based on the type of orogen in which it is preserved. Peripheral (or accretionary) orogens, which are largely comprised of accreted arcs, ophiolites and oceanic plateaus, occur around the edges of supercontinents; examples for Rodinia are the Antarctica segment of the Grenville Orogen at 1100–900 Ma and the Arabian–Nubian and Cadomia–Avalonia Orogens at 650–550 Ma (Figs 1 & 7). Internal orogens, which form between cratons that have collided, comprise chiefly of reworked older crust (Windley 1992); most of the Grenville and Pan-African orogens fall into this category. If the juvenile crust is divided into two age groups, terranes with ages of 1350–800 Ma occupy *c.* $10.1 \times 10^6 km^2$ (75%) and terranes with ages of 800–550 Ma comprise *c.* $4.2 \times 10^6 km^2$ (25%) of the total Neoproterozoic crust (Table 2, column 4). Areal distributions grouped according to orogen type suggest that, in both age categories, *c.* 50% of the juvenile crust occurs in peripheral orogens and 50% in internal orogens (Table 2, columns 2 and 3).

Because tectonic slices of Neoproterozoic rocks in younger orogens and underplated juvenile crust are not included in the estimates of juvenile crust, 15% has been added to each of the area estimates in Table 2 (column 5). Based on geologic maps of Phanerozoic orogens in Laurentia and Europe, this value is probably an upper limit for the amount of Neoproterozoic juvenile continental crust trapped in younger orogens. Results suggest that Neoproterozoic juvenile crust comprises *c.* 11% of the total continental crust: 8% in the 1350–800 Ma category and 3% in the 800–550 Ma category (Table 2, column 6). The earlier figure of 7–13% for the 1350–800 Ma category (Condie 2001*a*) has been refined based on more data. For an average crustal thickness of 35 km, these data indicate a growth rate of 0.70–0.74 km^3/a (Table 2, column 8), or, if an additional 10 km is restored due to erosion, 0.90–0.95 km^3/a. These values are near or somewhat below the average growth rate of continents during the Phanerozoic of *c.* 1 km^3/a (Reymer & Schubert 1984). It is noteworthy that continental growth rate during aggregation (1350–800 Ma) and breakup (800–550 Ma) phases of Rodinia are similar.

Assuming that the Arabian–Nubian Shield extends to the Zagros Suture in Iran and comprises Neoproterozoic juvenile crust (35 km thick), Reymer & Schubert (1986) calculated a crustal growth rate of 0.78 km^3/a, anomalously high for such a small volume of crust; similar rates were estimated by Stein & Goldstein (1996). Using only the exposed area of the Arabian Shield ($0.6 \times 10^6 km^2$), Pallister *et al.* (1990) estimate the rate to be only 0.07 km^3/a. Our rate, based on an area of $0.92 \times 10^6 km^2$ and a crust thickness of 41 km [to be comparable to the results of Pallister *et al.* (1990)] is 0.13 km^3/a, about twice the rate estimated by Pallister *et al.* (1990) but considerably less than that estimated by Reymer & Schubert (1986). If the Palaeoproterozoic and Late Archean crust exposed in southern end of the Arabian Peninsula (Windley *et al.* 1996; Whitehouse *et al.* 1998) underlies much of the Arabian Plate, as seems probable, Reymer & Schubert(1986) and Stein & Goldstein (1996) overestimated the volume of Neoproterozoic crust, thus accounting for their anomalously high growth rates.

Rodinia and superplume events

Unlike supercontinents in the Late Archean and Palaeoproterozoic, which were associated with the production and preservation of large volumes of new continental crust (Condie 1998, 2000), the results of this study suggest that Rodinia was not accompanied by the production of significant volumes of juvenile continental crust. In fact, the levels of crustal production are over an order of magnitude lower than those of two older supercontinents, and perhaps lower than Phanerozoic crustal production levels of *c.* 1 km^3/a. Why should Rodinia be different from older supercontinents? Perhaps the answer is the absence of a superplume event. Superplume events, which are short-lived, Earth-wide events ($<$100 Ma duration), during which large mantle plumes rapidly bombard the base of the lithosphere, may be responsible for the production of large volumes of juvenile crust. Both of the earlier supercontinents appear to have been associated with major superplume events at 2.7 and 1.9 Ga, and perhaps minor events at 2.5 and 1.7 Ga (Condie 1998; Isley & Abbott 1999).

