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Neoproterozoic crustal growth: The solid Earth system during a critical episode of Earth history

Robert J. Stern *

Geosciences Department, University of Texas at Dallas, Richardson TX 75083-0688, USA

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Abstract

The behavior of the solid Earth system is often overlooked when the causes of major Neoproteozoic (1000–542 Ma) climate and biosphere events are discussed although ~20% of the present continental crust formed or was remobilized during this time. Processes responsible for forming and deforming the continental crust during Neoproterozoic time were similar to those of the modern Earth and took place mostly but not entirely at convergent margin settings. Crustal growth and reworking occurred within the context of a supercontinent cycle, from breakup of Rodinia beginning ~830 Ma to formation of a new supercontinent Greater Gondwana or Pannotia, ~600 Ma. Neoproterozoic crust formation and deformation was heterogeneous in space and time, and was concentrated in Africa, Eurasia, and South America during the last 300 million years of Neoproterozoic time. In contrast, the solid Earth system was relatively quiescent during the Tonian period (1000–850 Ma). The vigor of Cryogenian and Ediacaran tectonic and magmatic processes and the similar timing of these events and development of Neoproterozoic glaciations and metazoa suggest that climate change and perhaps increasing biological complexity was strongly affected by the solid Earth system.

Keywords: Neoproterozoic; Continental crust; Ophiolites; Supercontinent cycle

1. Introduction

The Neoproterozoic Era (Fig. 1) is when Earth and life on it began to appear modern. This was when the first complex animals developed, ice ages became common, and the atmosphere became rich with oxygen. The great Pacific Ocean basin formed, alongside the Atlantic Ocean's ancestor Iapetus. Some of the diagnostic petrotectonic assemblages that are associated with modern-style plate tectonics first appear (blueschists, ultra-high pressure metamorphic assemblages) or become common (ophiolites) at this time (Stern, 2005). Major changes in Earth's biosphere and climate system at least partly reflected activity in the solid Earth system, but presently these relationships are only dimly discerned (Fairchild and Kennedy, 2007). Earth witnessed a single cycle of supercontinent breakup (Rodinia) and reformation (Greater Gondwana or Pannotia) during the Neoproterozoic (Fig. 1), but it is not clear how this affected climate and the biosphere. One possibility is that increased explosive volcanism during cooled climate sufficiently to cause glaciation (e.g. Stern et al., in press).

The purpose of this review is to focus attention on the significance of continental crust that formed in Neoproterozoic crust. This is motivated by the recognition that major crust-forming episodes are when the solid Earth system most directly impacts the biosphere, hydrosphere, and climate systems. How changes in the solid Earth affected climate and biological system during Neoproterozoic Era cannot be appreciated until this time's tectonic and magmatic activity is better known. The nature and timing of these interactions must be better resolved if we are to understand the complete Neoproterozoic Earth System. Fortunately, rapid advances in our understanding of Neoproterozoic events are occurring as a result of geochronologic breakthroughs, especially ion-probe dating of zircons. This has allowed Neoproterozoic time to be subdivided into three periods: Tonian (1000–850 Ma), Cryogenian (850–635 Ma), and Ediacaran (635–542 Ma) (Knoll, 2000; Knoll et al., 2006). Understanding is further advanced because rocks

* Tel.:+1 972 883 2442; fax: +1 972 883 2537.
E-mail address: rjstern@utdallas.edu.

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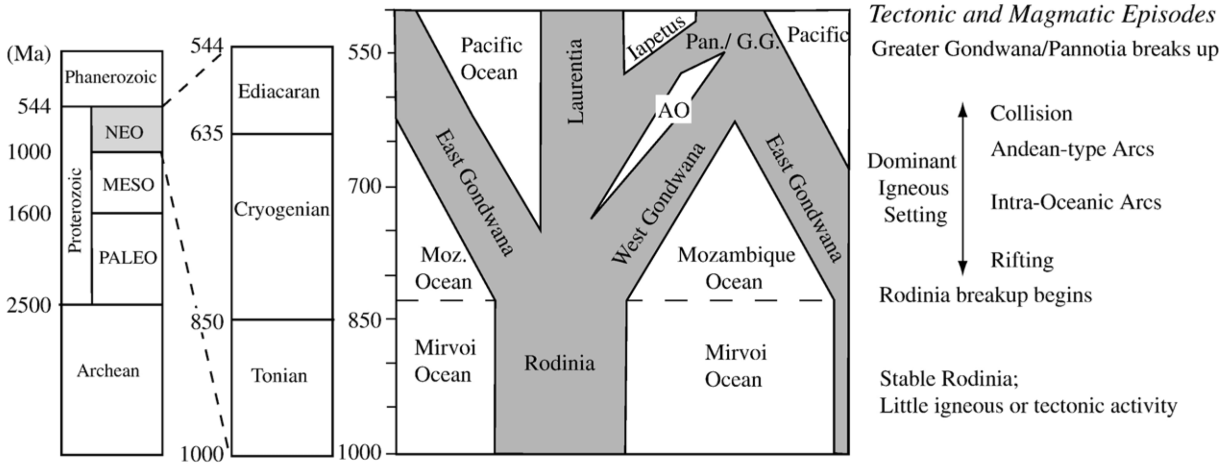

Fig. 1. Subdivisions of Precambrian (left) and Neoproterozoic time (center), from Knoll (2000). Also shown is schematic "tree diagram" representation (modified after Cawood 2005) of the Neoproterozoic transition from Rodinia into Gondwana through closure of the Mirvoi-Mozambique and Adamaster oceans and the opening of the Pacific and Iapetus oceans (right). Note that the left and right sides of the tree diagram are the same. Far right outlines some of the important Neoproterozoic tectonic events. AO = Adamaster Ocean; Pan./G.G. = Pannotia/ Greater Gondwana.

of Neoproterozoic age are globally abundant, as shown by Stewart (2007) presents a map that summarizes global abundance of exposed post-870 Ma Neoproterozoic rocks: sediments, igneous rocks, and metamorphic rocks (http://pubs.usgs.gov/of/2007/1087/). This further encourages a broadly international as well as interdisciplinary geoscientific study of Neoproterozoic Earth history.

There is increasing evidence that the Neoproterozoic was an important time of crustal growth. Goodwin (1991) estimated that ~17% of the present continental crust is Neoproterozoic (Table 2). Maruyama and Liou (1998) suggested that perhaps 20% of the area of all orogenic belts formed between 0.7 and 0.6 Ga. Table 4 of Artemieva (2006) splits late Precambrian time between two informal time periods, 750–550 Ma (11.7% of continental area) and 1.15–750 Ma (9.4%), so at least a tenth and as much as a fifth of crust is Neoproterozoic. In spite of these independent assessments, the significance of Neoproterozoic crustal growth is often underestimated (Condie, 2000).

There are several reasons why the volume of Neoproterozoic crust might be underestimated. First, this crust is found in fragments of former Gondwanaland or in Asia, regions where our understanding of crustal evolution lags behind that for N. America, and Europe. Geographic bias may also help explain why most reconstructions of mantle plume or hotspot activity through time seem to overlook important Neoproterozoic pulses in W. North America, China and Australia (Abbott and Isley, 2002). Our understanding of these less well-known regions is advancing rapidly, as modern geochronologic and isotopic techniques are increasingly used to recognize significant tracts of Neoproterozoic rocks. A second reason is that Neoproterozoic crust is often buried beneath sediments. This is largely due to the increase in subcontinental mantle lithosphere density, from Archean to Phanerozoic time (Artemieva and Mooney, 2001). Because the early Earth was hotter, melting of mantle to form continental crust resulted in greater depletion of especially Fe in the subjacent mantle lithosphere and made this lithosphere less dense than younger lithosphere. Mantle lithosphere beneath Neoproterozoic crust is less depleted in Fe and thus generally denser than that beneath Archean and Paleoproterozoic cratons. Isostatic considerations indicate that Neoproterozoic crust thus is more likely to be buried beneath Phanerozoic sediments. Third, Neoproterozoic crust, particularly in Eurasia, is commonly involved in younger orogenic events, and these unfossiliferous units are often mapped with associated younger units that do contain fossils. The age of the earlier crust-forming event is only revealed after careful geochronologic study (Neubauer, 2002). Fourth, Neoproterozoic structures are preferentially exploited by rifts (McConnell, 1972; Nyblade and Brazier, 2002), again burying Neoproterozoic crust beneath sediments after rift flanks subside.

In this review, we consider both juvenile Neoproterozoic crust (JNPC) and older crust that was melted, metamorphosed or otherwise reworked during the Neoproterozoic (MORN: note that Table 1 summarizes all acronyms used in this paper). We assume that JNPC is associated with a significant proportion of mafic and ultramafic igneous rocks (including ophiolites), that it has mantle-like isotopic compositions (e.g., ε-Nd > 0, $^{87}Sr/^{86}Sr < 0.705$, etc.), and contains zircons with U–Pb ages that are rarely older than Neoproterozoic. In contrast, MORN lacks ophiolites, has isotopic compositions appropriate for pre-Neoproterozoic continental crust, and has abundant older zircons. It should be emphasized that MORN and JNPC are endmember concepts, because in many situations the rejuvenation of older crust was associated with significant additions of mantle-derived (mafic) melts, but the concepts are useful for summarizing global Neoproterozoic crust. To a certain extent, JNPC and MORN are equivalents for exterior and interior orogens of Murphy and Nance (1991) or for the Pacific-type and Alpine-type orogens of Ernst (2005) respectively, although some JNPC forms in interior/Alpine-type orogens and some MORN forms in exterior/Pacific-type orogens.

The distinction between JNPC and MORN may not be very important in the context of how solid Earth processes – especially igneous activity – affect the biosphere and surficial

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Table 1
Acronyms used in this paper

ANS	Arabian–Nubian Shield
CAFB	Central African Fold Belt
CAOB	Central Asian Orogenic Belt
EAO	East African Orogen
EAAO	East African-Antarctic Orogen
GIS	Geographic Information System
IGCP	International Geologic Correlation Project, co-operative effort between UNESCO and IUGS
JNPC	Juvenile (mantle-derived) Neoproterozoic crust
LATEA	Central Hoggar (Algeria) composite terrane (Laouni, Azrou-n-Fad, Tefedest, Egéré-Aleksod)
LIP	Large igneous province
LREE	Light rare earth elements
MB	Mozambique Belt
MORB	Mid-ocean ridge basalt
MORN	Crust that was melted, metamorphosed, or otherwise reworked during Neoproterozoic time
OIB	Ocean island (intra-plate, hotspot) basalt
OJP	Ontong-Java Plateau
PNZ	Port Nolloth zone, Gariep belt
SHRIMP	Sensitive, high-resolution ion microprobe, used for obtaining U–Pb ages of zircons and other minerals.
SM	Saharan metacraton
SSZ	Super-subduction zone (usually pertains to ophiolite)
SWEAT	Hypothesis that the SW USA and East Antarctica were once a single block
UHP	Ultra-high pressure, metamorphism at $P>2.7$ GPa
VRM	Volcanic rifted margin
WAC	West African Craton

Earth systems. For example, explosive eruptions of volcanoes fed by melts extracted from the mantle or the crust may have similar effects on climate. It is probably preferable at this stage in our understanding of Neoproterozoic crust formation to minimize the differences between JNPC and MORN even as we recognize these endmembers.

In this review we do not address continental configurations and positions based on paleomagnetic data. These have been extensively discussed in the literature (Evans, 2000; Li et al., in press; Meert and Torsvik, 2003). The summary here is independent of, yet complementary to, these reconstructions.

2. Formation of continental crust today

Insights into how continental crust evolved in Neoproterozoic time can be gained by considering how it forms and is destroyed today (Fig. 2). Scholl and von Huene (in press) estimate that continental crust is presently produced at a rate of ~5.5 km^3/year. Crust is also destroyed, but this is more difficult to quantify. Scholl and von Huene (in press) estimate that ~3 km^3/year is lost, but this includes poorly constrained estimates for losses due to delamination and subduction of continental crust. There is also broad consensus that modern-style plate tectonics operated during Neoproterozoic time (Stern, 2007), so similar processes of crust formation are expected. With these caveats, it seems reasonable to assume that processes and rates were similar for modern and Neoproterozoic crustal growth.

Formation of continental crust today occurs mostly at Earth's ~45,000 km of convergent plate margins as one of the important products of the "Subduction Factory" (Davidson and Arculus, 2005; Tatsumi, 2005). New continental crust produced this way ultimately results from magmatic additions due to water-induced melting of mantle over subduction zones. Igneous rocks accumulate in the crust above long-lived subduction zones to produce magmatic arcs, at a global rate of a few cubic kilometers per year (Dimalanta et al., 2002; Scholl and von Huene in press). Magmatic arcs form on both oceanic lithosphere (intra-oceanic arcs) and continental lithosphere (Andean-type arcs). Because they are located at convergent plate boundaries, intra-oceanic arcs collide with other tracts of thickened crust to form increasingly large and differentiated "composite" arcs. These are recognized in ancient orogens as composite terranes, and are an important reason why the term "orogen" is largely synonymous with crustal growth. Subduction-related crustal growth also occurs at Andean-type continental margins, where pre-existing continental crust is modified at the same time that juvenile magmatic materials are added. Ancient Andean-type arcs are invariably deeply eroded and are recognized today as calc-alkaline batholiths and associated migmatitic sheaths (Hamilton and Myers, 1967; Ducea, 2001).

Convergent margins grow new crust principally by igneous activity, but accretion of sediments against the inner trench wall is also sometimes important. This occurs when sediments deposited on the subducting plate are transferred to – scraped off by – the overriding plate to form accretionary prisms. Accretionary prisms are mostly composed of recycled continental crust in the form of clastic sediments, shed from the overriding plate and thus are mostly MORN. Accretionary prisms are cemented to arcs when these collide to form accretionary orogens such as the western N. America cordillera (Dickinson, 2004). Most (~70%) arc-trench systems today lack accretionary prisms and are sites of tectonic erosion, where the inner trench wall is continuously being removed by subduction (von Huene and Scholl, 1991). Nearly all modern accretionary prisms are adjacent to Andean-type margins, where trenches are filled with sediments delivered by large rivers or glaciers (von

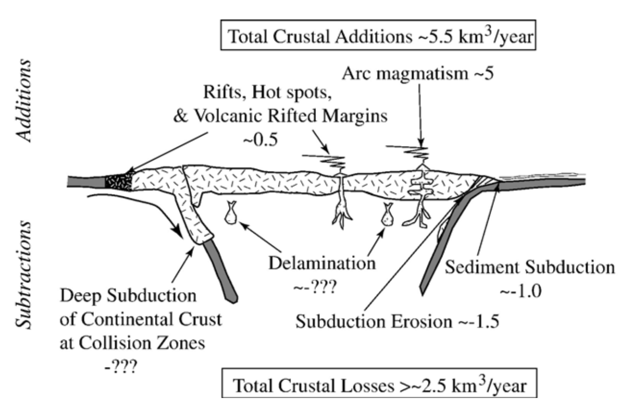

Fig. 2. Synopsis of how continental crust forms and is destroyed today, modified after Scholl and von Huene (2006). Also shown are estimated modern global additions to and subtractions from the crust.

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Huene and Scholl, 1991; Clift and Vannucchi, 2004). It seems likely that, by analogy with modern convergent margins, a similar minority of Neoproterozoic arc-trench systems were accretionary, and that these existed near continents.

