Continental growth in the Proterozoic: a global perspective

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Abstract: During Proterozoic time, growth of the continents took place by the addition of mantle-derived, juvenile material to pre-existing continental blocks. This accretion took place largely within three tectonic environments: (1) most importantly, in accretionary orogens such as the Birimian, the Baltic Shield, the Arabian–Nubian Shield and the early Altaids in Central Asia – these orogens grew largely by the accretion of island arcs, oceanic plateaus, accretionary prisms and ophiolites; (2) in the juvenile parts of collisional orogens as in the Trans-Hudson and Grenville; (3) within supercontinents that underwent rifting and breakup, giving rise to continental flood basalts and mafic dyke swarms. In addition to plate tectonics, the role of plume tectonics is increasingly emphasized as a fundamental process in Earth evolution. A mantle superplume may increase the oceanic spreading rate, the subduction rate and thus the island-arc production rate. It may also be responsible for the formation of a supercontinent, thus preserving the juvenile parts of collisional orogens, and it may be instrumental in the fragmentation of a supercontinent, giving rise to juvenile continental flood basalts. The balance between these processes is still poorly understood, as are calculated growth rates of Proterozoic crust.

Throughout Earth history the development of the continents has been influenced by the formation of two major types of orogens; continent–continent collisional and accretionary (Windley 1995). Modern analogues of the former are the Himalayas and the European Alps, and of the latter the circum-Pacific belts extending from the Japanese islands via Taiwan and the Philippines to Indonesia, and from Alaska via the Cordillera of Canada and the USA to Mexico. These two types are just the ends of a spectrum of orogens because some are intermediate in type, e.g. the western Himalayas contains a major juvenile island arc in Kohistan. The formation of collisional orogens involved little or no crustal growth, but much reactivation of older continental crust. In contrast, accretionary orogens formed by the accretion of oceanic crust, accretionary prisms, subduction-generated island arcs, plume-generated oceanic plateaus, seamounts and oceanic islands, and thus are characterized by much crustal growth of juvenile material derived directly or indirectly from the mantle. Therefore, when considering Proterozoic orogens, we will be largely concerned with accretionary, rather than collisional, types (Windley 1992).

Most juvenile Proterozoic continental crust formed in three tectonic environments.

(1) Accretionary orogens: western and central Baltic Shield (2.50–1.75 Ga), Birimian, West Africa (2.1 Ga), SW USA, Yavapai (1.8–1.6 Ga), Arabian–Nubian Shield (1.0–0.5 Ga), Cadomian in NW Europe (0.6–0.5 Ga), the early (0.75–0.54 Ga) Altaids in Central Asia. Most accretionary orogens, present and past, are defined largely by the geological, structural and geochemical characteristics of particular rock suites that are diagnostic of, for example, island arcs, accretionary prisms and oceanic plateaus.

(2) The juvenile, minor parts of collisional orogens that involved the incorporation of mid-ocean ridge basalt (MORB)-type basalts, island arcs, oceanic plateaus and ocean island basalts, e.g. the 1.92–1.84 Ga Flin Flon Belt in the Trans-Hudson Orogen (Lucas *et al.* 1996), and within the Grenville Orogen (Condie 2001).

(3) In supercontinents or large continental blocks that were undergoing breakup as a result of mantle upwellings at 2.4–2.1, 1.5–1.3 and 1.0–0.6 Ga. Examples of products are: (a) continental flood basalts (Circum-Superior Province, 1.96 Ga; Coppermine River, northern Canada, 1.27 Ga; the Mid-Continent Rift, USA, 1.1 Ga; Kola Peninsula, 0.6 Ga); (b) giant mafic dyke swarms (worldwide at 2.4–2.0 Ga; Gardar, South Greenland, 1.2 Ga; Sudbury–Mackenzie dykes, 1.2 Ga; Grenville dykes 590 Ma).

The aim of this paper is to present the key features of the main components (e.g. island arcs, oceanic plateaus, continental flood basalts) of the Proterozoic continents that provide evidence of crustal growth. Also, some important criteria, such as Nd isotopes and seismic data, are singled out as useful constraints on juvenile growth, and, in the final section, crustal growth rates are reviewed.