In the case of a Late Archean supercontinent, it

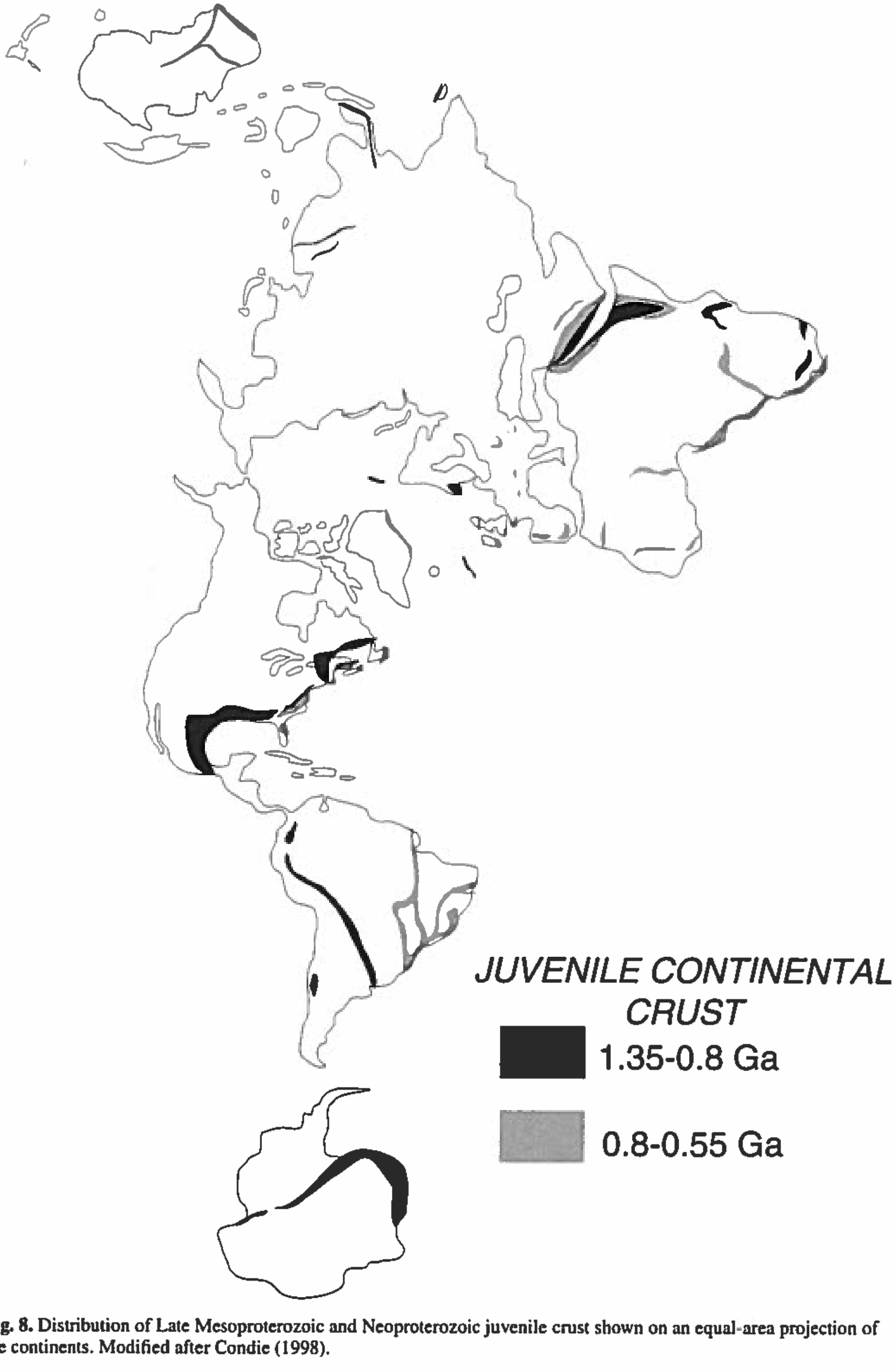

Fig. 8. Distribution of Late Mesoproterozoic and Neoproterozoic juvenile crust shown on an equal-area projection of the continents. Modified after Condie (1998).

Table 2. *Growth rate of juvenile crust in Rodinia*

Age (Ma)	Peripheral Orogen ($\times 10^6 km^2$)*	Internal Orogen ($\times 10^6 km^2$)*	Total ($\times 10^6 km^2$)	Total +15%†	Percentage‡	Growth Rate (km^2/a)§	Growth Rate (km^3/a)‖
1350–800	5.17	4.94	10.1	11.6	7.9	0.021	0.74 (0.95)
800–550	2.20	2.02	4.22	4.9	3.3	0.020	0.70 (0.90)

* Areas scaled from geologic maps and corrected for reworked crust with Nd isotope data.
† Fifteen per cent added for tectonic slices in younger orogens.
‡ Percentage of the total continental crust above sea level ($147 \times 10^6 km^2$).
§ Based on 550 Ma for the 1350–800 Ma crust and 250 Ma for the 800–550 Ma crust.
‖ Assumed crustal thickness of 35 km (45 km thickness given in parenthesis).

would appear that breakup of the supercontinent at 2.2–2.1 Ga may have triggered a 1.9 Ga superplume event (Larson 1991; Maruyama 1994; Condie 1998). If supercontinent breakup is required to produce a superplume event, perhaps by initiating catastrophic collapse of lithospheric slabs at the 660 km seismic discontinuity, the absence of a Mesoproterozoic–Neoproterozoic superplume event may mean that a Palaeoproterozoic supercontinent did not fully fragment before Rodinia began to form (Condie 2001c). Rogers (1996) suggested that two supercontinents