New continental crust also forms where thick piles of oceanic crust – oceanic plateaux, formed above mantle "hotspots" – collide with arcs, such as the Miocene collision of the Ontong-Java Plateau with the Solomon arc (Mann and Taira, 2004). Continued subduction as a result of subduction polarity reversal following collision (Stern, 2004) resulted in continued igneous activity, which served to further hybridize and process OJP crust. This tectonic and magmatic mixing of arc and hotspot crust is another way to add juvenile continental crust, as inferred for the Canadian cordillera during Mesozoic time (Lapierre et al., 2003). Hotspot magmas also contribute directly to the continental crust by underplating, intrusion, anatexis and delamination, such as occurring beneath the Yellowstone hotspot today (Bryan et al., 2002) and has been associated with Large Igneous Provinces (LIPs) in the past; the Deccan Traps and Columbia River flood basalt province are examples. Finally, additions to the continental crust are made at magmatic rifts and at volcanic rifted margins (VRMs) when continents break up. Breakup is often associated with eruption of thick piles of mostly mafic lava at the continent–ocean crustal boundary (Menzies et al., 2002), which can ultimately become new additions to the continental crust. These additions are only significant during the rift-to-drift transition. Overall, hotspot additions to the continental crust today are estimated to be ~10% of arc additions (Scholl and von Huene, 2006), but this contribution was probably more important when large supercontinents rifted apart. Addition of mafic melts to existing continental crust at Andean-type margins and hotspots alongside collisional reworking are the most important mechanisms for producing MORN.

Magmatic additions to the continental crust at Andean-type convergent margins and at hotspots are immediate, but attaching composite arcs and VRMs requires that this material be attached to a continent, usually by collisional suturing. This happens either at accretionary "Pacific-type" or collisional "Alpine" type orogens (Ernst, 2005). Suture zones are often marked by ophiolites, which are best preserved where collision is of moderate intensity. Terminal collision between continents results in intense deformation and erosion that can largely remove ophiolitic suture markers, but even in the case of the severe India-Asia collision there are some ophiolitic fragments to mark the Indus suture (Corfield et al., 2001). Most Cenozoic ophiolites formed in forearcs when subduction began (Stern, 2004), and Neoproterozoic ophiolites probably have a similar significance.

Further processing of mafic JNPC must occur to yield true continental crust, which has an intermediate (andesitic) composition. The moho beneath active zones of crustal growth must be open to mass transfers in both directions. Mafic magmas move from the mantle up into the crust, where these spur felsic magmagenesis by differentiation as well as by melting lower crust (Brown, 2007). This yields granodioritic upper crust as well as residual and cumulate gabbroic lower crust (Rudnick and Gao, 2003; Hawkesworth and Kemp, 2006). Differentiation into felsic upper crust and mafic lower crust is an important aspect of crust formation, especially where magmatic refinement allows dense lower crustal materials to separate and sink into the mantle. Lower crustal delamination happens at all stages, from thickening of the growing intra-oceanic arc (Jull and Kelemen, 2001) to the point where terminal continental collision (Anderson, 2005), and can also occur after plate convergence ends (Ducea and Saleeby, 1998).

Both processes – mafic crust formation and intra-crustal refining – operate continuously in zones of active igneous activity to produce JNPC. Some dacitic juvenile crust that does not need further processing is generated by melting of young subducted slabs (Defant and Drummond, 1990), but these "adakitic" melts comprise a small fraction of juvenile crust production at arcs today and presumably during the Neoproterozoic as well. JNPC and MORN, along with their mantle lithosphere root, ultimately stabilize to become cratons (Flowers et al., 2004).

3. Continental crust formation and the Neoproterozoic supercontinent cycle

Modern plate tectonic processes produce JNPC and MORN within the context of a supercontinent cycle, and this probably was also true for Neoproterozoic crustal growth (Murphy and Nance, 2003). The Neoproterozoic witnessed the breakup of the supercontinent Rodinia and the reassembly of these fragments into a new supercontinent, known as Greater Gondwana (Stern, 1994) or Pannotia (Dalziel, 1997), by the end of the era (Fig. 1). It should be noted that the existence of the Rhodinia supercontinent and the configuration of continental fragments within it remains controversial (Torsvik, 2003), but the majority view that there was an end-Mesoproterozoic supercontinent is adopted here. Accepting a Neoproterozoic supercontinent cycle implies systematic changes in the relative contributions of igneous activity at the different sites discussed above (Fig. 3A–D) and a progressive variation in overall magma compositions (Fig. 3E). Rodinia remained intact during the Tonian (*tonos* is Greek for tension or stretching, as Rodinia experienced before it ruptured in Cryogenian time); consequently the first 150 Ma of Neoproterozoic time witnessed little crust formation. In contrast, the Cryogenian period was a period of intense magmatic activity, beginning with a Rodinia breakup ~830 Ma ago (Li et al., 1999; Torsvik, 2003). Li et al. (in press) infer that widespread rifting of Rodinia occurred between this time and 740 Ma, with episodic plume events at ~825 Ma, ~780 Ma and ~750 Ma.

Early Cryogenian igneous rocks related to Rodinia breakup are well preserved in western North America, China, Australia, and Siberia, leading to the inference that some of these that are now found on either side of the Pacific Ocean were adjacent before early Cryogenian rifting (Burret and Berry, 2000; Sears and Price, 2003; Wang and Li, 2003). This line of argument is explored in the SWEAT hypothesis (Moores, 1991) and its variations (e.g. Karlstrom et al., 1999; Burret and Berry, 2000; Wingate et al., 2002). Rodinia fragmentation required formation of new subduction zones, and Cryogenian magmatism reflected the increasing vigor of complementary rifts, VRMs, and island

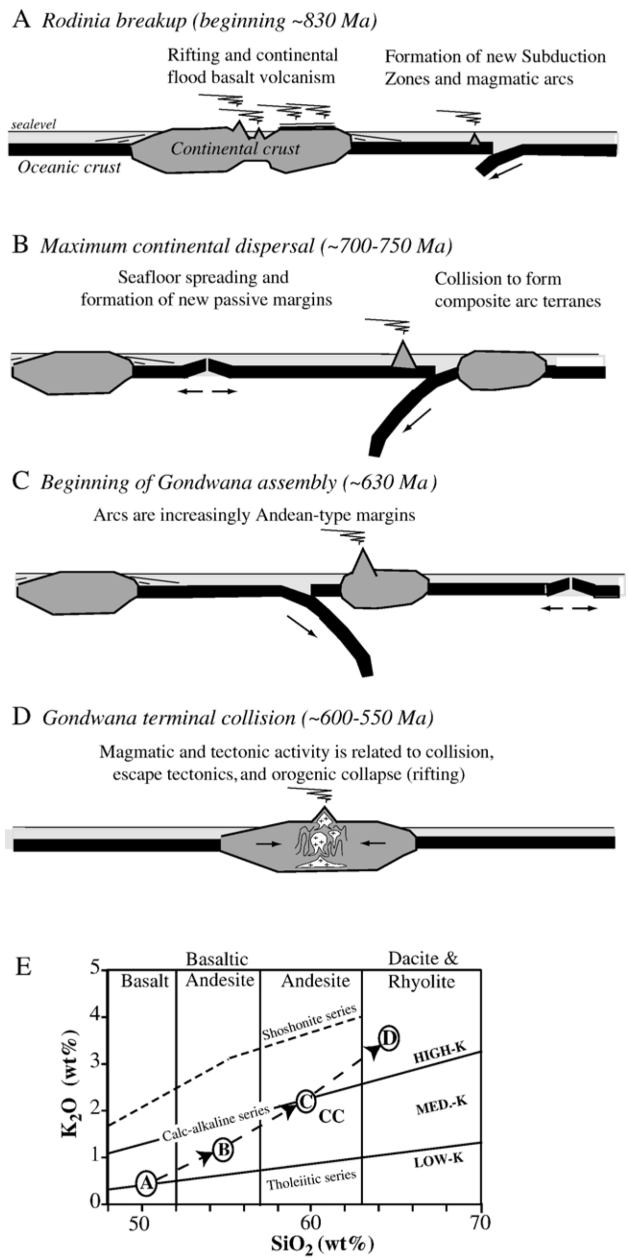

Fig. 3. A–D: Evolution of magmatic and other crust-forming processes over the Neoproterozoic supercontinent cycle, greatly simplified. Modified after Stern et al. (in press). E summarizes the expected progression in potassium and silica contents of predominant magmas over the four stages of the supercontinent cycle shown in panels A–D. "CC" indicates bulk composition of the continental crust, after Rudnick and Gao (2003).

arcs. Cryogenian subduction zones ultimately became sites of collision and new orogenic belts.

Breakup continued throughout the rest of Neoproterozoic time (Li et al., in press), even as the new supercontinent coalesced. The birth of the Pacific Ocean basin dates from Neoproterozoic time (Fig. 1; Cawood, 2005) and this hemispheric feature has persisted since. The Ediacaran period also witnessed considerable igneous activity and crust formation as the new supercontinent formed. This igneous activity occurred at old and new subduction zones, at collision zones, and where new rifts continued to form.

Over the Neoproterozoic supercontinent cycle, crustal growth at first was largely concentrated at rifts, with arcs and collision-related igneous activity becoming progressively more important. This progression can be inferred because maturing rifts evolve to become zones of seafloor spreading, and oceanic crust is a very minor contribution to the continental crust. In contrast, arc volcanoes build progressively thicker and more buoyant crust that is increasingly likely to be preserved in an orogen as JNPC. Convergent margin magmas are predominantly low-K and basaltic for intra-oceanic arcs but more potassic and felsic for Andean-type arcs (Stern, 2002b). Consequently, arc magmas evolved from predominantly mafic to increasingly felsic over the Neoproterozoic supercontinent cycle, as predominantly intra-oceanic arcs were progressively replaced as a result of collisions by dominantly felsic Andean-type arcs. Magmatic activity at the beginning of the supercontinent cycle thus would have been dominated by juvenile, intra-oceanic arcs and generation of low-K tholeiitic and medium-K calc-alkaline mafic magmas. Over the supercontinent cycle, juvenile arcs coalesced to produce thicker crust, and arc magmas evolved to more felsic and potassic compositions, as shown in Fig. 3E.

The Ediacaran period witnessed significant igneous activity at diverse tectonic settings, reflecting magmatic activity associated with collision and post-orogenic collapse. Such a transition is observed in the Arabian–Nubian Shield at ~600 Ma, when convergent margin magmatism was replaced by post-tectonic or anorogenic magmatism (e.g., Beyth et al., 1994); a similar transition is inferred for Neoproterozoic crust of eastern North America and western Europe (Nance et al., in press). The end-Neoproterozoic supercontinent started to break apart almost immediately, with transtension and rifting to form Iapetus on its northern and western margins (Nance et al., in press) along with formation of new subduction zones with Andean-type arcs along its southern margins (Fig. 1; Terra Australis Orogen of Cawood, 2005).

4. Limitations of orogenic nomenclature for correlating Neoproterozoic crustal growth

One of the challenges facing any global overview of tectonic and magmatic activity is the multiplicity of names for the events of different region, such as "Pan-African", "Brasiliano", "Timanian", "Baikalian", "Jiangnan", etc. These are useful in advancing studies of regions with protracted tectonic histories, for example in discussing basement geology of the British isles, where the Neoproterozoic Dalradian orogeny needs to be distinguished from the Paleozoic Caledonian orogeny. In spite of the utility of such terms for regional studies, the large number of such names often causes frustration when correlating Neoproterozoic crust-forming events around the world, particularly because of the often informal and poorly-defined meaning of these terms. An example is the term "Pan-African", which was originally developed by Kennedy (1964) to describe tectonothermal events in Africa at the end of the Precambrian and beginning of Paleozoic time (500 ± 50 Ma). The term is now commonly applied to crust formation and reworking in Africa anytime in the Neoproterozoic and Early Paleozoic. The term is sometimes applied to Neoproterozoic orogeny elsewhere in Gondwana, but is rarely applied to Neoproterozoic crust-forming episodes of similar age in N. America or Eurasia.

Another reason that orogenic nomenclature impedes global correlation of Neoproterozoic tectonic and magmatic episodes is that these focus exclusively on orogenic events. Orogenies are only one aspect of the supercontinent cycle and crust formation. Other aspects such as rifting or arc magmatic activity are just as important and encompass as much or more time but are ignored by the nomenclature-driven focus on orogenies. As a result, there is a strong temptation to expand the timespan and, implicitly, the definition of the orogeny to include all related tectonic and magmatic events. This problem is also illustrated by the Pan-African “event”, “orogeny”, or “orogenic cycle”, which is increasingly used to encompass all aspects of Neoproterozoic crust formation in Africa and surrounding parts of reconstructed Gondwana. The Pan-African as presently used is simply the most extensive, protracted, and best preserved of the great Neoproterozoic crust-forming episodes. This is why the term as presently applied encompasses tectonic and magmatic activity over a period of ~300 million years, which is an order of magnitude longer than well-defined Phanerozoic orogenies such as the Grampian (Dewey, 2005).

Use of names for regional tectonic and magmatic events of Neoproterozoic age is less important today because the proliferation of good quality U–Pb zircon ages along with geochemical and other indicators of tectonic setting allow much better resolution of timing and tectonic setting of igneous activity and metamorphism. Traditional names are nevertheless used below in describing Neoproterozoic crust formation, simply because these terms are deeply embedded in the pertinent geoscientific literature. Nevertheless, it seems that the effort to advance global synthesis of Neoproterozoic crustal growth benefits from stressing radiometric ages and inferred tectonic settings wherever possible.