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Neodymium (Nd) isotopic mapping

Patchett & Arndt (1986) used Nd isotopes to demonstrate that >80% of the 1.96–1.6 Ga belt that extends from Colorado and Arizona through Michigan and South Greenland to the Baltic Shield, consists of newly differentiated, juvenile material. This belt consists of three provinces that decrease in age from NW–SE (present coordinates): (1) 1.9–1.8 Ga: the Penokean in the Lake Superior region, the Makkovik of Labrador, the Ketilidian of South Greenland and the Svecofennian of the Baltic Shield; (2) 1.8–1.7 Ga: the Yavapai of Arizona and Colorado, the Central Plains Orogen and the Killarney Belt near Lake Huron; (3) 1.7–1.0 Ga: the Matzatzal Belt of Arizona and New Mexico, the Labradorian Orogen, and the Trans-Scandinavian Batholith Belt of Sweden. The juvenile origin of the crust in Texas and Mexico is indicated by εNd values of +3–+5 (Patchett & Ruiz 1989; Ortega-Gutierrez *et al.* 1995).

These conclusions were confirmed by Bennett & DePaolo (1987), who concluded that crustal formation in the northern provinces contained only *c.* 20% of pre-existing crust, and that the southernmost province, most distal from the Archaean nucleus to the north, was derived almost entirely from Proterozoic mantle. In the formation of these provinces an aggregate area of new crust up to 1500 km wide and 5000 km long was accreted in *c.* 300 Ma (Hoffman 1988).

Seismic reflection studies

Over the last decade, seismic reflection profiling by COCORP-type on-land vibrators or BIRPS-type seaborne surveys have provided new insights to the crust–mantle structure of many Precambrian orogens, demonstrating that they are, in general, geometrically similar to modern orogens. Seismic profiling across the Svecofennian, Lewisian (Britain and Ireland) and Trans-Hudson Orogens have revealed that, in all cases, juvenile Palaeo-Proterozoic arc, oceanic and composite terranes have been detached from their lower crust and mantle during accretion to Archaean cratons. Thus, the juvenile terranes seen at the surface are only crustal flakes, imbricated during collisional and post-collisional events (Snyder *et al.* 1996). Korja & Heikkinen (1995) demonstrated that listric shear zones in the Svecofennian Orogen flatten at major detachments at depths of 35–40 and 48 km. This conclusion has major implications for interpretations of crustal thickening by juvenile magmatism, which cannot necessarily be extrapolated to Moho depths (e.g. Luosto 1997). Because Reymer & Schubert (1986) were not able to take account of this possibility in the Arabian–Nubian Shield at that time, they assumed that the juvenile greenstone belts at the surface continued all the way down to the Moho, and therefore they arrived at an anomalously high crustal growth rate. In the eastern Arabian Shield of Yemen, Windley *et al.* (1996) and Whitehouse *et al.* (1998*b*) demonstrated that westerly dipping, Early Precambrian gneiss terranes are imbricated with Pan-African island arcs, and this raises the likely possibility that the eastern shield under Saudi Arabia has a more complicated structure than has so far been realized.

Island arcs

Island arcs are an important component of modern accretionary orogens of the western Pacific, and of equivalent Proterozoic orogens: there are several well-documented Proterozoic examples. The Amisk Collage of the Flin Flon Belt and the Reindeer Zone in Canada is a tectonically dismembered collection of accreted juvenile terranes (Stern *et al.* 1995; Lucas *et al.* 1996; Lewry & Stauffer 1997; Leybourne *et al.* 1997). Between 1.904 and 1.890 Ga, tholeiitic and related calc-alkaline basalt–basaltic andesite and rare high-Ca boninites were dominant. The tholeiitic rocks are similar to modern island-arc tholeiites, having low high field strengths (HFSE) and rare earth element (REE) abundances, and chondrite-normalized light REE depletion to slight enrichment. The boninites have even lower HFSE and REE abundances. Between 1.89 and 1.864 Ga, calc-alkaline andesite–rhyolite and rare shoshonite and trachyandesite erupted with strong arc trace element signatures [e.g. high Th/Nb, La/Nb, strong negative Nb anomalies and large-ion lithophile element (LILE) enrichment], and initial εNd values (+2.3–+4.6) indicate depleted mantle contributions (Stern *et al.* 1995).