formed at 1.9 Ga: the first, which he called Atlantica, comprises Amazonia and other Archean Cratons in South America, and the Congo, West Africa and North Africa Cratons in Africa; Nena, the second supercontinent, includes Laurentia, Baltica, Siberia and Antarctica. These two supercontinents may have survived breakup of a Palaeoproterozoic supercontinent at 1.6–1.4 Ga, and they were later incorporated intact into Rodinia (Condie 2001*b*). Why two supercontinents survived a Mesoproterozoic breakup is unknown. An alternative interpretation is that that two supercontinents formed at 1.9 Ga, and that neither was large enough to provide adequate lithospheric shielding for the production of mantle upwellings large enough to break the continental lithosphere (Lowman & Jarvis 1999).

Because superplume events may have been triggered by the breakup of a Late Archean supercontinent, when Rodinia broke up at 850–600 Ma, it may also have triggered a superplume event. Is there any evidence for such an event? The recent recognition of a superchron in the Ordovician centred at *c.* 480 Ma presents the possibility of a superplume event at this time (Johnson *et al.* 1995). Consistent with such a superplume event is a peak in eustatic sea level and in a calculated production rate of oceanic crust (Condie 2001*b*). Black shales are also widespread in the Ordovician. A steep fall in the $^{87}Sr/^{86}Sr$ ratio of seawater at this time could be a response to enhanced input of mantle Sr accompanying such a superplume event.