5. Distribution of Neoproterozoic ophiolites

Location of JNPC is outlined by the distribution of Neoproterozoic ophiolites (Fig. 4). Ophiolites mark sites of former subduction zones and are emplaced when the subduction of buoyant crustal blocks (arcs, oceanic plateaus, or continents) is attempted and fails. This destroys the subduction zone and accretes the buoyant crust, with the ophiolite marking the suture. Ophiolites older than about 1.0 Ga are rare, but Neoproterozoic and younger ophiolites are abundant (Yakubchuk et al., 1994; Dilek, 2003; Stern, 2005; Dilek et al., 2007), found

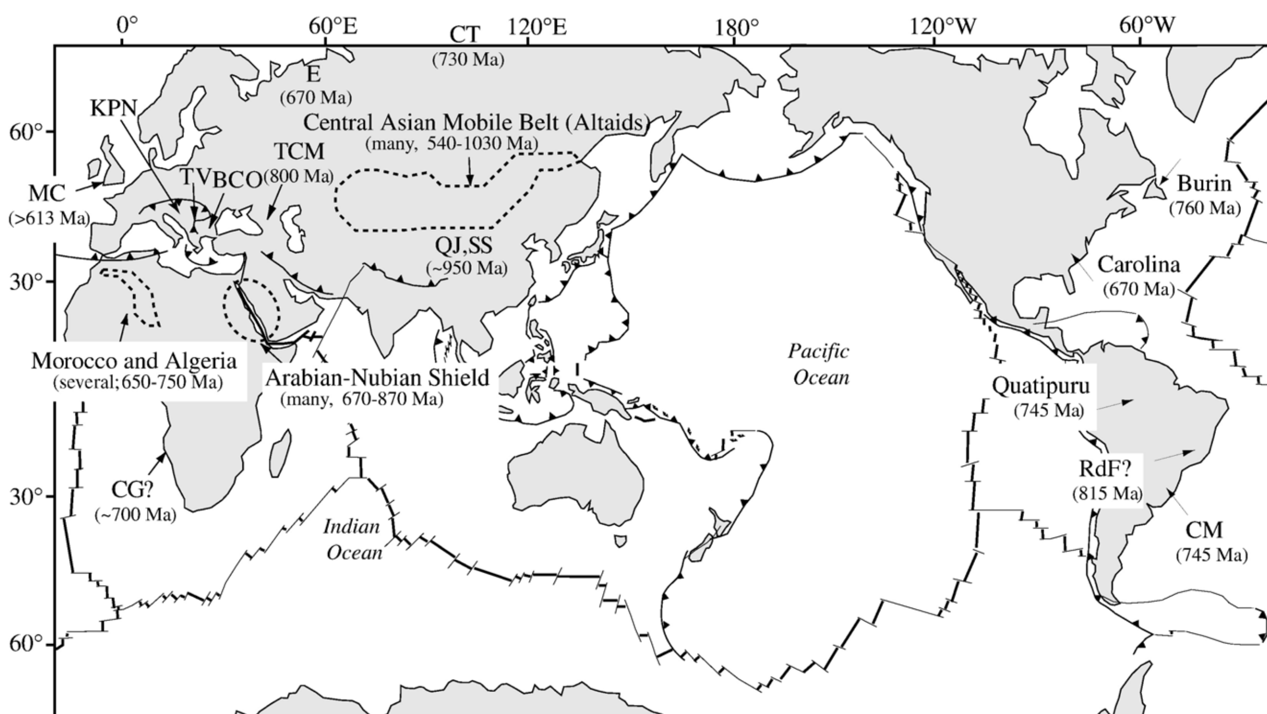

Fig. 4. Neoproterozoic ophiolites and ophiolitic regions of the world. Data sources and abbreviations were as follows: *Eurasia*: BC (Balkan–Carpathian ophiolites: Savov et al., 2001); E (Enganepe: Scarrow et al., 2001); KPN (870–550 Ma Kraubath-Pernegg-Hochgrössen, Austria: Melcher et al., 2002; Malitch, 2004); MC (Mona Complex, Wales: Thorpe, 1978; Tucker and Pharaoh, 1991); TCM (800±100 Ma Trans-Caucasian massif melange mafic–ultramafic association, Georgia (Zakariadze et al., 2007). *Asia*: China: SS (Shimian ophiolite, Sichuan: Shen et al., 2002); QJ (Qinling-Jiangxi ophiolite: Zhang et al., 2003); Russia: (570 Ma Agardagh Tes-Chem ophiolite: Pfänder et al., 2002); (697 Ma Enisei ophiolite: Vernikovsky et al., 2001); CT (~730 Ma Central Taimyr ophiolites: Vernikovsky and Vernikovskaya, 2001); (1020 Ma Dunzhugur ophiolite: Khain et al., 2002); *Africa*: Arabian–Nubian Shield (Dilek and Ahmed, 2003; Stern et al., 2004); Algeria (Black et al., 1994; Caby, 2003a); Morocco (Hefferan et al., 2000; Thomas et al., 2002; Samson et al., 2004); GC (Chameis Complex, Gariep Belt, Namibia: Frimmel et al., 1996); *S. America*: Rdf (Ribeirão da Folha, Minas Gerais, Brazil: Pedrosa-Soares et al., 1998); Quatipuru (Paixao and Nilson, 2001); CM (Cerro Mantiqueiras, Rio Grande do Sul, Brazil: Leite et al., 1998); *N. America*: Carolina (Hibbard et al., 2002); Burin (O’Driscoll et al., 2001).

on all continents except Australia and Antarctica. They are especially common in the Arabian–Nubian Shield (ANS) and the Central Asian Orogenic Belt (CAOB). ANS ophiolites have been recently reviewed (Stern et al., 2004), but an overview of the ages and tectonic setting of CAOB ophiolites is needed. ANS ophiolites range in age from ~870 to 690 Ma and mostly have chemical and mineralogic (esp. chromite) compositions consistent with formation in a suprasubduction zone (SSZ) tectonic setting, although the Gerf ophiolite in SE Egypt has compositions that indicate a MORB setting (Zimmer et al., 1995).

The ages of CAOB ophiolites straddle those of ANS ophiolites. Some are older than ANS ophiolites, such as the ~1040 Ma Dunzhugur ophiolite of southern Siberia (Khain et al., 2002). Other CAOB ophiolites are younger than 690–870 Ma ANS ophiolites, such as Mongolian ophiolites (~570 Ma Bayankhongor, Dariv, and Khawtaishir; Khain et al., 2003) and the ~570 Ma Agardagh-Tes-Chem ophiolite (Pfänder et al., 2002; Pfänder and Kröner, 2005) and the 627± 25 Ma Chaya ophiolite of Siberia (Amelin et al., 1997). These ophiolites mark the growth and destruction of the large, Pacific-scale, Paleo-Asian ocean basin between E. Gondwana and Siberia. The fact that the Paleo-Asian ocean existed all through Neoproterozoic time implies great size, perhaps 4000 km or more across (Dobretsov et al., 2003; Khain et al., 2003). Closing the Paleo-Asian ocean was accompanied by the development of numerous arcs that existed well into Paleozoic time, although arc-continent collisions began as early as ~800 Ma in Siberia (Kuzmichev et al., 2001). The long history of the Paleo-Asian ocean thus implies a long history of arc magmatism around its margins, suggesting that the CAOB was a very important site of JNPC formation, similar to the ANS. Possibly the Paleo-Asian Ocean and Mozambique Ocean were really parts of the same large ocean basin.

In addition to the great concentrations of ophiolites in the ANS and CAOB, there are also ophiolites in China and farther west in Europe and Africa. Chinese ophiolites are found in the northern Qinling mountains (S of Tarim) and in NE Jiangxi province (Zhang et al., 2003). These are mostly older than the bulk of Chinese Neoproterozoic igneous activity, ranging in age from ~1030 Ma ophiolites in East Qinling (e.g., Songshugou ophiolite; Yunpeng et al., 1997) and Anhui (Fuchuan ophiolite; Shen et al., 1992) to ~940 Ma ophiolites in NE Jiangxi and Sichuan (Yunpeng et al., 1997; Zhang et al., 2007). The ~940 Ma ophiolites have MORB affinities, whereas the ~1030 Ma Fuchuan ophiolite may have formed in a back-arc basin. Possible Neoproterozoic ophiolite sequences in China should be carefully examined, as shown by controversy over the

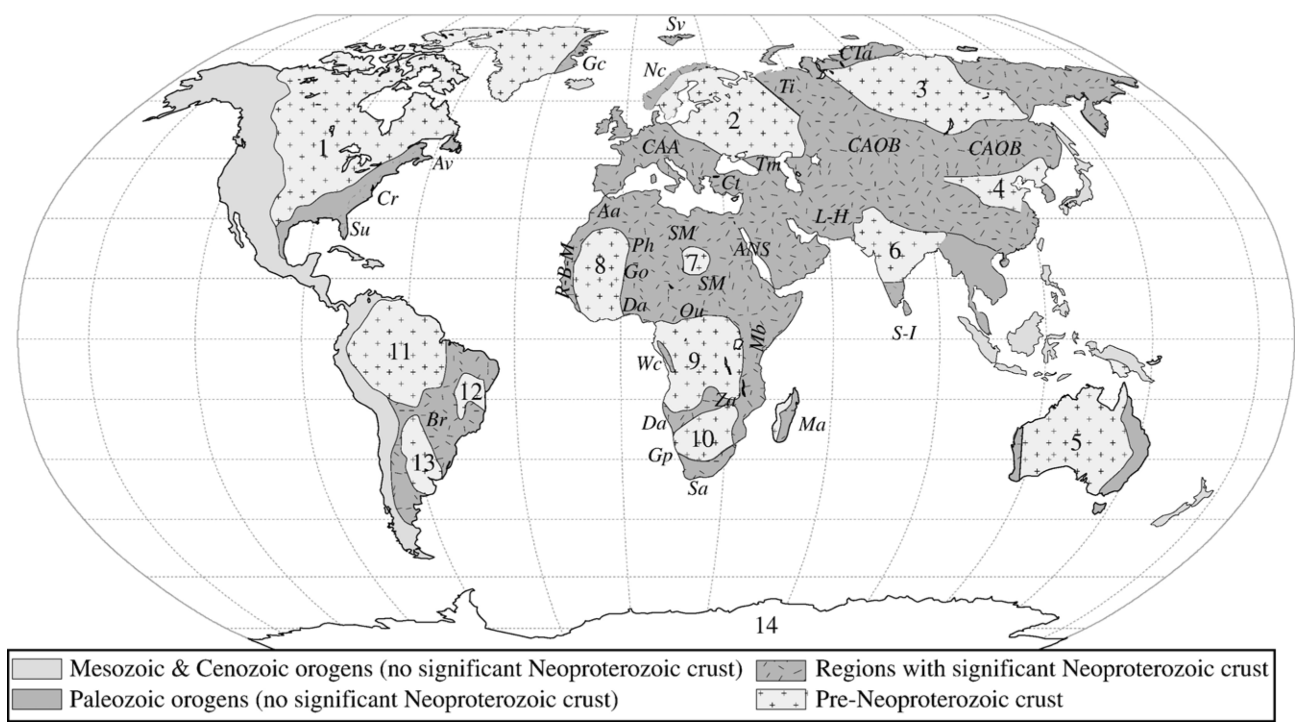

Fig. 5. Global distribution of Neoproterozoic crust, map modified after (Ernst et al., in press). Numbers correspond to pre-Neoproterozoic cratons: 1 = North American-Greenland craton; 2 = Baltic-East European craton; 3 = Siberian craton; 4 = N. China craton; 5 = Australia craton; 6 = Indian craton; 7 = Uweinat; 8 = West African craton; 9 = Congo craton; 10 = S. African craton; 11 = Amazon craton; 12 = Sao Francisco craton; 13 = Rio de la Plata craton; 14 = E. Antarctic craton. Letters in italics correspond to Neoproterozoic terranes and crustal tracts: *N. America-Greenland*: *Su* = Suwanee terrane; *Ca* = Carolina terrane; *Av* = Avalonia terrane; *Gc* = E. Greenland Caledonides. *S. America*: *Br* = Brasiliano orogens. *Europe*: *Sv* = Svalbard; *Nc* = Norwegian Caledonides; *Ti* = Timarides; *CAA* = Caledonian, Amorican, and Avalonian terranes. *Africa*: *R-B-M* = Rokelides, Bassarides, and Mauritanian belt; *Aa* = Anti-Atlas; *Ph* = Pharusian belt; *Go* = Gourma belt; *Da* = Dahomides; *Ou* = Oubangides; *SM* = Saharan Metacraton; *Wc* = West Congo belt; *Da* = Damaran orogen; *Gp* = Gariep belt; *Sa* = Saldanian belt; *Za* = Zambezi belt; *Mb* = Mozambique belt; *ANS* = Arabian–Nubian Shield; *Ma* = Madagascar. *Asia*: *Ct* = Cadomian of Turkey; *LH* = Lut block & Helmand block; *Tm* = Trans-causacian massif; *CAOB* = Central Asian Orogenic Belt; *S-I* = Sri Lanka and southern India; *CTa* = Central Taimyr accretionary belt.

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Longsheng "ophiolite", which Ge et al. (2000) regard as an intrusion into the supracrustal sequences of the Nanhua rift.

There are quite a few Neoproterozoic ophiolites in Europe, as shown in Fig. 4. These include those in the Balkan–Carpathian region (Savov et al., 2001), Enganepe in the Russian Arctic (Scarrow et al., 2001), Kraubath-Pernegg-Hochgrössen complexes in Austria (Melcher et al., 2002; Malitch, 2004), and the Mona Complex in Anglesy, Wales (Thorpe, 1978; Tucker and Pharaoh, 1991). This abundance indicates that significant oceanic realms must have existed between European Neoproterozoic exposures.

6. Global distribution of Neoproterozoic crust

Neoproterozoic crust is found on all of the continents. Fig. 5 is a qualitative assessment based on literature survey that summarizes the distribution of "significant" Neoproterozoic crust. Below these occurrences are briefly summarized and discussed, on a continent-by-continent basis. Table 2 presents the only estimate that we know for the volume of Neoproterozoic crust. Below we discuss the nature of Neoproterozoic crust on each of the continents in order of decreasing area of Neoproterozoic crust.

6.1. Africa

Africa was greatly affected by Neoproterozoic igneous activity, reflecting pervasiveness of Neoproterozoic tectonics and magmatism in Gondwanaland and Africa's location at the center of this supercontinent. Mobile belts dominated by Neoproterozoic igneous and metamorphic activity and deformation define the "Pan-African" orogenic belts that rim all African cratons. Other cratonic tracts, such as the Saharan Metacraton, were pervasively remobilized. An up-to-date overview of the Neoproterozoic evolution of Africa can be found in (Kröner and Stern, 2004). There are marked differences between the Neoproteozoic crust formation of northern and southern Africa. Northern Africa contains vast tracts of JNPC whereas Neoproterozoic orogens of southern Africa are mostly JNPC, including the Gariep, Saldania, Damaran, West Congo, and Zambezi belts. In this sense, the Neoproterozoic of southern Africa is very similar to that of South America.

The great Eburnian (~2.1 Ga) West Africa Craton (WAC) is surrounded for ~6000 km by Neoproterozoic orogens. The Rokelide, Mauritinide, and Bassalide mobile belts define its western margin (Culver et al., 1991; Lécorché et al., 1991), whereas JNPC exposed in the Anti-Atlas of Morocco define its northern margin (Hefferan et al., 2000; Thomas et al., 2002). The >3000 km-long Trans-Saharan or Pharusian belt of Algeria, Mauritania, and Niger extends south along the eastern margin of the craton into the Dahomeyides of Togo and Nigeria (Caby, 2003b). The Pharusian belt contains abundant JNPC and significant MORN, including the oldest known (~620 Ma) tract of ultra-high pressure metamorphism on Earth (Jahn et al., 2001). The Central Hoggar block in Algeria contains Archean/Paleoproterozoic and Neoproterozoic tracts thought to comprise a Neoproterozoic passive margin on the western margin of the Saharan metacraton (Liégeois et al., 2003).

The Saharan Metacraton (SM) is a huge (~5,000,000 km^2) tract of older crust in N. Africa that was extensively affected by Neoproterozoic tectonomagmatic activity. The SM extends from the Arabian–Nubian Shield in the east to the Hoggar/Tuareg Shield to the west and from the Congo craton in the south to the Mediterranean. The igneous rocks that have been dated generally give Neoproterozoic Rb–Sr whole-rock and U–Pb zircon ages but much older Nd model ages (Abdelsalam et al., 2002; Liégeois et al., 2003). Neoproterozoic remobilization of older crust was accompanied by significant volumes of crust formation related to rifting, small ocean basin opening, and subduction (Sultan et al., 1994; Suayah et al., 2006). In some ways the SM is reminiscent of better documented MORN of the S. China craton. Regardless, the SM is one of the largest tracts of poorly known crust on Earth and should be the subject of a regional isotopic and geochronologic study.

Insights about the SM come from recent studies of the Central African Fold Belt (CAFB), also known as the Oubangides. This Neoproterozoic orogen marks the northern margin of the Paleoproterozoic–Archean Congo craton (Van Schmus et al., in press) through Cameroon, southern Chad, and the Central African Republic. CAFB igneous rocks in Cameroon mostly yield crystallization ages <700 Ma but much older Nd model ages (Toteu et al., 2004) indicating MORN, but in Chad the igneous rocks are 570–740 Ma and have Nd isotopic compositions indicating abundant JNPC (Penaye et al., 2006). The CAFB may have formed by collision between the Congo craton and the SM (Toteu et al., 2004).