In the SW Baltic Shield at 1.69–1.65 Ga, felsic and basaltic lavas have markedly primitive trace element signatures and depleted Nd isotopic compositions, all consistent with derivation in an oceanic island-arc setting, possibly similar to those in the Philippine Sea (Brewer *et al.* 1998). Elsewhere in the Swedish Svecofennian at 1.758 Ga, meta-andesites are chemically similar to primitive modern oceanic island arcs and an εNd value of +4.3 indicates a depleted mantle source (Åhäll & Daly 1989).

The Birimian Orogen of West Africa underwent major crustal growth at *c.* 2.1 Ga, which lasted <50 Ma (Boher *et al.* 1992). Arc-derived andesites, dacites and rhyolites are an important component, and calc-alkaline granites make up almost half the Birimian terranes. However, it is strange that the pillowed tholeiitic basalts are overlain by pelagic cherts, shales and clastic turbidites, because that type of oceanic plate stratigraphy in the modern accretionary orogen of Japan is characteristic not of island arcs but of off-scraped oceanic floor, the sedimentary sequence indicating transport from a ridge to a trench. The detailed study by Béziat *et al.* (2000) demonstrated that calc-alkaline basalts and rhyolites, which are associated with pyroclastics, show classic features of arc magmatic suites, namely LILE (Large Ion Lithophile Element) and Pb enrichment, depleted HFSE patterns and high Ce/Nb and Th/Nb ratios; associated wehrlites and gabbros represent the roots of the island arc.

The Early Proterozoic Ungava Orogen in Canada is well known for the 2.00 Ga Portuniq Ophiolite. An island arc was built on the oceanic crust between 1.90 and 1.86 Ga, and a younger subduction zone gave rise to 1.85–1.83 Ga quartz diorite to granite plutons that intruded the older oceanic crust and arc (Lucas *et al.* 1992).

In French Guiana early continental crust accretion is characterized by the formation of volcanic centres with calc-alkaline tuff lavas and pyroclastites associated with plutonic–volcanic complexes composed largely of calc-alkaline tonalite–diorite batholiths dated at 2144 ± 6 and 2115 ± 7 Ma (Vanderhaeghe *et al.* 1998). Also in French Guiana the 2.11 ± 0.09 Ga Inini Greenstone is dominated by calc-alkaline andesite to rhyolite, intruded by plutons of tonalite and trondhjemite, all belonging to an island arc; diamond-bearing, mantle-derived, ultramafic komatiites form part of the volcanic sequence (Capdevila *et al.* 1999).

Further examples of Proterozoic island arcs include: the 1.87 Ga South Harris igneous complex, Outer Hebrides, NW Scotland (Whitehouse & Bridgwater 2001), a 1.92 Ga arc in the Nagssugtoqidian Orogen of West Greenland (Whitehouse *et al.* 1998*a*); a 1.742 Ga arc in the Grand Canyon in SW USA (Ilg *et al.* 1996); the Kaourera Island Arc in the Kibaran Belt of central Africa that is tectonically interleaved with the 1.393 Ga Chewore Ophiolite (Johnson & Oliver, 2000), and a variety of arcs formed during the 1.950–1.700 Ga assembly of the West, North and South Australian Cratons (Myers *et al.* 1996). The Cadomian–Avalonian accretionary orogen of NW Europe was constructed from oceanic crust and island arcs in the period of 800–640 Ma, creating the arc-dominated terrane of Avalonia [e.g. in NW Iberia (Fernández-Suárez *et al.* 2000) and in Wales (Bevins *et al.* 1995)].