Unlike Rodinia and Pangaea, which had lifetimes of *c.* 100–150 Ma (Rodinia from 900 to 800 Ma and Pangaea from 320 to 180 Ma), Gondwana appears to have survived for 370 Ma, from *c.* 550 to 180 Ma (Fig. 9). The Late Archean and Palaeoproterozoic supercontinents may have survived for even longer time periods. Why did Rodinia and Pangaea begin to fragment so soon after their formation? Perhaps it had to do with their relatively large sizes. As shown by the numerical models of Lowman & Jarvis (1999) and Lowman & Gable (1999), a plate carrying a supercontinent must be very large to effectively shield the underlying mantle from cooling by subduction. Gondwana, which contained only about half of the continental crust at 500 Ma, may not have been large enough to provide this shielding, and hence it survived until it became part of Pangaea. Likewise, the pre-Rodinian supercontinents contained only 40–80% of the present continental crust and thus may have required a longer period of time to effectively shield the underlying mantle to produce a mantle upwelling.

Conclusions

(1) Thick Archean lithosphere in Laurentia, Siberia and Baltica may have resulted in the Neoproterozoic supercontinent Rodinia splitting around the margins rather than through the centre during the 850–600 Ma breakup.

(2) Opening of the Iapetus Ocean at 620–550 Ma occurred at the same time as collisions between East and West Gondwana. These

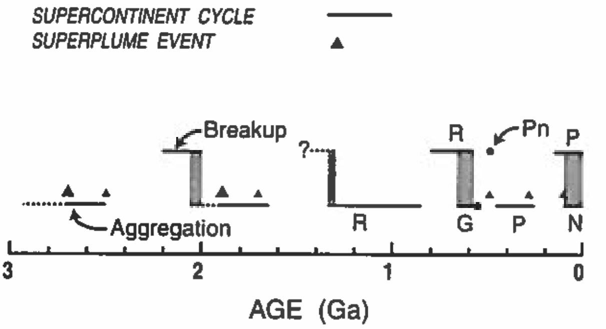

Fig. 9. The supercontinent cycle shown in relation to possible superplume events. R, Rodinia; G, Gondwana; Pn, Pannotia; P, Pangaea; N, possible new supercontinent. Size of triangles proportional to size of superplume events. Vertical grey bars are areas of assembly/breakup overlap. Modified after Condie (1998).

simultaneous events may have been responsible for closing small ocean basins in West Gondwana, producing the Pan-African and Braziliano orogenies.

(3) The Sr isotope record of seawater records the early fragmentation of Rodinia at 850–750 Ma and the formation of Pannotia–Gondwana at 600–500 Ma; the record is less clear between 750 and 600 Ma, when both fragmentation and collision events occurred simultaneously.

(4) At least 50% of the known juvenile continental crust produced between 750 and 550 Ma occurs in the Arabian–Nubian Shield and in other terranes that formed along the northern border of Amazonia and West Africa. An additional 20% is found in the Pan-African orogens of Amazonia, and *c.* 16% in the Adamastor and West African Orogens.

(5) Most juvenile continental crust produced during the 1350–500 Ma time interval formed in arcs, chiefly continental margin arcs. Only *c.* 11% of the total preserved continental crust formed during this period.

(6) About 50% of the Neoproterozoic juvenile crust occurs in peripheral orogens and 50% in internal orogens. In addition, an unknown but probably small amount of juvenile crust may have been added from mantle plumes by crustal underplating.

(7) Continental growth rates during the time period of 1350–500 Ma were similar or less than the average rate of continental growth during the Phanerozoic of 1 km^3/a. These small growth rates apply to both formation and breakup stages of supercontinents, and thus continental crustal growth is not dependent upon the stage of the supercontinent cycle.

(8) The low rates of continental growth during the Neoproterozoic may be due to the absence of superplume events associated with the formation of Rodinia or Gondwana. The breakup of Rodinia between 800 and 600 Ma, however, may have triggered a superplume event at *c.* 480 Ma.

(9) If supercontinent breakup is required to produce a superplume event, perhaps by initiating catastrophic collapse of lithospheric slabs at the 660 km seismic discontinuity, the absence of a Mesoproterozoic– Neoproterozoic superplume event may mean that a Palaeoproterozoic supercontinent did not fully fragment.

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