Like the WAC, the Congo craton is also surrounded by Neoproterozoic orogens, in addition to the Oubangides on its northern flank. The West Congo belt lies near its western margin, stretching over 1300 km from southwestern Gabon into northwestern Angola. A thrust-fold-belt with NE-verging transport direction in the west grades into a foreland basin built on Paleoproterozoic crust in the east. The succession resulted from Tonian rifting (1000–910 Ma) along the western margin of the Congo craton (Tack et al., 2001) followed by subsidence and formation of a carbonate-rich foreland basin, in

Table 2
Neoproterozoic (NPZ) crust on Earth

Continent	Area**	% of continental area	% NPZ*	NPZ area **	% of Earth's NPZ crust
Africa	30.37	20.36	50.60	15.37	62.33
Asia	43.81	29.38	9.10	3.99	16.17
S. America	17.84	11.96	14.90	2.66	10.78
Antarctica	13.72	9.20	8.30	1.14	4.62
Europe	10.4	6.97	8.40	0.87	3.54
N. America	24.49	16.42	1.60	0.39	1.59
Australia & New Guinea	8.5	5.70	2.80	0.24	0.97
Total	149.13			24.65 ~17% of Earth's crust	

* From Table 5-1 of Goodwin (1991).
** Area in (10^6 km^2).

which the West Congolian Group was deposited between ca. 900 and 570 Ma ago. In the west, an allochthonous thrust-and-fold stack of Palaeo- to Mesoproterozoic basement rocks overrode the West Congolian foreland sequence. The West Congo belt may be the eastern part of an orogenic system with the western part, including an 800 Ma ophiolite, exposed in the Aracuaí belt of Brazil (Pedrosa-Soares et al., 2001).

A complex Neoproterozoic orogenic belt separates the Congo craton from the Kalahari craton to the south. From east to west, this consists of the Zambezi belt, Lufilian arc, and Damaran belt. The Zambezi belt connects eastwards with the East African Orogen and records interactions between the Congo and Kalahari cratons during collisional assembly of the Gondwana supercontinent at the end of the Neoproterozoic. The Zambezi belt mostly consists of strongly deformed amphibolite- to granulite-facies Tonian and early Cryogenian ortho- and parag-neisses, locally intruded by ~860 Ma layered gabbro-anorthosite bodies and generally display S-vergent thrusting and transpressional shearing (Hargrove III et al., 2003).

The Zambezi belt transitions westwards into the Lufilian arc along the ~530 Ma transcurrent Mwembeshi shear zone (Hanson et al., 1993). In contrast to the dominantly high-grade metamorphic nature of the Zambezi belt, the Lufilian arc is mostly low-grade metasediments of the ~10 km thick Katanga Supergroup (880–500 Ma; Wendorff, 2005), which contains lavas with U–Pb zircon ages between 765 and 735 Ma (Key et al., 2001). The outer (northern) Lufilian arc is a northeast-verging, thin-skinned, low-grade fold-and thrust belt, whereas the inner (southern) Lufilian arc has basement-involved thrusts. The Lufilian arc continues southwestward into the Damara belt of Namibia, connected through isolated outcrops in northern Botswana.

The Damara Supergroup records basin formation and rift-related magmatism at ~760 Ma, followed by the formation of a broad carbonate shelf in the north and a turbidite basin in the south. Crustal shortening between the Congo and Kalahari Cratons mainly occurred between 550 and ~500 Ma (Johnson et al., 2006). The Damara belt underwent north- and south-vergent thrusting along its respective margins, whereas the deeply eroded central zone exposes medium to high-grade ductilely deformed rocks, widespread migmatization and anatexis in which both the Damara supracrustal sequence and a 1.0–2.0 Ga old basement are involved (Jung and Mezger, 2001). Damaran intrusive rocks are 840–460 Ma and are mostly MORN (Jung et al., 1998). The Damaran belt in Namibia branches near the Atlantic coast and continues southwards into the Gariep and Saldania belts and northwards into the Kaoko belt.

The Gariep, Saldania, and Kaoko belts are interpreted to result from oblique closure of the Adamaster Ocean, which opened ~780–700 Ma, separating the Rio de la Plata Craton (South America) from the Kalahari and Congo cratons. Closing of this ocean basin led to continental collision and deformation of the coast-parallel Kaoko, Gariep, and Saldania belts (Rozendaal et al., 1999).

The Kaoko belt is mostly MORN and extends 700 km NW from the Damara belt into southwestern Angola (Goscombe et al., 2005). Like the W. Congo belt, Neoproterozoic continental margin sequences of the Congo craton were overthrust eastwards, by a tectonic mixture of pre-Neoproterozoic basement and Neoproterozoic rocks during an oblique transpressional event following closure of the Adamastor Ocean (Goscombe et al., 2005). High-grade metamorphism and migmatization dated between 650 and 530 Ma affected both basement and cover rocks, and granitoids were emplaced between 733 and 550 Ma. The western part of the Kaoko belt is dominated by ~550 Ma crustal melt granites.

The Gariep Belt lies along the western margin of the Kalahari craton. It is subdivided by Frimmel et al. (1996) into: 1) an eastern, para-autochthonous, predominantly sedimentary rift and passive continental margin succession (Port Nolloth Zone, PNZ); and 2) a western, allochthonous, predominantly basaltic Marmora Terrane. The latter has been thrust southeast over the former. Early igneous activity in the PNZ accompanied rifting, beginning ~740 Ma (Frimmel and Fölling, 2004). The Gariep basin opened ~717 Ma. There is evidence of oceanic seamounts and MORB in the Marmora Terrane, and Nd isotopic compositions indicate significant JNPC (Frimmel et al., 1996). Peak metamorphism associated with closing the Adamaster Ocean occurred at ~545 Ma (Frimmel et al., 1996). The likely southward continuation of the Gariep belt is the Saldania Belt of the Western Cape Province, South Africa (Belcher and Kisters, 2003). The Saldania belt is mostly underlain by low-grade pelitic and psammitic metasediments and subordinate mafic volcanic rocks of the Malmesbury Group. Granitic intrusions to form the Saldania batholith occurred from ~550 to 515 Ma (Scheepers and Armstrong, 2002).

The locus of collision between east and west Gondwana (Fig. 1) is marked by the East African Orogen (EAO: Stern, 1994; Meert, 2003). The EAO consists of the high-grade metamorphic Mozambique Belt in the south (Tanzania, Kenya, Madagascar, and Mozambique; Kröner, 2001; Sommer et al., 2003) and the mostly lower-grade Arabian–Nubian Shield (ANS) in the north (Stern, 2002a). This along-strike variation probably reflects the fact that terminal collision was much more intense in the Mozambique Belt than in the ANS (Stern, 1994). The southern EAO continues through the largely MORN Mozambique Belt south through reconstructed Gondwana into Antarctica, such that (Jacobs and Thomas, 2004) renamed the entire ~8000 km long belt as the East Africa-Antarctica Orogen (EAAO), making it one of the largest orogenic belts on Earth.

The ANS is exposed for about 1,000,000 km^2, extending from Israel to Ethiopia and from the Nile to central Arabia. The ANS is mostly JNPC that formed within and adjacent to a large oceanic basin known as the Mozambique Ocean (Stern, 1994; Johnson and Woldehaimanot, 2003). The ANS mostly formed as a result of accretion of Mozambique Ocean intra-oceanic arcs during the Cryogenian Period (Roobol et al., 1983; Schandelmeier et al., 1994; Tadesse et al., 1999; Katz et al., 2004; Teklay, 2006). ANS formation began ~870 Ma and ended with terminal collision along the EAAO ~630 Ma (Meert and Torsvik, 2003). The Mozambique Belt also includes a significant proportion of Neoproterozoic arc-related igneous rocks, although these mostly formed at Andean-type margins (Handke et al., 1999).

The Mozambique Belt suffered intense collision-related metamorphism and magmatism ~650–600 Ma, followed by orogenic collapse that continued into Lower Paleozoic time (De Wit et al., 2001; Tsige, 2006). Although the ANS was less intensely affected by the collision, its magmatic and tectonic styles also reflect mostly collisional and post-collisional settings during the Ediacaran Period (Beyth et al., 1994; Blasband et al., 2000; Asrat et al., 2004).

6.2. Asia

Our understanding of Neoproterozoic crust in Asia is advancing rapidly, although much is buried or overprinted by younger tectonothermal episodes. Neoproterozoic crust is estimated to underlie ~9.1% of the continent (Table 2), and is concentrated in SW and S Asia, the Central Asian Orogenic Belt, and China.

SW Asia contains abundant Neoproterozoic crust. This is especially true for Jordan, Israel, Arabia, and Yemen which are underlain by the Arabian Shield, which is the eastern half of the Neoproterozoic Arabian–Nubian Shield (ANS: Jarrar et al., 2003; Johnson and Woldehaimanot, 2003), a large expanse of mostly JNPC discussed further in the next section. The eastern and northern boundary of the ANS is poorly known because this is buried beneath Phanerozoic sediments, but Neoproterozoic crust appears to underlie much of this region. Scattered basement exposures east of the Arabian Shield yield Neoproterozoic ages, and the Arabian peninsula appears to be underlain by mostly Neoproterozoic crust (Johnson and Kattan 2007). The crust north of Arabia also appears to be largely Neoproterozoic, including much of western Turkey (Loos and Reischmann, 1999; Koralay et al., 2004; Ustaömer et al., 2005), and Iran (Ramezani and Tucker, 2003). The crust of Syria, Iraq, Afghanistan, and Pakistan is poorly known, but the abundance of Hormuz salt, which formed in end-Neoproterozoic salt basins, suggests that the underlying crust was affected by strong extension and perhaps igneous activity at the end of Neoproterozoic time (Bahroudi and Talbot, 2003). The recognition of early Cryogenian igneous rocks just north of the Caucasus Mountains in Georgia (Zakariadze et al., 2007) suggests that the broad tract of Neoproterozoic crust beneath the Middle East continues across the Tethyan sutures north to the Timanides in Russia.

Neoproterozoic metamorphosed and deformed igneous rocks – mostly MORN – are common in southern India and Sri Lanka. Extensive Neoproterozoic magmatism in the Malani igneous province of NW India (771–751 Ma), along with the Seychelles islands (mostly 748–755 Ma) and central-northern Madagascar (824–720 Ma) may have constituted a single Andean-type arc on the western margin of East Gondwana (Torsvik et al., 2001).

Asian crust is anchored by four great Archean cratons (Fig. 5): East Europe-Baltica in E. Europe; Siberia; North China; and India in the south. Neoproterozoic crust is tectonically mixed in with fragments of Paleozoic and Mesozoic crust, and these polycyclic orogenic tracts are sandwiched between the cratons. The greatest of the Neoproteozoic–Phanerozoic polyorogenic tracts is the CAOB, also known as the Altaids. This is an accretionary orogen stretching from Mongolia to the Urals that began to grow ~1.0 Ga and continued until Mesozoic time (Yakubchuk, 2004; Windley et al., 2007). The Neoproterozoic history of the CAOB is also known as the Baikalides or pre-Uralides. Evidence of Neoproterozoic igneous activity is found in ophiolites, fragments of continental-margin and rift-related magmatic belts, bimodal volcanic associations, mafic intrusions, and granites (850–700 Ma). Neoproterozoic igneous rocks are also associated with rifts along the southern and southwestern margins of the Siberian craton, including basaltic dyke swarms, ultramafic rock-carbonatite complexes. Along the western margin of the Siberian craton, Vernikovsky et al. (2003) identified three main Neoproterozoic tectonic events involved in forming the Yenisey Ridge fold-and-thrust belt: 880–860 Ma, 760–720 Ma and 700–630 Ma. It is difficult to assess the amount of Neoproterozoic igneous rocks in a region that is as large and complex as the CAOB, but Neoproterozoic Nd model ages are common (Hong et al., 2004; Kovalenko et al., 2004), indicating that significant JNPC underlies much of the CAOB.

The Precambrian basement of China can be subdivided into the N. China or Sino-Korean craton, which was not affected by Neoproterozoic tectonics and magmatism, and the S. China block, which is MORN: largely Paleo-and Mesoproterozoic crust that was partly remobilized in early Neoproterozoic time, along with new additions from the mantle to the crust. The relationship of the N. and S. China blocks to basement farther west is poorly understood, but Neoproterozoic igneous and metamorphic rocks are common beneath and around the Tarim Basin (Tibet; Xu et al., 2005), where Chinese Neoproterozoic crust grades into the CAOB.

Most Neoproterozoic magmatism in S. China occurred 950–760 Ma ago (Zhou et al., 2002a). Crustal growth in S. China for the first 100 million years of the Neoproterozoic is thought to have occurred at intra-oceanic or intra-continental arcs. This phase ended when the Yangtze and Cathysia blocks collided ~850 Ma to form a ~1500-km long suture and the composite South China block (Jiangnan orogen; Wang et al., 2004b). Arc volcanism on the northern and western margins of the Yangtze craton continued from ~850 Ma until ~760 Ma (Zhou et al., 2002a; Yan et al., 2004), with suturing of the arc to S. China ~690–660 Ma (Yan et al., 2004).

Igneous activity in S. China reached a maximum ~820 Ma (Li et al., 2005). This episode was associated with the emplacement of granitoids over a broad area (700×1000 km) of the Yangtze craton. Granitic rocks are diverse, including two types of peraluminous, S-type granitoids (muscovite-bearing leucogranite and cordierite-bearing granodiorite) and two types of I-type granitoids (K-rich calc-alkaline granitoids and tonalite–trondhjemite–granodiorite). Nd isotopic data indicates that all were generated by crustal anatexis with little involvement of new mantle-derived magmas and thus these plutons are clearly MORN. Li et al. (2003) suggested that these granitoids formed by crustal melting above a mantle plume, but this is controversial (Wang et al., 2004a). Controversy also persists about whether mantle-derived mafic melts emplaced ~820–780 Ma are plume-related (Li et al., 1999) or are arc-related (Zhou et al., 2002a,b). This or a related mantle plume also appears to have affected southern Australia about the same time. Following the Jiangnan orogeny, S. China experienced four episodes of Neoproterozoic rifting, which correlates well with the Neoproterozoic rift history of Australia (Wang and Li, 2003). This has led to inferences that S. China and Australia-Antarctica were adjacent during much of Neoproterozoic time (Li et al., in press).

It is difficult to unequivocally interpret the significance of Nd isotopic data for S. Chinese Neoproterozoic igneous rocks as exclusively MORN or JNPC. Epsilon-Nd are generally low positive or negative, suggesting that Neoproterozoic granioids largely formed by remelting older continental crust (Chen and Jahn, 1998). This interpretation is equivocal because many Chinese Neoproterozoic mafic igneous rocks with little evidence for fractionation or assimilation, as indicated by moderate to high Mg#, also have low or even negative epsilon-Nd (Ling et al., 2003; Li et al., 2005). Furthermore, inherited zircons of pre-Neoproterozoic age are not very common in these igneous rocks, as would be expected if older crust was involved. The low and variable epsilon-Nd of mafic igneous rocks may instead reflect enriched mantle sources, probably ancient subcontinental lithosphere. A few mafic intrusions and high-Mg andesites show strongly positive epsilon-Nd(t) values, consistent with juvenile additions from the mantle to the crust (e.g., 820 Ma Wangjiangshan intrusion, epsilon-Nd = +3.5 to +5.9; Zhou et al., 2002a; 950–895 Ma Xixiang volcanics, epsilon-Nd = +2.0 to +8.8; Ling et al., 2003), but these are exceptional. Nevertheless, there is strong evidence that the early Neoproterozoic witnessed significant JNPC in S. China, because sediments of southern China show a dramatic decrease in Nd model ages at ~800 Ma (Li and McCulloch, 1996).