Neoproterozoic island arcs are common in the Arabian–Nubian Shield, where they were mostly formed in the period of 900–700 Ma. The oldest arcs (900–850 Ma) in Saudi Arabia consist of tholeiitic andesites and are thought to represent young immature island arcs. Thickening and melting of the immature tholeiitic crust caused the formation of more mature island arcs made up of calc-alkaline low- to high-K tonalites, trondhjemites and andesites in the period of 825–730 Ma (Blasband *et al.* 2000). In their plume-oriented model for the Arabian–Nubian Shield, Stein & Goldstein (1996) suggested that an oceanic plateau resisted subduction and that this enabled arcs to develop by subduction on its margins.

A recent development has been the recognition of juvenile island-arc material within deep crustal rocks, such as the enderbite–charnockite suite of the Umba Complex in the Kola Peninsula (Glebovitsky *et al.* 2001).

Oceanic plateaus

Many oceanic plateaus are prominent in the present Pacific Ocean and increasing evidence is accruing that similar plateaus accreted during the Mesozoic–Cenozoic to parts of the circum-Pacific accretionary orogens. An oceanic plateau implies the presence of a mantle plume not related to ridge-subduction tectonics.

Few detailed studies have yet been made of Proterozoic oceanic plateaus, but information is beginning to appear of their presence. Abouchami *et al.* (1990) suggested that many Birimian basalts represent oceanic plateaus, and Boher *et al.* (1992) argued that the island arcs they discovered (referred to above) formed on top of the assumed oceanic plateaus. Stein & Goldstein (1996) interpreted isotopic and geochemical data from the Arabian–Nubian Shield to indicate that a plume head generated an oceanic plateau, which later resisted subduction during convergence and was overprinted with continent-like characteristics. However, a worrying aspect of this work is that neither the authors or subsequent workers have described lithologies, structure and petrology, which would confirm or constrain the isotopic–geochemical model.

In the *c.* 1.9 Ga Flin Flon Belt of Canada, the Sandy Bay Assemblage is a *c.* 3 km thick, monotonous sequence of basalt flows, synvolcanic diabase and gabbro sills. Lucas *et al.* (1996) found that the basalts are geochemically distinct from those of arc and ocean-floor basalts. Their trace element characteristics include strong enrichment in HFSE (Nb, Zr, Ti), light REE enrichment, high Ti/V and low Zr/Nb, juvenile Nd isotopes, and, most importantly, fractionated heavy REE, suggesting the involvement of residual garnet during melting. Lucas *et al.* (1996) followed Stern *et al.* (1995) in proposing an oceanic plateau or oceanic island origin.

The Loch Maree Group in the Lewisian of NW Scotland contains basalts associated with abyssal sediments and ferruginous hydrothermal deposits (Park *et al.* 2001). Geochemistry of the basalts suggests derivation from an oceanic plateau or primitive arc, and the associated Ard Gneiss, which has primitive geochemical patterns, possibly formed by melting of an underplated oceanic plateau.

Continental flood basalts

Continental flood basalts represent major additions to the crust of plume-generated, mantle-derived magmas. The breakup of a 2.5 Ga supercontinent (Windley 1995) was expressed by global mafic magmatism at 2.45 Ga, the remnants of which are so abundant and extensive that they reasonably constitute a large igneous province (Heaman 1997). They include continental flood basalts, mafic dyke swarms, layered mafic–ultramafic intrusions and rift-related alkaline igneous rocks. The volume of this magmatic material certainly rivals that of Mesozoic large igneous provinces. It is known that the fragmentation of the supercontinent of Pangaea was diachronous and lasted for at least 250 Ma. Similarly, the huge number of mafic dykes and associated igneous rocks worldwide that intruded in the period of 2.45–*c.* 2.1 Ga probably resulted from episodic, semi-continuous attempts at further continental breakup. We are only witnessing the failed attempts at such plume-generated continental breakup preserved on exposed continental margins; many other areas are no doubt under sedimentary basins or were destroyed by collisional orogenesis.