Another noteworthy feature of mid-Neoproterozoic volcanism in China is that some volcanic successions inferred to have erupted in extensional environments show arc-like geochemical features, especially Nb–Ta depletions, such as is seen for the ~755 ma Beiyixi bimodal volcanics of the Tarim block in NW China (Xu et al., 2005). It is also noteworthy that even relatively primitive Beiyixi basalts have strongly negative epsilon-Nd (−9 to−11; Xu et al., 2005). These authors concluded that the presence of arc-like geochemical features in rift-related volcanics manifested a subcontinental lithosphere melt source rather than being due to crustal contamination. This may be part of the reason that Neoproterozoic igneous rocks around the Yangtze craton are thought to be arc-related. Alternatively, this may reflect tectonic evolution from early subduction-related magmatism to later plume-related magmatism, as argued for volcanic sequences around the NW Yangtze craton. In contrast, Cathaysian Neoproterozoic volcanics have OIB-like trace element compositions (Li et al., 2004).

Neoproterozoic crust formation in South Korea reflects that of S. China on a smaller scale, and the Gyeonggi massif in particular may be a part of the South China Block. Lee et al. (2003a) interpreted 742 ± 13 A-type granitic magmatism in the Gyeonggi massif to have been associated with extension. Similar ages and interpretations exist for other S. Korean igneous and metamorphic rocks: Kim et al. (2006) inferred that 762 ± 7 Ma bimodal volcanics from the Okcheon metamorphic belt represents the NE extension of the Nanhua rift of S. China. Rifting and bimodal magmatism postdates the emplacement of more primitive mafic magmas, such as those in the Imjingang belt (SHRIMP U–Pb zircon age of 861 ± 7 Ma; Cho et al., 2001, quoted in Lee et al., 2003a), and amphibolites of the central Gyeonggi massif (Sm–Nd whole-rock age of ca. 850 Ma; Lee and Cho, 1995). The Neoproterozoic basement of S. Korea is probably MORN, as shown by the fact that Nd model ages for Gyeonggi massif igneous rocks are 2.9–2.5 Ga and 1.9–1.8 Ga (Lee et al., 2003b).

6.3. South America

Neoproterozoic crust is estimated to underlie ~15% of South America (Table 2), making this the continent with the second largest proportion of Neoproterozoic crust. The South American Platform makes up the stable Precambrian crust of South America and covers an area of about 15,000,000 km^2, some 40% of which is exposed in three shields: Guiana, Guaporé, and Atlantic. About 80% of the basement exposures formed during Archean and Paleoproterozoic time, and these rocks are principally preserved in three cratons: Amazonian, Sao Francisco, and Rio de la Plata–Luis Alves. These cratons are flanked by Neoproterozoic mobile belts: the Amazonian craton is flanked by the Paraguai–Araguaia–Tocatins belts, the Sao Francisco craton is surrounded by the Borborema, Brasilia, Ribeira, Mantequeira, and Araçuai belts, and the Rio de la Plata- and Luis Alves cratonic fragments are flanked by the Dom Feliciano belt (Cordani and Sato, 1999). Neoproterozoic mobile belts are common in the southeastern part of the platform and are rare in the northwest.

Neoproterozoic tectonomagmatic events are grouped together as expressions of the Brasiliano orogeny. This has been subdivided into Brasiliano I (~850–700 Ma), Brasiliano II (650–600 Ma), and Brasiliano III (590–540 Ma; Da Silva et al., 2005). Brasiliano I was largely JNPC and included development of juvenile intra-oceanic magmatic arcs, whereas Brasiliano II and III episodes were MORN, involving collision and tectonothermal reworking of older crust. Some Brasiliano orogens correlate across the Atlantic Ocean into Neoproterozoic mobile belts of Africa.

Neoproterozoic igneous rocks are abundant in South America. Most contain abundant evidence for the involvement of older (Archean–Mesoproterozoic) crust, especially in terms of inherited zircons and isotopic compositions of Sr and Nd. Da Silva et al. (2005) noted "…the crucial distinction between the Neoproterozoic evolution in South America and Africa is based neither on the timing of the successive events, nor on the orogenic architecture, but on the scale of the earliest orogenic accretionary events. In South America, the known early juvenile crustal growth was very restricted, totaling perhaps less than 10% of the total exposed Brasiliano crust. On the other hand, the Pan-African orogens, especially from the north-west and east African continent, were much more efficient in terms of generation of new crust."

Much of the migmatitic, anatectically-reworked Neoproterozoic crust of S. America may reflect exhumed middle crust,

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which was largely molten during late Neoproterozoic time. Vauchez et al. (2007) report that the northern Ribeira-Araçuaí orogen along the Brazil coast was pervasively heated at high T and low P (>700 °C, 600 MPa, suggesting a ~35 °C km geotherm). This caused extensive melting in the middle crust, which may therefore represent an eroded analogue of the partially molten middle crust that today lies beneath the Tibetan and Altiplano orogenic plateaus (Vauchez et al., 2007). A similar mechanism of heating the middle crust to>700 °C may have been an important way to weaken and thermally rework other pre-Neoproterozoic crustal blocks, such as the S. China block and Saharan metacraton.

6.4. Antarctica

Antarctica is mostly buried beneath ice but Neoproterozoic crust is estimated by Goodwin (1991) to underlie ~8.3% of its area (Table 2). The continent is naturally divided by the Transantarctic Mountains into western and eastern portions. The crust of W. Antarctica was made by Phanerozoic terrane accretion (Mukasa and Dalziel, 2000). We know much less about the crust beneath E. Antarctica, and it is controversial whether its interior is underlain by a great pre-Neoproterozoic craton. Where exposed, East Antarctica basement is dominated by high-grade gneiss and comprise a number of Archean Paleoproterozoic crustal tracts and younger mobile belts (Fitzsimons, 2000a). Traditional explanations of E. Antarctic geology (summarized by Fitzsimons, 2000a) involve a 3-stage tectonic history: 1) stabilization of Archean–Paleoproterozoic craton by 1600 Ma; 2) Development of a high-grade "Grenville" orogen (~1300–900 Ma) that approximates the present coastline; and 3) late Neoproterozoic to early Paleozoic tectonism in the Ross orogen with modest reheating of the E. Antarctic craton. More recently Fitzsimons (2000b) concluded that Neoproterozoic orogenic belts traverse the E. Antarctic Shield, and these cut across three late Mesoproterozoic and early Neoproterozoic orogens. On this basis Fitzsimons (2000a) concluded that East Gondwana comprises at least three major continental fragments assembled during the late Neoproterozoic to Early Cambrian time, further suggesting that the proportion of Neoproterozoic crust in Antarctica is significantly greater than Goodwin (1991) estimated.

Some Neoproterozoic crust in Antarctica is related to the southern continuation of the East African Orogen (discussed in Section 6.1), such that Jacobs and Thomas (2004) redefined it as the ~8,000 km long East Africa-Antarctica Orogen (EAAO). There is also a lot of late Neoproterozoic crust that was affected by the Terra Australis Orogen (Cawood, 2005) and by rifting along the Trans-Antarctic mountains (Goodge et al., 2004).

6.5. Europe

Neoproterozoic crust surrounds the Paleoproterozoic–Archean Baltic–East European craton and is estimated to underlie ~8.5% of Europe (Table 2). It is especially abundant south of the craton, in the Avalonian and Cadomian terranes, on the SW side of the buried Trans-European suture zone (aka the "Tornquist line"). This suture has been traced geophysically for 3000 km SE from Denmark through Poland to the Black Sea (Dadlez et al., 2005). Neoproterozoic crust fragments are found in all of the Variscan massifs south and west of the Tornquist line, from Iberia–France–Britain through Germany and Austria in central Europe and farther south in the Balkans. Avalonian and Cadomian crust formed between ~650 and 600 Ma, followed by granitic plutonism in an Andean-type continental margin setting between ~570 and 520 Ma (Dörr et al., 2002; Neubauer, 2002; Linnemann et al., 2004). These late Neoproterozoic units locally overlie 2.0 Ga or 750 Ma basement (Chantraine et al., 2001). Similar and older Cryogenian terranes are found in Romania (Liégeois et al., 1996; Seghedi et al., 2005), Bulgaria (Savov et al., 2001), and Greece (Anders et al., 2006). This crust extends eastward along the southern margin of the East European craton as the Skythian plate (Seghedi et al., 2005), so that the Neoproterozoic crust of SE Europe is continuous with that of SW Asia, in spite of the fact that Europe and SW Asia are separated by Tethyan sutures of Mesozoic and Cenozoic age.

Even the relatively stable East European platform was affected by Neoproterozoic igneous activity, associated with rifting of Archean and Paleoproterozoic crust (Artemieva, 2003). On the eastern flank of the Baltic-East European craton, extensive Neoproterozoic crust is found in the Timanide orogen also known as the "Baikalides". The Timanides extend from the southern Ural Mountains of Kazakhstan to northernmost Norway, a distance of at least 3000 km (Gee and Pease, 2005). Neoproterozoic igneous rocks of the Timanian orogeny are also buried beneath Phanerozoic cover and the Arctic continental shelf and obscured by Uralian deformation.

6.6. North America

Neoproterozoic rocks are a relatively minor component of the North American crust, comprising ~2% of the continent (Table 2). Thus is concentrated around the margins of the Laurentian craton. As found for Laurasia, Paleozoic orogenic belts of eastern North America in particular contain abundant Neoproterozoic igneous rocks. These are especially common in the Carolinia and Suwanee terranes of the eastern US seaboard and the W. Avalonia, Ganderia and Meguma terranes in maritime Canada (Hibbard et al., 2002; Wortman et al., 2000; Nance et al., in press). Nance et al. (in press) linked these terranes with similar Neoproterozoic crustal tracts in western Europe as "peri-Gondwanan" terranes, which record evolution of one or more convergent margins along the northern flank of western Gondwana beginning ~760 Ma. Subduction was followed by development of a magmatically active transform margin of ~610–540 Ma.

Neoproterozoic igneous rocks are also found in western N. America and Alaska, mostly related to Neoproterozoic rifting. The most important may be the ~780 Ma old Laurentian LIP, manifested by a giant radiating dyke swarm which can be traced from the Mackenzie Mountains of northwestern Canada to the Wyoming province of the western USA. These dikes supplied magma to a large flood basalt province above it (Goddéris et al., 2003).

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There is little Neoproterozoic crust beneath Mexico and Central America, but that beneath Yucatan may be latest Neoproterozoic (Krogh et al., 1993).

6.7. Australia–New Guinea

Australia–New Guinea has the least Neoproterozoic crust of any continent (Table 2). Australia was affected by early Cryogenian crustal extension, beginning at 827±6 Ma (the age of the Gairdner dyke swarm; Wingate et al., 1998). This presumably fed eruption of flood basalts, although only mafic dikes and gabbros over >1000 km of southern Australia are preserved. The 827 Ma igneous rocks show remarkably uniform geochemical and isotopic features, including LREE-enriched patterns and a limited range of epsilon-Nd (800 Ma) values (+2.4 to+4.2) indicating derivation from enriched mantle (Zhou et al., 2002c). Their trace element characteristics resemble OIB and continental flood basalts, features that suggest generation by decompression melting of a large-scale, uniform asthenospheric mantle plume.

Another Neoproterozoic igneous episode is preserved in the Mundine Well dike swarm of W. Australia, which yields a ^{207}Pb/^{206}Pb age of 754±5 Ma for zirconolite (Rasmussen and Fletcher, 2004). These dikes are one manifestation of crustal extension; subsequent thermal subsidence formed large sedimentary basins in central-southern Australia (Preiss, 2000). These basins formed by multiple periods of mostly Neoproterozoic subsidence (Preiss, 2000): ~830 Ma, ~800 Ma, ~780 Ma, and ~700 Ma, followed by broad thermal subsidence from ~700 Ma to ~630 Ma, with a resumption of rifting ~630 Ma and culminating with renewed volcanism and rifting. Rifting events between ~780 Ma and ~580 Ma formed the Pacific Ocean basin (Meffre et al., 2004).

7. Concluding remarks

This review is surely oversimplified and incomplete, but it has hopefully made the point that the Neoproterozoic Era was an important time of crustal growth. Perhaps 20% of Earth's continental crust formed or was intensely reworked during this time. This emphasizes the importance of better understanding how the solid Earth system behaved during this time, knowledge that is important for its own sake as well as for understanding what caused and influenced the spectacular climatic and biological changes of this time. This review demonstrates that the evidence of Neoproterozoic crustal growth is found to very different extents on the continents. Africa by far preserves the greatest proportion of all Neoproterozoic crust (62%), followed by Asia (16%), South America (11%), Antarctica (5%), Europe (4%), North America (2%) and Australia (1%).

This review supports the interpretation that Neoproterozoic crustal growth took place within the context of a supercontinent cycle, from breakup of Rodinia beginning ~830 Ma and ending with the formation of a new supercontinent near the end of the Era. Neoproterozoic crust formation was quite similar to that of the Phanerozoic, dominated by convergent margin magmatism but strongly supplemented by intra-plate, rifting-related, and "hotspot" melts, especially during times of continental breakup. This may be why there was little crust formation during the first 150 Ma of the Neoproterozoic Era, a time when Rodinia remained intact. The paucity of the Tonian record, coupled with the fact that this time predates the major changes in climate, crust formation, and the biosphere, needs to be reviewed so that the solid Earth system before the major climatic and biospheric changes.

We agree with Windley (2003) that "…most estimates of the rate of growth of Proterozoic crust are premature and unreliable", especially as this pertains to quantitative assessments, but qualitative estimates are useful. This review rejects the interpretation that the Neoproterozoic Era was a time of reduced crustal growth, although there is much uncertainty: about how much Neoproterozoic crust is preserved today; about how much has been destroyed during Phanerozoic time; and about what proportion of Neoproterozoic crust were juvenile additions from the mantle as opposed to reworked, older continental crust. Reducing these uncertainties is feasible but requires a concerted, international effort, requiring the development of a global GIS for basement rocks, something that does not yet exist. Perhaps a new IGCP project to inventory Neoproterozoic crustal growth is needed.

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References

Abbott, D.H., Isley, A.E., 2002. The intensity, occurrence, and duration of superplume events and eras over geologic time. Journal of Geodynamics 34, 265–307.

Abdelsalam, A.G., Liégeois, J.-P., Stern, R.J., 2002. The Saharan metacraton. Journal of African Earth Sciences 34, 119–136.

Amelin, Y.V., Ritsk, E.Y., Neymark, L.A., 1997. Effects of interaction between ultramafic tectonite and mafic magma on Nd–Pb–Sr isotopic systems in the Neoproterozoic Chaya Massif, Baikal-Muya ophiolite belt. Earth and Planetary Science Letters 148, 299–316.

Anders, B., Reischmann, T., Kostopoulos, D., Poller, U., 2006. The oldest rocks of Greece: first evidence for a Precambrian terrane within the Pelagonian Zone. Geological Magazine 143, 41–58.

Anderson, D.L., 2005. Large igneous provinces, delamination, and fertile mantle. Episodes 1, 271–275.

Artemieva, I.M., 2003. Lithospheric structure, composition, and thermal regime of the East European Craton: implications for the subsidence of the Russian platform. Earth and Planetary Science Letters 213, 431–446.

Artemieva, I.M., 2006. Global 1°×1° thermal model TC1 for the continental lithosphere: implications for lithosphere secular evolution. Tectonophysics 416, 245–277.

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Artemieva, I.M., Mooney, W.D., 2001. Thermal thickness and evolution of Precambrian lithosphere: a global study. Journal of Geophysical Research B 106, 16387–16414.

Asrat, A., Barbey, P., Ludden, J.L., Reisberg, L., Gleizes, G., Ayalew, D., 2004. Petrology and isotope geochemistry of the Pan-African Negash Pluton, Northern Ethiopia: Mafic–Felsic magma interactions during the construction of shallow-level calc-alkaline plutons. Jounal of Petrology 45, 1147–1179.

Bahroudi, A., Talbot, C.J., 2003. The configuration of the basement beneath the Zagros Basin. Journal of Petroleum Geology 26, 257–282.

Belcher, R.W., Kisters, A.F.M., 2003. Lithostratigraphic correlations in the western branch of the Pan-African Saldania belt, South Africa: the Malmesbury Group revisited. South African Journal of Geology 106, 327–342.

Beyth, M., Stern, R.J., Altherr, R., Kröner, A., 1994. The late Precambrian Timna igneous complex, southern Israel: evidence for comagmatic-type sanukitoid monzodiorite and alkali granite magma. Lithos 31, 103–124.

Black, R., Latouche, L., Liégeois, J.P., Caby, R., Bertrand, J.M., 1994. Pan-African displaced terranes in the Tuareg shield (Central Sahara). Geology 22, 641–644.

Blasband, B., White, S., Brooijmans, P., DeBoorder, H., Visser, W., 2000. Late Proterozoic extensional collapse in the Arabian–Nubian Shield. Journal of the Geological Society of London 157, 615–628.

Brown, M., 2007. Crustal melting and melt extraction, ascent and emplacement in orogens: mechanisms and consequences. Journal of the Geological Society of London 164, 709–739.

Bryan, S.E., Riley, T.R., Jerram, D.A., Stephens, C.J., Leat, P.T., 2002. Silicic volcanism: an undervalued component of large igneous provinces and volcanic rifted margins. In: Menzies, M.A., Klemperer, S.I., Ebinger, C., Baker, J. (Eds.), Volcanic Rifted Margins. Geological Society of America Special Paper, vol. 362. Boulder CO, pp. 99–120.

Burret, C., Berry, R., 2000. Proterozoic Australia-Western United States (AUSWUS) fit between Laurentia and Australia. Geology 28, 103–106.

Caby, R., 2003a. Terrane assembly and geodynamic evolution of central-western Hoggar: a synthesis. Journal of African Earth Sciences 37, 133–159.

Caby, R., 2003b. Terrane assembly and geodynamic evolution of central-western Hoggar: a synthesis. Journal of African Earth Sciences 37, 133–159.

Cawood, P.A., 2005. Terra Australis orogen: Rodinia breakup and development of the Pacific and Iapetus margins of Gondwana during the Neoproterozoic and Paleozoic. Earth-Science Reviews 69, 249–279.

Chantraine, J., Egal, E., Thielemont, D., Le Goff, E., Guerrot, C., Ballevre, M., Guennoc, P., 2001. The Cadomian active margin (North Armorican Massif, France): a segment of the North Alantic Panafrican belt. Tectonophysics 331, 1–18.

Chen, J., Jahn, B.-M., 1998. Crustal evolution of southeastern China: Nd and Sr isotopic evidence. Tectonophysics 284, 101–133.

Clift, P., Vannucchi, P., 2004. Controls on tectonic accretion versus erosion in subduction zones: implications for the origins and recycling of the continental crust. Reviews of Geophysics 42. doi:10.1029/2003RG000127.

Condie, K.C., 2000. Episodic continental growth models: afterthoughts and extensions. Tectonophysics 322, 153–162.

Cordani, U.G., Sato, K., 1999. Crustal evolution of the South American Platform, based on Nd isotopic systematics on granitoid rocks. Episodes 22 (3), 167–173.

Corfield, R.I., Searle, M.P., Pedersen, R.B., 2001. Tectonic setting, origin, and obduction history of the Spontang Ophiolite, Ladakh Himalaya, NW India. Journal of Geology 109, 715–736.

Culver, S.J., Williams, H.R., Venkatakrishnan, R., 1991. The Rokelide orogen. In: Dallmeyer, R.D., Lécorché, J.P. (Eds.), The West African Orogens and Circum-Atlantic Correlatives. Springer-Verlag, Berlin, pp. 123–150.

Dadlez, R., Grad, M., Guterch, A., 2005. Crustal structure beneath the Polish Basin: is it composed of proximal terranes derived from Baltica? Tectonophysics 411, 111–128.

Dalziel, I.W.D., 1997. Neoproterozoic–Paleozoic geography and tectonics: review, hypothesis, environmental speculation. Geological Society of America Bulletin 109, 16–42.

Da Silva, L.C., McNaughton, N.J., Armstrong, R., Hartmann, L.A., Fletcher, I.R., 2005. The Neoproterozoic Mantiquiera Province and its African connections: A zircon-based U–Pb geochronologic subdivision for the Brasiliano/Pan-African system of orogens. Precambrian Research 136, 203–240.

Davidson, J.P., Arculus, R.J., 2005. The significance of Phanerozoic arc magmatism in generating continental crust. In: Brown, M., Rushmer, T. (Eds.), Evolution and Differentiation of the Continental Crust. Cambridge U. Press, pp. 135–172.

Defant, M.J., Drummond, M.S., 1990. Derivation of some modern arc magmas by melting of young subducted lithosphere. Nature 347, 662–665.

Dewey, J.F., 2005. Orogeny can be very short. Proceedings of the National Academy of Science 102 (43), 15286–15293.

De Wit, M.J., Bowring, S.A., Ashwal, L.D., Randrianasolo, L.G., Morel, V.P.I., Rambleson, R.A., 2001. Age and tectonic evolution of Neoproterozoic ductile shear zones in southwestern Madagascar, with implications for Gondwana studies. Tectonics 20 (1), 1–45.

Dickinson, W.R., 2004. Evolution of the North American Cordillera. Annual Reviews of Earth and Planetary Sciences 32, 13–45.

Dilek, Y., 2003. Ophiolite pulses, mantle plumes, and orogeny. In: Dilek, Y., Robinson, P.T. (Eds.), Ophiolites in Earth History. Geological Society of London Special Publication, vol. 218, pp. 9–20. London.

Dilek, Y., Ahmed, Z., 2003. Proterozoic ophiolites of the Arabian Shield and their significance in Precambrian tectonics. In: Dilek, Y., Robinson, P.T. (Eds.), Ophiolites in Earth History. Special Publications, London, vol. 218. Geological Society, London, pp. 685–700.

Dilek, Y., Furnes, H., Shallo, M., 2007. Suprasubduction zone ophiolite formation along the periphery of Mesozoic Gondwana. Gondwana Research 11, 453–475.

Dimalanta, C., Taira, Jr., A., G.P.Y., Tokuyama, H., Mochizuki, K., 2002. New rates of western Pacific island arc magmatism from seismic and gravity data. Earth and Planetary Science Letters 202, 105–115.

Dobretsov, N.L., Buslov, M.M., Vernikovsky, V.A., 2003. Neoproterozoic to Early Ordovician evolution of the Paleo-Asian ocean: implication to the breakup of Rodinia. Gondwana Research 6, 143–159.

Ducea, M., 2001. The California arc: thick granitic batholiths, eclogitic residues, lithosphere-scale thrusting, and magmatic flare-ups. GSA Today 11, 4–10.

Ducea, M., Saleeby, J., 1998. A case for delamination of the deep batholithic crust beneath the Sierra Nevada, California. International Geology Review 40, 78–93.

Dörr, W., Zulauf, G., Fiala, J., Franke, W., Vejnar, Z., 2002. Neoproterozoic to early Cambrian history of an active plate margin in the Teplá-Barrandian unit-a correlation of U–Pb isotopic-dilution-TIMS ages (Bohemia, Czech Republic). Tectonophysics 352, 65–85.

Ernst, W.G., 2005. Alpine and Pacific styles of Phanerozoic mountain building: subduction-zone petrogenesis of continental crust. Terra Nova 17, 165–188.

Ernst, W.G., Hacker, B.R., Liou, J.G., in press. Petrotectonics of ultrahigh-pressure crustal and upper mantle rocks-implications for Phanerozoic collisional orogens. In: Sears, J., Harms, T., Evanchick, C. (Eds.), Whence the Mountains? Inquiries into the Evolution of Orogeic Systems: A Volume in Honor of Raymond A. Price. Geological Society of America Special Paper vol. 433. The Geological Society of London, Boulder.

Evans, D.A.D., 2000. Stratigraphic, geochronological, and paleomagnetic constraints upon the Neoproterozoic climate paradox. American Journal of Science 300, 347–433.

Fairchild, I.J., Kennedy, M.J., 2007. Neoproterozoic glaciation in the Earth System. Journal of the Geological Society of London 164, 895–921.

Fitzsimons, I.C.W., 2000a. Grenville-age basement provinces in East Antarctica: evidence for three separate collisional orogenies. Geology 28, 879–882.

Fitzsimons, I.C.W., 2000b. A review of tectonic events in the East Antarctic Shield and their implications for Gondwana and earlier supercontinents. Journal of African Earth Sciences 31, 3–23.

Flowers, R.M., Royden, L.H., Bowring, S.A., 2004. Isostatic constraints on the assembly, stabilization, and preservation of cratonic lithosphere. Geology 32, 321–324.

Frimmel, H.E., Fölling, P.G., 2004. Late Vendian closure of the Adamastor Ocean: timing of tectonic inversion and syn-orogenic sedimentation in the Gariep Basin. Gondwana Research 7 (3), 685–699.

Frimmel, H.E., Hartnady, C.J.H., Koller, F., 1996. Geochemistry and tectonic setting of magmatic units in the Pan-African Gariep Belt, Namibia. Chemical Geology, 130, 101–121.

Ge, W., Li, X., Li, Z., Zhou, H., Wang, J., Lee, C.-Y., 2000. Longsheng Ophiolite in North Guangxi revisited (in Chinese with English abstract). Acta Petrologica Sinica 16 (1), 111–118.

Gee, D.G., Pease, V., 2005. The Neoproterozoic Timanide Orogen of eastern Baltica: introduction. In: Gee, D.G., Pease, V. (Eds.), The Neoproterozoic Timanide Orogen of Eastern Baltica. Geological Society of London Memoir, vol. 30. The Geological Society of London, London, pp. 1–3.

Goddéris, Y., Donnadieuu, Y., Nedelec, A., Dupre, B., Dessert, C., Grard, A., Ramstein, G., Francois, L.M., 2003. The Sturtian "snowball" glaciation: fire and ice. Earth and Planetary Science Letters 211, 1–12.

Goodge, J.W., Williams, I.S., Myrow, P., 2004. Provenance of Neoproterozoic and lower Paleozoic siliciclastic rocks of the central Ross Orogen, Antarctica: detrital record of rift-, passive- and active-margin sedimentation. Geological Society of America Bulletin 116, 1253–1279.

Goodwin, A.M., 1991. Precambrian Geology: The Dynamic Evolution of the Continental Crust. Academic Press, London. 666 pp.

Goscombe, B., Gray, D., Hand, M., 2005. Extrusional tectonics in the core of a transpressional orogen; the Kaoko Belt, Namibia. Journal of Petrology 46, 1203–1241.

Hamilton, W., Myers, W.B., 1967. The Nature of Batholiths. 554-C. US Geological Survey, Washington DC.

Handke, M.J., Tucker, R.D., Ashwal, L.D., 1999. Neoproterozoic continental arc magmatism in west-central Madagascar. Geology 27, 351–354.

Hanson, R.E., Wardlaw, M.S., Wilson, T.J., Mwale, G., 1993. U–Pb zircon ages from the Hook granite massif and Mwembeshi dislocation: constraints on Pan-African deformation, plutonism, and transcurrent shearing in Central Zambia. Precambrian Research 63 (3–4), 189–209.

Hargrove, U.S., Hanson, R.E., Martin, M.W., Blenkinsop, T.G., Bowring, S.A., Walker, N., Munanyiwa, H., 2003. Tectonic evolution of the Zambezi orogenic belt: geochronological, structural, and petrological constraints from northern Zimbabwe. Precambrian Research 123, 159–186.

Hawkesworth, C.J., Kemp, A.I.S., 2006. Evolution of the continental crust. Nature 443, 811–817.

Hefferan, K.P., Admou, H., Karson, J.A., Saquaque, A., 2000. Anti-Atlas (Morocco) role in Neoproterozoic Western Gondwana reconstruction. Precambrian Research 103, 89–96.

Hibbard, J.P., Stoddard, E.F., Secor, D.T., Dennis, A.J., 2002. The Carolina Zone: overview of Neoproterozoic to early Paleozoic per-Gondwanan terranes along the eastern Flank of the southern Appalachians. Earth-Science Reviews 57, 299–339.

Hong, D., Zhang, J., Wang, T., Wang, S., Xie, X., 2004. Continental crustal growth and the supercontinental cycle: evidence from the Central Asian Orogenic Belt. Journal of Asian Earth Sciences 23, 799–813.

Jacobs, J., Thomas, R.J., 2004. Himalayan-type indenter-escape tectonics model for the southern part of the late Neoproterozoic-early Paleozoic East African-Antarctic orogen. Geology 32, 721–724.

Jahn, B.-M., Caby, R., Monie, P., 2001. The oldest UHP eclogites of the world: age of UHP metamorphism, nature of protoliths and tectonic implications. Chemical Geology 178, 143–158.

Jarrar, G., Stern, R.J., Saffarini, G., Al-Zubi, H., 2003. Late- and post-orogenic Neoproterozoic intrusions of Jordan: implications for crustal growth in the northernmost segment of the East African Orogen. Precambrian Research, 123, 295–320.

Johnson, P.R., Woldehaimanot, B., 2003. Development of the Arabian–Nubian Shield: perspectives on accretion and deformation in the East African orogen and the assembly of Gondwana. In: Yoshida, M., Windley, B.F., Dasgupta, S. (Eds.), Proterozoic East Gondwana: Supercontinent Assembly and BreakupSpecial Publication, vol. 206. Geological Society, London, pp. 289–325. London.

Johnson, P.R., Kattan, F.H., 2007. Geochronological dataset of Precambrian rocks in the Arabian Peninsula: a catalogue of U–Pb, Rb–Sr, Ar–Ar, and Sm–Nd ages. Saudi Geological Survey Open File report SGS-OF-2007-3. 40 p.

Johnson, S.D., Poujol, M., Kisters, A.F.M., 2006. Constraining the timing and migration of collisional tectonics in the Damara Belt, Namibia: U–Pb zircon ages for the syntectonic Salem-type Stinkbank granite. South African Journal of Geology 109, 611–624.

Jull, M., Kelemen, P.B., 2001. On the conditions for lower crustal convective instability. Journal of Geophysical Research 106 (B4), 6423–6446.

Jung, S., Mezger, K., 2001. Geochronology in migmatites—a Sm–Nd, U–Pb and Rb–Sr study from the Proterozoic Damara belt (Namibia): implications for polyphase development in high-grade terranes. Journal of Metamorphic Geology 19, 77–97.

Jung, S., Mezger, K., Hoernes, S., 1998. Petrology and geochemistry of syn- to post-collisional A-type granites—a major and trace element and Nd–Sr–Pb–O isotopic study from the Proterozoic Damara Belt, Namibia. Lithos 45, 147–175.

Karlstrom, K.E., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W., Ahall, K.-I., 1999. Refining Rodinia: geologic evidence for the Australia-western U.S. connection in the Proterozoic. GSA Today 9 (10), 1–7.

Katz, O., Beyth, M., Miller, N., Stern, R.J., Avigad, D., Basu, A., Anbar, A., 2004. A late Neoproterozoic (~630 Ma) boninitic suite from southern Israel: implications for the consolidation of Gondwanaland. Earth and Planetary Science Letters 218, 475–490.