Continental flood basalts formed at two periods in the Paleoproterozoic in the northern Baltic Shield. Numerous well-dated, large, layered mafic–ultramafic intrusions were emplaced in the period of 2.4–2.5 Ga (Amelin *et al.* 1995). They vary in composition from lherzolite and olivine gabbronorite to norite, anorthosite and hypersthene diorite, and represent two types of flood basalt series, whose parental magmas were generated in a mantle plume (Amelin & Semenov 1996). Interestingly, Kempton *et al.* (2001) found that granulite-facies xenoliths hosted in Devonian lamprophyres within the same area in the Kola Peninsula represent high-grade equivalents of the continental flood basalts. Also, at 2.4 Ga in the southeastern Baltic Shield, plume-generated komatiitic basalts were erupted in an abortive continental rift (Puchtel *et al.* 1997). In the Kola Peninsula the 1000 km long Pechenga Greenstone Belt has a total tectonic thickness of *c.*16 km (probably thickened by imbrication) and formed for an unknown duration during the period of 2.50–1.80 Ga in a vast continental rift according to Melezhik & Sturt (1994). The rift contains many plume-derived layered gabbro–norite complexes enriched in PGE (Platinum Group Element) and chromite mineralization. Ferropicrites from *c.* 1.98 Ga are associated with major Ni–Cu deposits, most likely derived from Fe-rich, ^{187}Os-enriched mantle plumes (Walker *et al.* 1997).

A second vast continental flood basalt province in the northern Baltic Shield was reported by Puchtel *et al.* (1998). Submarine basalts, reaching up to 4.5 km in thickness, occur within several epicontinental basins that are remnants of a once-continuous cover over earlier basement. These plume-generated rocks occur in an area of *c.* 600 000 sq. km^2 of the northern Baltic Shield and represent a major contribution of juvenile material to the existing continental crust. Sm–Nd mineral and Pb–Pb whole-rock isochron ages of 1975 ± 24 and 1980 ± 57 Ma, respectively, from the upper part, and a SHRIMP U–Pb zircon age of 1976 ± 9 Ma from the lower part of the basalt pile imply a short time span of formation. Uppermost lavas have high $(Nb/Th)_N (= 1.4–2.4)$ and $(Nb/La)_N$ $(= 1.1–1.3)$ values, an εNd (T) value of $+3.2$, and an unradiogenic Pb-isotope composition ($\mu_1 = 8.57$), all comparable with those of modern oceanic plume-derived magmas (oceanic flood basalt and oceanic island basalt). Puchtel *et al.* (1998) also concluded that the estimated Nb/U ratios of 53 ± 4 in the uncontaminated lavas are similar to those found in the modern mantle (*c.* 47), suggesting that by 2.0 Ga a volume of continental crust similar to the present-day value already existed.

Many, now isolated, segments of continental basalts that now rim the Superior Province of northern Quebec probably once belonged to an extensive continental flood basalt province, the best continuous sequence of which is exposed in the Belcher Islands (Legault *et al.* 1994); ages range from 1798 ± 38 (whole-rock Rb–Sr isochron) to 1960 ± 80 Ma (Pb–Pb isochron).

The 1.27 Ga Coppermine River flood basalts in the Northwest Territories of Canada were emplaced during the Mackenzie igneous event, that included the coeval, layered, mafic–ultramafic Muskox Intrusion and the vast Mackenzie Dyke Swarm that radiates from the Coppermine River basalts (Dupuy *et al.* 1992; Griselin *et al.* 1997). The basalts that overlie continental basement comprise *c.* 150 flows, each 4–100 m thick, many of which can be traced laterally for several tens of kilometres (maximum total thickness is 4.7 km). They were emplaced in a short time period of <5 Ma. Petrochemistry is interpreted to indicate that the lowermost lavas were produced by melting in the garnet stability field at a depth >90 km, and probably in a mantle plume beneath the continental lithosphere. Upper lavas were partly contaminated with crustal rocks as the magmas passed through the lower and upper crust.