Kennedy, W.Q., 1964. The structural differentiation of Africa in the Pan-African (±500 m.y.) tectonic episode. Leeds University Research Institute for African Geology Annual Report 8, 48–49.

Key, R.M., Liyungu, A.K., Njamu, F.M., Somwe, V., Mosley, P., Armstrong, R.A., 2001. The western arm of the Lufilian Arc in NW Zambia and its potential for copper mineralization. Journal of African Earth Sciences 33, 503–528.

Khain, E.V., Bibikova, E.V., Kröner, A., Zhuravlev, D.Z., Sklyarov, E.V., Fedotova, A.A., Kravchenko-Berezhnoy, I.R., 2002. The most ancient ophiolite of the Central Asian fold belt: U–Pb and Pb–Pb zircon ages for the Dunzhugur Complex, Eastern Sayan, Siberia, and geodynamic implications. Earth and Planetary Science Letters 199, 311–325.

Khain, E.V., Bibikova, E.V., Salnikova, E.B., Kröner, A., Gibsher, A.S., Didenko, A.N., Degtyarev, K.E., Fedotova, A.A., 2003. The Palaeo-Asian ocean in the Neoproterozoic and early Palaeozoic: new geochronologic data and palaeotectonic reconstructions. Precambrian Research 122, 329–358.

Kim, S.W., Oh, C.W., Ryu, I.-C., Williams, I.S., Sajeev, K., Santosh, M., Rajesh, V.J., 2006. Neoproterozoic bimodal volcanism in the Okcheon Belt, South Korea, and its comparison with the Nanhua Rift, South China: implications for rifting in Rodinia. Journal of Geology 114, 717–733.

Knoll, A.H., 2000. Learning to tell Neoproterozoic time. Precambrian Research 100, 3–20.

Knoll, A.H., Walter, M.R., Narbonne, G.M., Christie-Blick, N., 2006. The Ediacaran period: a new addition to the geologic time scale. Lethaia 39, 13–30.

Koralay, O.E., Dora, O.Ö., Chen, F., Satir, M., Candan, O., 2004. Geochemistry and geochronology of orthogneiss in the Derbent (Alasehir) Area, eastern part of the Ödemis-Kiraz Submassif, Menderes Massif: Pan-African magmatic activity. Turkish Journal of Earth Science 13, 37–61.

Kovalenko, V.I., Yarmolyuk, V.V., Kovach, V.P., Kozakov, I.K., Salnikova, E.B., Larin, A.M., 2004. Isotope provinces, mechanisms of generation and sources of the continental crust in the Central Asian mobile belt: geological and isotopic evidence. Journal of Asian Earth Sciences 23, 605–627.

Krogh, T.E., Kamo, S.L., Sharpton, V.L., Marin, L.E., Hildebrand, A.R., 1993. U–Pb ages of single shocked zircons linking distal K/T ejecta to the Chicxulub crater. Nature 366, 731–734.

Kröner, A., 2001. DuToit Memorial Lecture 1999: The Mozambique belt of East Africa and Madagascar: significance of zircon and Nd model ages for Rodinia and Gondwana supercontinent formation and dispersal. South African Journal of Geology 104, 151–166.

Kröner, A., Stern, R.J., 2004. Africa: Pan-African orogeny. In: Selley, R., Cocks, L.R.M., Plimer, I.R. (Eds.), Encyclopedia of Geology. Elsevier, pp. 1–12.

Kuzmichev, A.B., Bibikova, E.V., Zhuravlev, D.Z., 2001. Neoproterozoic (~800 Ma) orogeny in the Tuva-Mongolia Massif (Siberia): island arc-continent collision at the northeast Rodinia margin. Precambrian Research 110, 109–126.

Lapierre, H., Bosch, D., Tardy, M., Struik, L.C., 2003. Late Paleozoic and Triassic plume-derived magmas in the Canadian Cordillera played a key role in continental crust growth. Chemical Geology 201, 55–89.

Lee, S.R., Cho, M., 1995. Tectonometamorphic evolution of the Chuncheon amphibolite, central Gyeonggi massif, South Korea. Journal of Metamorphic Geology 13, 315–328.

Lee, S.R., Cho, M., Cheong, C.-S., Kim, H., Wingate, M.T.D., 2003a. Age, geochemistry, and tectonic significance of Neoproterozoic alkaline granitoids in the northwestern margin of the Gyeonggi massif, South Korea. Precambrian Research 122, 297–310.

Lee, S.R., Cho, M., Hwang, J.H., Lee, B.-J., Kim, Y.-B., Kim, J.C., 2003b. Crustal evolution of the Gyeonggi massif, South Korea: Nd isotopic evidence and implications for continental growths of East Asia. Precambrian Research 121, 25–34.

Leite, J.A.D., Hartmann, L.A., McNaughton, N.J., Chemale Jr., F.C., 1998. SHRIMP U/Pb Zircon geochronology of Neoproterozoic juvenile and crustal-reworked terranes in southernmost Brazil. International Geology Review 40, 688–705.

Li, X.H., McCulloch, M.T., 1996. Secular variation in the Nd isotopic composition of Neoproterozoic sediments from the southern margin of the Yangtze Block: evidence for a Proterozoic continental collision in southeast China. Precambrian Research 76, 67–76.

Li, Z.X., Li, X.H., Kinney, P.D., Wang, J., 1999. The breakup of Rodinia: did it start with a mantle plume beneath South China? Earth and Planetary Science Letters 173, 171–181.

Li, X.-H., Li, Z.-X., Ge, W., Zhou, H., Li, W., Liu, Y., Wingate, M.T.D., 2003. Neoproterozoic granitoids in South China: crustal melting above a mantle plume at ca. 825 Ma? Precambrian Research 122, 45–83.

Li, Z.X., Evans, D.A.D., Zhang, S., 2004. A 90° spin on Rodinia: possible causal links between the Neoproterozoic supercontinent, superplume, true polar wander and low-latitude glaciation. Earth and Planetary Science Letters 220, 409–421.

Li, X.H., Su, L., Chung, S.-L., Li, Z.X., Liu, Y., Song, B., Liu, D.Y., 2005. Formation of the Jinchuan ultramafic intrusion and the world's third largest Ni–Cu sulfide deposit: associated with the ~825 Ma South China mantle plume? Geochemistry, Geophysics, Geosystems 6 (11). doi:10.1029/2005GC001006.

Li, Z.X., Bogdanova, S.V., Collins, A.S., Davidson, A., DeWael, B., Ernst, R.E., Fitzsimons, I.C.W., Fuck, R.A., Gladkochub, D.P., Jacobs, J., Karlstrom, K.E., Lu, S., Natapov, L.M., Pease, V., Pisarevsky, S.A., Thrane, K., Vernikovsky, V., in press. Assembly, configuration, and break-up history of Rodinia: a synthesis. Precambrian Research.

Ling, W., Gao, S., Zhang, B., Li, H., Liu, Y., Cheng, J., 2003. Neoproterozoic tectonic evolution of the northwestern Yangtze craton, South China: implications for amalgamation and breakup of the Rodinia supercontinent. Precambrian Research 122, 111–140.

Linnemann, U., McNaughton, N.J., Romer, R.L., Gehmlich, M., Drost, K., Tonk, C., 2004. West African provenance for Saxo-Thuringia (Bohemian Massif): did Armorica ever leave pre-Pangean Gondwana?-U/Pb-SHRIMP zircon evidence and the Nd isotopic record. International Journal of Earth Sciences 93, 683–705.

Liégeois, J.P., Berza, T., Tatu, M., Duchesne, R., 1996. The Neoproterozoic Pan-African basement from the Alpine lower Danubian nappe system (Southern Carpathians, Romania). Precambrian Research 80, 281–301.

Liégeois, J.P., Latouche, L., Boughrara, M., Navez, J., Guiraud, M., 2003. The LATEA metacraton (Central Hoggar, Tuareg Shield, Algeria): behaviour of an old passive margin during the Pan-African orogeny. Journal of African Earth Sciences 37, 133–159.

Loos, S., Reischmann, T., 1999. The evolution of the southern Menderes Massif in SW Turkey as revealed by zircon dating. Journal of the Geological Society of London 156, 1021–1030.

Lécorché, J.P., Bronner, G., Dallmeyer, R.D., Rocci, G., Roussel, J., 1991. The Mauritanide orogen and its northen extensions (western Sahara and Zemmour), West Africa. In: Dallmeyer, R.D., Lécorché, J.P. (Eds.), The West African Orogens and Circum-Atlantic Correlatives. Springer-Verlag, Berlin, pp. 187–227.

Malitch, K.N., 2004. Osmium isotope constraints on contrasting sources and prolonged melting in the Proterozoic upper mantle: evidence from ophiolitic Re–Os sulfides and Ru–Os–Ir alloys. Chemical Geology 208, 157–173.

Mann, P., Taira, A., 2004. Global tectonic significance of the Solomon Islands and Ontong Java Plateau. Tectonophysics 389, 137–190.

Maruyama, S., Liou, J.G., 1998. Ultrahigh-pressure metamorphism and its significance on the Proterozoic–Phanerozoic boundary. The Island Arc 7, 6–35.

McConnell, R.B., 1972. Geological development of the rift system of eastern Africa. Geological Society America Bulletin 83, 2549–2572.

Meert, J.G., 2003. A synopsis of events related to the assembly of eastern Gondwana. Tectonophysics 362, 1–40.

Meert, J.G., Torsvik, T.H., 2003. The making and unmaking of a supercontinent: Rodinia revisited. Tectonophysics 375, 261–288.

Meffre, S., Direen, N.G., Crawford, A.J., Kamenetsky, V., 2004. Mafic volcanic rocks on King Island, Tasmania: evidence of 579 Ma break-up in East Gondwana. Precambrian Research 135, 177–191.

Melcher, F., Meisel, T., Puhl, J., Koller, F., 2002. Petrogenesis and geotectonic setting of ultramafic rocks in the Eastern Alps: constraints from geochemistry. Lithos 65, 69–112.

Menzies, M.A., Klemperer, S., Ebinger, C., Baker, J., 2002. Characteristics of volcanic rifted margins. In: Menzies, M.A., Klemperer, S., Ebinger, C., Baker, J. (Eds.), Volcanic Rifted Margins. Geological Society of America Special Publication, vol. 362. Boulder CO, pp. 1–14.

Moores, E.M., 1991. Southwest U.S.-East Antarctica (SWEAT) connection: a hypothesis. Geology 19, 425–428.

Mukasa, S.B., Dalziel, I.W.D., 2000. Marie Byrd Land, West Antarctica: evolution of Gondwana's Pacific margin constrained by zircon U–Pb geochronology and feldspar common-Pb isotopic compositions. Geological Society of America Bulletin 112, 611–627.

Murphy, J.B., Nance, R.D., 1991. Supercontinent model for the contrasting character of late Proterozoic orogenic belts. Geology 19, 469–471.

Murphy, J.B., Nance, R.D., 2003. Do supercontinents introvert or extrovert? Sm–Nd isotope evidence. Geology 31, 873–876.

Nance, R.D., Murphy, J.B., Strachan, R.A., Keppie, J.D., Gutiérrez-Alonso, G., Fernández-Suárez, J., Quesada, C., Linnemann, U., D'Lemos, R.S., Pisarevsky, S.A., in press. Neoproterozoic–early Paleozoic paleogeography of the peri-Gondwanan terranes: Amazonian versus West African connections. In: Nasser, E., Liégeois, J.-P. (Eds.), The Boundaries of the West African Craton. Geological Society of London Special Publication, London.

Neubauer, F., 2002. Evolution of late Neoproterozoic to early Paleozoic tectonic elements in Central and Southeast European Alpine mountain belts: review and synthesis. Tectonophysics 352, 87–103.

Nyblade, A.A., Brazier, R.A., 2002. Precambrian lithospheric controls on the development of the East African rift system. Geology 30, 755–758.

O'Driscoll, C.F., Dean, M.T., Wilton, D.H.C., Hinchley, J.G., 2001. The Burin Group: A Late Neoproterozoic Ophiolite Containing Shear-zone Hosted Mesothermal-style Gold Mineralization in the Avalon Zone, Burin Peninsula, Newfoundland, Newfoundland Department of Mines and Energy.

Paixao, M.A.P., Nilson, A.A., 2001. Characterization of the Moho Transition Zone in the Quatipuru Complex—a slab of oceanic lithosphere in southern Pará, Brazil—and implications for the tectonic evolution of the Araguaia Belt (abstract). International Conference on the Geology of Oman, Muskat (Oman), p. 68.

Pedrosa-Soares, A.C., Vidal, P., Leonardos, O.H., deBrito-Neves, B.B., 1998. Neoproterozoic oceanic remnants in eastern Brazil: further evidence and refutation of an exclusively ensialic evolution for the Araçuaí-West Congo orogen. Geology 26, 519–522.

Pedrosa-Soares, A.C., Noce, C.M., Wiedemann, C.M., Pinto, C.P., 2001. The Araçuaí-West Congo orogen in Brazil: an overview of a confined orogen formed during Gondwanaland assembly. Precambrian Research 110, 307–323.

Penaye, J., Kröner, A., Toteu, S.F., Schmus, W.R.V., Doumnang, J.-C., 2006. Evolution of the Mayo Kebbi region as revealed by zircon dating: an early (ca. 740 Ma) Pan-African magmatic arc in southwestern Chad. Journal of African Earth Sciences 44, 530–542.

Pfänder, J., Kröner, A., 2005. Tectonomagmatic evolution, age and emplacement of the Agardagh Tes-Chem ophiolite in Tuva, Central Asia: crustal growth by island arc accretion. In: Kusky, T.E. (Ed.), Precambrian Ophiolites and Related Rocks. Elsevier, Amsterdam, pp. 207–222.

Pfänder, J.A., Jochum, K.P., Kozakov, I., Kröner, A., Todt, W., 2002. Coupled evolution of back-arc and island arc-like mafic crust in the late-Neoproterozoic Agardagh Tes-Chem ophiolite, Central Asia: evidence from trace element and Sr–Nd–Pb isotope data. Contributions to Mineralogy and Petrology 143, 154–174.

Preiss, W.V., 2000. The Adelaide Geosyncline of South Australia and its significance in Neoproterozoic continental reconstruction. Precambrian Research, 100, 21–63.

Ramezani, J., Tucker, R.D., 2003. The Saghand region, central Iran: U–Pb geochronology, petrogenesis and implications for Gondwana tectonics. American Journal of Science 303, 622–665.

Rasmussen, B., Fletcher, I.R., 2004. Zirconolite: a new U–Pb chronometer for mafic igneous rocks. Geology 32, 785–788.

Roobol, M.J., Ramsay, C.R., Jackson, N.J., Darbyshire, D.P.F., 1983. Late Proterozoic lavas of the Central Arabian Shield: evolution of an ancient volcanic arc system. Journal of the Geological Society of London 140, 185–202.

Rozendaal, A., Gresse, P.G., Scheepers, R., LeRoux, J.P., 1999. Neoproterozoic to early Cambrian crustal evolution of the Pan-African Saldania Belt, South Africa. Precambrian Research 97, 303–323.

Rudnick, R.A., Gao, S., 2003. 3.01 composition of the continental crust. Treatise on Geochemistry. Elsevier, pp. 1–64.

Samson, S.D., Inglis, J.D., D'Lemos, R.S., Admous, H., Blichert-Toft, J., Hefferean, K., 2004. Geochronological, geochemical, and Nd–Hf isotopic constraints on the origin of Neoproterozoic plagiogranites in the Tasriwine ophiolite, Anti-Atlas, Morocco. Precambrian Research 135, 133–147.