Mafic dykes

The numbers of mafic dykes and dyke swarms intruded worldwide, especially in the Palaeoproterozoic and Late Neoproterozoic, are too numerous to enumerate; most are recorded in Ernst & Buchan (2001a). Giant, radiating dyke swarms that may be linked with the evolution of mantle plumes were reviewed by Ernst & Buchan (2001a). Many dyking events between 2.4 and 2.0 Ga can be correlated across continents, e.g. between the Canadian and Fennoscandian Shields (Park 1994; Vogel *et al.* 1998). From a study of the Th/Ta and La/Yb ratios of Proterozoic mafic dykes, Condie (1997*a*) concluded that there is an overall shift in composition of dykes from high Th/Ta ratios in the Palaeoproterozoic to low ratios in the Neoproterozoic. This reflects a decrease in the importance of Archaean subcontinental lithospheric sources and an increase in importance of plumes containing enriched mantle components such as recycled sediments and oceanic lithosphere.

The breakup of Baltica from Laurentia is marked by huge doleritic sill complexes in Fennoscandia at 1.27 Ga (Elming & Mattsson 2001). Park *et al.* (1995) proposed that during the breakup of Rodinia a giant, radiating, plume-generated mafic dyke swarm was emplaced at *c.* 780 Ma, and is now fragmented by the separation of western North America and eastern Australia. However, Wingate *et al.* (1998) found that the Gairdiner Dyke Swarm in Australia has a U–Th–Pb baddeleyite–zircon age of 827±6 Ma, *c.* 40 Ma too early.

Finally, the opening of the Iapetus Ocean was marked by the emplacement of the 616 Ma Egersund Dyke Swarm in SW Norway (Bingen *et al.* 1998), and of the 700 km long, plume-generated, radiating 590 Ma Grenville Dyke Swarm in Canada (Seymour & Kumarapeli 1995).

Ophiolites and the ocean floor

Proterozoic ophiolites are not common, but neither are ophiolites common in the Himalayas (collisional orogen) or Japan (accretionary orogen). However, the fact that some examples with a full ophiolite stratigraphy do occur (Anon 1972), indicates that oceanic ridge and subduction processes were in operation back in time to at least the Palaeo-Proterozoic. Although volumetrically small, these ophiolites do provide key information on the role of plate tectonic processes.

The 1.998 Ga Portuniq Ophiolite in the Cape Smith Belt of Canada is situated on the continental margin of an Archaean craton. It consists of two magmatic suites (Scott *et al.* 1992; St-Onge *et al.* 1997). The lower one has a complete ophiolite stratigraphy and its composition is similar to rocks formed at present-day mid-oceanic ridges. The younger suite has sheeted mafic dykes and mafic to ultramafic cumulate rocks that are geochemically similar to tholeiites found in modern plume-generated oceanic islands, such as Hawaii.

The 1.96 Ga Jormua Ophiolite in Finland is situated on the margin of an Archaean craton (Kontinen 1987). The gabbros are similar to those in high-Ti ophiolites dredged from present-day mid-oceanic ridges. The chemistry of the basalts indicates they are comparable to Red-Sea-type basalts (Peltonen *et al.* 1996). The 1.901 Ga Elbow–Athapapuskow Ophiolite in the Flin Flon Belt of Canada has pillow basalts intruded by diabase sills, gabbros and mafic–ultramafic cumulates (Lucas *et al.* 1996).

The 1.73 Ga Payson Ophiolite in Arizona, within the juvenile Yavapai–Mazatzal Orogen, has submarine basalts, sheeted dykes (1–2 km thick) and gabbro, and was erupted upon a magmatic arc (Dann & Bowring 1997). The sheeted dykes are tholeiitic basalts with an island-arc affinity.

There are not many mid-Proterozoic orogenic belts or ophiolites. However, the Zambezi Belt between the Congo and Zimbabwe Cratons contains the remains of the 1.4 Ga (zircon age of a plagiogranite dyke) Chewore Ophiolite, which occurs as discontinuous pods associated with yoderite-bearing whiteschists within a subduction complex. Serpentinites, gabbros, sheeted dolerite dykes and basalts can be recognized at a greenschist grade (Oliver *et al.* 1998). One group of meta-basalts is similar to modern N-MORB, and another resembles modern island-arc basalts. This is the oldest dated remnant of Proterozoic oceanic crust in Africa.