Savov, I., Ryan, J., Haydoutov, I., Schijf, J., 2001. Late Precambrian Balkan–Carpathian ophiolite—a slice of the Pan-African ocean crust? Geochemical and tectonic insights from the Tcherni Vrah and Deli Jovan massifs, Bulgaria and Serbia. Journal of Volcanology and Geothermal Research 110, 299–318.

Scarrow, J.H., Pease, V., Fleuelot, C., Dushin, V., 2001. The late Neoproterozoic Enganepe ophiolite, Polar Urals, Russia: an extension of the Cadomian arc? Precambrian Research 110, 255–275.

Schandelmeier, H., Abdel-Rahman, E.M., Wipfler, E., Küster, D., Utke, A., Matheis, G., 1994. Late Proterozoic magmatism in the Nakasib suture, Red Sea Hills, Sudan. Journal of Geological Society, London 151, 485–497.

Scheepers, R., Armstrong, R., 2002. New U–Pb SHRIMP zircon ages of the Cape Granite Suite: implications for the magmatic evolution of the Saldania Belt. South African Journal of Geology 105, 241–256.

Scholl, D.W., von Huene, R., 2006. Estimating the global volume of deeply recycled continental crust at continental collision zones. Eos Transaction AGU 87 (52), V43F-02 (abstract).

Scholl, D.W., von Huene, R., in press. Crustal recycling at modern subduction zones applied to the past-issues of growth and preservation of continental basement, mantle geochemistry, and supercontinent reconstruction. In: Hatcher, R.D., Carlson, M.P. McBride, J.H., Catalan, J.M. (Eds.), The 4D Framework of Continental Crust. Geological Society of America Memoir 200, Boulder CO.

Sears, J.W., Price, R.A., 2003. Tightening the Siberian connection to western Laurentia. Geological Society of America Bulletin 115, 943–953.

Seghedi, A., Berza, T., Iancu, V., Maruntiu, M., Oaie, G., 2005. Neoproterozoic terranes in the Moesian basement and in the Alpine Danubian Nappes of the South Carpathians. Geologica Belgica 8 (4), 4–19.

Shen, W., Xu, S., Gao, J., Yang, Z., Yang, Q., 2002. Sm–Nd dating and Nd–Sr isotopic characteristics of the Shimian ophiolite suite, Sichuan Province. Chinese Science Bulletin 47, 1897–1901.

Shen, W., Zou, H., Chu, Z., Zhou, X., Yang, J., Wang, Y., 1992. Nd, Sr, and O isotopic study on Fuchuan ophiolite suite in Ahui Province (in Chinese with English abstract). Scientia Geologica Sinica 10, 333–341.

Sommer, H., Kröner, A., Hauzenberger, C., Muhongo, S., Wingate, M.T.D., 2003. Metamorphic petrology and zircon geochronology of high-grade rocks from the central Mozambique Belt of Tanzania: crustal recycling of Archean and Palaeoproterozoic material during the Pan-African orogeny. Journal of Metamorphic Geology 21, 915–934.

Stern, R.J., 1994. Arc assembly and continental collision in the Neoproterozoic East African orogen: implications for the consolidation of Gondwanaland. Annual Reviews of Earth and Planetary Sciences 22, 319–351.

Stern, R.J., 2002a. Crustal evolution in the East African Orogen: a neodymium isotopic perspective. Journal of African Earth Sciences 34, 109–117.

Stern, R.J., 2002b. Subduction zones. Reviews of Geophysics 40, 1012. doi:10.1029/2001RG000108.

Stern, R.J., 2004. Subduction initiation: spontaneous and induced. Earth and Planetary Science Letters 226, 275–292.

Stern, R.J., 2005. Evidence from ophiolites, blueschists, and ultra-high pressure metamorphic terranes that the modern episode of subduction tectonics began in Neoproterozoic time. Geology 33, 557–560.

Stern, R.J., 2007. When did plate tectonics begin on Earth? Theoretical and empirical constraints. Chinese Bulletin of Science 52, 578–591.

Stern, R.J., Johnson, P.J., Kröner, A., Yibas, B., 2004. Neoproterozoic ophiolites of the Arabian–Nubian Shield. In: Kusky, T. (Ed.), Precambrian Ophiolites. Elsevier, Amsterdam, pp. 95–128.

Stern, R.J., Avigad, D., Miller, N., Beyth, M., in press. From Volcanic Winter to Snowball Earth: An alternative explanation for Neoproterozoic Biosphere Stress. In: Dilek, Y., Muehlkenbachs K., (Eds.), Links Between Geological Processes, Microbial Activities, and Evolution of Life. Springer.

Stewart, J.H., 2007. World Map Showing Surface and Subsurface Distribution, and Lithologic Character of Middle and Late Neoproterozoic Rocks. 2007-1087, U.S. Geological Survey.

Suayah, I.B., Miller, J.S., Miller, B.V., Bayer, T.M., Rogers, J.J.W., 2006. Tectonic significance of late Neoproterozoic granites from the Tibesti massif in southern Libya inferred from Sr and Nd isotopes and U–Pb zircon data. Journal of African Earth Sciences 44, 561–570.

Sultan, M., Tucker, R.D., Alfy, Z.E., Attia, R., Ragab, A.G., 1994. U–Pb (zircon) ages for the gneissic terrane west of the Nile, southern Egypt. Geologiische Rundschau 83, 514–522.

Tack, L., Wingate, M.T.D., Liégeois, J.-P., Fernandez-Alonso, M., Deblond, M., 2001. Early Neoproterozoic magmatism (1000–910 Ma) of the Zadinian and Mayumbian Groups (Bas-Congo): onset of Rodinia rifting at the western edge of the Congo craton. Precambrian Research 110, 277–306.

Tadesse, T., Hoshino, M., Sawada, Y., 1999. Geochemistry of low-grade metavolcanic rocks from the Pan-African of the Axum area, northern Ethiopia. Precambrian Research 99, 101–124.

Tatsumi, Y., 2005. The subduction factory: how it operates in the evolving Earth. GSA Today 15 (7), 4–10.

Teklay, M., 2006. Neoproterozoic arc-back-arc system analog to modern arc-back-arc systems: Evidence from tholeiite–boninite association, serpentinite mudflows and across-arc geochemical trends in Eritrea, southern Arabian–Nubian shield. Precambrian Research 145, 81–92.

Thomas, R.J., Chevallier, L.P., Gresse, P.G., Harmer, R.E., Eglington, B.M., Armstrong, R.A., DeBeer, C.H., Martini, J.E.J., de Kock, G.S., Macey, P.H., Ingraham, B.A., 2002. Precambrian evolution of the Sirwa Window, Anti-Atlas Orogen, Morocco. Precambrian Research 118, 1–57.

Thorpe, R.S., 1978. Tectonic emplacement of ophiolitic rocks in the Precambrian Mona Complex of Anglesey, North Wales. Geological Magazine, 130, 85–91.

Torsvik, T.H., 2003. The Rodinia jigsaw puzzle. Science 300, 1379–1381.

Torsvik, T.H., Carter, L.M., Ashwal, L.D., Bhushan, S.K., Prandt, M.K., Jamtvelt, B., 2001. Rodinia refined or obscured: paleomagnetism of the Malani igneous suite (NW India). Precambrian Research 108, 319–333.

Toteu, S.F., Penaye, J., Djomani, Y.P., 2004. Geodynamic evolution of the Pan-African belt in central Africa with special reference to Cameroon. Canadian Journal of Earth Science 41, 73–85.

Tsige, L., 2006. Metamorphism and gold mineralization of the Kenticha-Katawicha area: Adola belt, southern Ethiopia. Journal of African Earth Sciences 45, 16–32.

Tucker, R.D., Pharaoh, T.C., 1991. U–Pb zircon ages for late Precambrian igneous rocks in southern Britain. Journal of the Geological Society, London 148, 435–443.

Ustaömer, P.A., Mundil, R., Renne, P.R., 2005. U/Pb and Pb/Pb zircon ages for arc-related intrusions of the Bolu Massif (W Pontides, NW Turkey): evidence for late Precambrian (Cadomian) age. Terra Nova 17, 215–223.

Van Schmus, W.R., Oliveira, E.P., da Silva Filho, A.F., Toteu, S.F., Pennaye, J., Guimaraes, I.P., in press. Proterozoic Links between the Borborema Province, NE Brazil, and the Central African Fold Belt. In: Pankhurst, R.J., Trouw, R.A.J., BritoNeves B.B., de Wit, M.J. (Eds.), West Gondwana: pre-Cenozoic correlations across the South Atlantic region. Geological Society, London, London.

Vauchez, A., Egydio-Silva, M., Babinski, M., Tommasi, A., Uhlein, A., Liu, D., 2007. Deformation of a pervasively molten middle crust: insights from the Neoproterozoic Ribeira-Araçuaí orogen (SE Brazil). Terra Nova 19, 278–286.

Vernikovsky, V.A., Vernikovskaya, A.E., 2001. Central Taimyr accretionary belt (Arctic Asia): Meso-Neoproterozoic tectonic evolution and Rodinia breakup. Precambrian Research 110, 127–141.

Vernikovsky, V.A., Vernikovskaya, A.E., Chernykh, A.I., Sal'nikova, E.B., Kotov, A.B., Kovach, V.P., Yakovleva, S.Z., Fedoseenko, A.M., 2001. Porozhnaya granitoids of the Enisei Ophiolite Belt: indicators of Neoproterozoic events on the Enisei Ridge. Doklady Earth Sciences 381A (9), 1043–1046.

Vernikovsky, V.A., Vernikovsky, A.E., Sal'nikova, E.B., Kotov, A.B., Kovach, V.P., 2003. Neoproterozoic accretion-collisional events on the western margin of the Siberian Craton: new geological and geochronological evidence from the Yenisey Ridge. Tectonophysics 375, 147–168.

Von Huene, R., Scholl, D.W., 1991. Observations at convergent margins concerning sediment subduction, subduction erosion, and the growth of continental crust. Reviews of Geophysics 29, 279–316.

Wang, J., Li, Z.-X., 2003. History of Neoproterozoic rift basins in South China: implications for Rodinia break-up. Precambrian Research 122, 141–158.

Wang, X., Zhou, J., Qiu, J., Gao, J., 2004a. Comment on "Neoproterozoic granitoids in South China: crustal melting above a mantle plume at ca. 825 Ma?" by Xian-Hua Li et al. [Precambrian Res. 122(2003) 45–83]. Precambrian Research 132, 401–403.

Wang, X., Zhou, J., Qiu, J., Gao, J., 2004b. Geochemistry of the Meso- to Neoproterozoic basic–acid rocks from Hunan Province, South China: implications for the evolution of the western Jiangnan orogen. Precambrian Research 135, 79–103.

Wendorff, M., 2005. Evolution of Neoproterozoic–Lower Palaeozoic Lufilian arc, Central Africa: a new model based on syntectonic conglomerates. Journal of the Geological Society of London 162, 5–8.

Windley, B.F., 2003. Continental growth in the Proterozoic: a global perspective In: M., M., Yoshida, Windley, B.F., Dasgupta, S. (Eds.), Proterozoic East Gondwana: Supercontinent Assembly and Break-upSpecial Publication Geological Society of London, vol. 206, pp. 23–34. London.

Windley, B.F., Alexeiev, D., Xiao, W., Kröner, A., Badarach, G., 2007. Tectonic models for accretion of the Central Asian Orogenic Belt. Journal of the Geological Society of London 164, 31–47.

Wingate, M.T.D., Campbell, I.H., Compston, W., Gibson, G.M., 1998. Ion microprobe U–Pb ages for Neoproterozoic basaltic magmatism in south-central Australia and implications for the breakup of Rodinia. Precambrian Research 87, 135–159.

Wingate, M.T.D., Pisarevsky, S.A., Evans, D.A.D., 2002. Rodinia connections between Australia and Laurentia: no SWEAT, no AUSWUS? Terra Nova 14, 121–128.

Wortman, G.L., Samson, S.D., Hibbard, J.P., 2000. Precise U–Pb zircon constraints on the earliest magmatic history of the Carolina Terrane. Journal of Geology 108, 321–338.

Xu, B., Jian, P., Zheng, H., Zou, H., Zhang, L., Liu, D., 2005. U–Pb zircon geochronology and geochemistry of Neoproterozoic volcanic rocks in the Tarim Block of northwest China: implications for the breakup of Rodinia supercontinent and Neoproterozoic glaciations. Precambrian Research 136, 107–123.

Yakubchuk, A., 2004. Architecture and mineral deposit setting of the Altaid orogenic collage: a revised model. Journal of Asian Earth Sciences 23, 761–779.

Yakubchuk, A.S., Nikishin, A.M., Ishiwatari, A., 1994. A late Proterozoic ophiolite pulse. Proc. 29th Int'l. Geol. Congr., Part D, pp. 273–286.

Yan, Q., Hanson, A.D., Wang, Z., Druschke, P.A., Yan, Z., Wang, T., Liu, D., Song, B., Jian, P., Zhou, H., Jiang, C., 2004. Neoproterozoic subduction and rifting on the northern margin of the Yangtze Plate, China: implications for Rodinia reconstruction. International Geology Review 46, 817–832.

Yunpeng, D., Dingwu, Z., Liang, L., Qi, Z., Zongqing, Z., 1997. Sm–Nd isotopic ages of the Songshugou ophiolite from the east Qinling and its geological significance (in Chinese with English abstract). Regional Geology of China 16 (2), 217–221.

Zakariadze, G.S., Dilek, Y., Adamia, S.A., Oberhänsli, R.A., Karpenko, S.F., 2007. Geochemistry and geochronology of the Neoproterozoic Pan-African Transcaucasian Massif (Republic of Georgia) and implications for island arc evolution of the late Precambrian Arabian–Nubian Shield. Gondwana Research 11, 92–108.

Zhang, Q., Wang, Y., Zhou, G.Q., Qian, Q., Robinson, P.T., 2003. Ophiolites in China: their distribution, ages and tectonic settings. In: Dilek, Y., Robinson, P.T. (Eds.), Ophiolites in Earth History. Geological Society of London, London, pp. 541–566.

Zhang, C.-L., Li, X.-H., Li, Z.-X., Lu, S.-N., Ye, H.M., Li, H.-M., 2007. Neoproterozoic ultramafic–mafic–carbonatite complex and granitoids of northeastern Tarim Block, western China: geochronology, geochemistry and tectonic implications. Precambrian Research 152, 149–169.

Zhou, M.-F., Kennedy, A.K., Sun, M., Malpas, J., Lesher, C.M., 2002a. Neoproterozoic Arc-related mafic intrusions along the northern margin of South China: implications for the accretion of Rodinia. Journal of Geology 110, 611–618.

Zhou, M.-F., Yan, D.-P., Kennedy, A.K., Li, Y., Ding, J., 2002b. SHRIMP U–Pb zircon geochronological and geochemical evidence for Neoproterozoic arc-magmatism along the western margin of the Yangtze Block, South China. Earth and Planetary Science Letters 196, 51–67.

Zhao, J.-X., McCulloch, M.T., Korsch, R.J., 2002c. Characterisation of a plume-related ~800 Ma magmatic event and its implication for basin formation in central-southern Australia. Earth and Planetary Science Letters 121, 349–367.

Zimmer, M., Kröner, A., Jochum, K.P., Reischmann, T., Todt, W., 1995. The Gabal Gerf complex: a Precambrian N-MORB ophiolite in the Nubian Shield, NE Africa. Chemical Geology 123, 29–51.

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