Yakubchuk *et al.* (1994) pointed out that there was an ophiolite pulse in the Neoproterozoic spread over the full age range from 1000 to 570 Ma, with a pronounced concentration at 750 Ma and a lesser pulse at 600 Ma. Ophiolites of 750 Ma and older are common in Arabia, Africa and South America, whereas 600 Ma ophiolites are abundant in Central Asia. A prominent, semi-continuous belt of Neoproterozoic ophiolites extends around the margin of the Siberian Craton from the Taimyr Peninsula to the Yenesei Range, and Eastern Sayan to the Baikal Uplands and Transbaikalia (Khain *et al.* 1997).

In the Arabian–Nubian Shield there are 15 ophiolites with the full Penrose-type stratigraphy (but highly imbricated); eight in Saudi Arabia, five in Sudan and two in Egypt (P. R. Johnson, pers. comm.). They range in age from *c.* 870 to 730 Ma and most are allochthonous sheets situated in arc–arc sutures (Berhe 1997). For example, the 740 Ma Gabal Gerf Ophiolite on the Egypt–Sudan border has pillow lavas and sheeted dykes whose major- and trace-element data, including REE, are indistinguishable from modern high-Ti N-MORB. In fact, this is the only Precambrian ophiolite with N-MORB chemistry (Zimmer *et al.* 1995). Together with the island arcs mentioned above, these ophiolites in the Arabian–Nubian Shield document a major phase of juvenile crustal growth in the Neoproterozoic.

The early development of the Central Asian Mobile Belt or Altaids (Sengör *et al.* 1993) during the Neoproterozoic was dominated by oceanic plate accretion and subduction, and the emplacement of ophiolites. The largest ophiolite in Central Asia (with complete Penrose stratigraphy) is the 300 km long, 20 km wide, 569±21 Ma Bayankhongor Ophiolite in Mongolia, which probably occupies a suture zone between two continental blocks (Buchan *et al.* 2001). Several other ophiolites, e.g. Dariv, Khantaishir and Tuva (southern Russia), have similar ages.

Crustal growth rates

It is important, if possible, to make quantitative estimates of the growth rate of Proterozoic continental crust in specific orogens or parts of the world. Note that the rates are discussed in terms of volumes of continental crust. According to Reymer & Schubert (1986), the average production rate of continental crust during the Phanerozoic has been 1.1±0.5 km^3/a. They went on to calculate and suggest that the crust formation rates during some geological periods and places, e.g. the Arabian–Nubian Shield and the Superior Province, exceeded by far the Phanerozoic rate. However, they assumed, reasonably at that time, that the juvenile greenstone belts on the surface continue all the way to the Moho; this was a necessary assumption in order to arrive at a volume calculation. However, it is now known from gravity data and especially from seismic profiling projects such as LITHOPROBE in Canada, that greenstone belts only continue to a few kilometres depth. For example, the greenstone belts of the Superior Province, which have a thickness of only *c.* 5 km, are situated on a major shear zone below which is reflective deeper crust (Clowes *et al.* 1996). The nearby Kapuskasing Uplift, which provides a complete crustal section, shows that the deeper crust consists of granulite facies mafic, and felsic, gneisses and anorthosite (Percival & West 1994). Gravity anomaly data suggest that the Barberton Greenstone Belt in South Africa does not extend beyond 6 km in depth, and 3 km is most likely (Darracott 1975). Although these examples are from Archaean juvenile greenstone belts, Snyder *et al.* (1996) concluded from their overall survey of seismic reflection profiles in the Early Proterozoic Svecofennian, Lewisian and Trans-Hudson Orogens that these juvenile terranes were detached from their lower crust during accretion to Archaean margins, and that the juvenile terranes are only crustal flakes, imbricated and internally deformed. This idea is supported by the discovery that Early Precambrian gneiss terranes in the eastern Arabian Shield of Yemen are imbricated with Neoproterozoic island arcs (Windley *et al.* 1996; Whitehouse *et al.* 1998*b*). Therefore, it can no longer be assumed that Proterozoic juvenile material in arc-rich greenstone belts continues to the Moho, in which case the very high crustal production rates quoted above are readily explicable, i.e. the actual rates were much less than those calculated.

Patchett & Arndt (1986) estimated that the overall crust production rates in central Laurentia and Baltica in the period of 1.9–1.7 Ga was *c.* 1.2 km^3/a, which is slightly greater than the total Phanerozoic island-arc accretion rate. Because this rate was derived only from the Yavapai–Penokean–Svecofennian Accretionary Orogen, they concluded that the global Early Proterozoic rate was around double the present rate. Condie (2001) estimated, from the aerial extent of crust on geological and tectonic maps, that crustal production rates during the formation of Rodinia in the period of 1.35–0.9 Ga fell within the 1.1 km^3/a average Phanerozoic production rate. But most parts of the Grenvillian Orogen lack sufficient geophysical data to make accurate estimates of volume. Also, the average rate for the Phanerozoic, quoted above, was calculated before the enormous extent of the largely Palaeozoic, juvenile, Altaid Orogen in Central Asia was known, and therefore must be suspect for purposes of comparison with Proterozoic rates.

The above estimates were of crustal production rates. However, the crustal net growth rate, which is equal to the mantle extraction rate minus the recycling rate, would provide a minimum estimate of crustal growth rates in the Proterozoic. The recycling rate includes all forms of recycling, including crustal delamination, subduction of sediments, oceanic crust, oceanic plateaus and island arcs, and erosion and redeposition in sedimentary basins. However, quantitative estimates of these Proterozoic processes are not known.

There seems little point in using the present-day thickness of Proterozoic continental crust anywhere in the world for these calculations, because many of the orogens concerned probably had a much greater crustal thickness at the time of their formation, and these thicknesses have been subsequently modified by erosion and/or extensional collapse. Abbott *et al.* (2000) calculated that the average thickness of Early Proterozoic crust was between 48±9 and 60±7 km (depending on the methods of calculation), but hardly anywhere do Early Proterozoic orogens today have this thickness. Only the Svecofennian Orogen has (up to) 60 km thick crust today and the original crust was even thicker, because the surface rocks are in the amphibolite or granulite facies. So, because most Proterozoic orogens today probably have a crustal thickness which is less than when originally formed, it is difficult to use present-day thicknesses in any calculations of production rate or net growth rate, and in any case the juvenile orogens at the present surface most likely do not continue to the Moho. Therefore, it seems that most estimates of the rate of growth of Proterozoic crust are premature and unreliable.

Discussion

The above brief summary of published magmatic rocks worldwide illustrates the current state of knowledge of the contribution of juvenile material to continental growth in the Proterozoic and provides a useful databank or background for discussion.

The magmatic rocks that contributed to crustal growth in the Proterozoic were generated in part by plate tectonic processes, such as oceanic plate accretion and subduction giving rise to ophiolites and island arcs, and in part by plume tectonic processes, giving rise to oceanic islands and plateaus, continental flood basalts and radiating dyke swarms. Supercontinents were important in this story because, on the one hand, they trapped fragments derived from the oceans and, on the other, because they acted as the framework for plume-generated breakup.

In the last few decades a variety of very different, theoretically based models have been produced to explain the secular development of continental crust with time. In recent years compilations of Nd isotopic data and U–Pb zircon ages suggest the following:

(1) At least 27–30% (Abbott *et al.* 2000, fig. 5) or 29–45% (Abbott *et al.* 2000, abstract) of the continental crust was extracted from the mantle by the start of the Proterozoic at 2.5 Ga.
(2) Between 50 and 52% (Abbott *et al.* 2000, fig. 5) or between 51 and 79% (Abbott *et al.* 2000, abstract) of the present volume of the continents existed by 1.8–2.0 Ga.
(3) Between 7 and 13% of the continental crust was formed between 1.35 and 0.9 Ga (Condie 2001)

In a detailed survey, Condie (1997*b*) concluded that of 96 post-Archaean greenstones, only *c.* 10% have oceanic plateau–MORB affinities, the bulk of the greenstones having arc signatures. From a study of the La/Nb ratio and Ni contents of mafic lower crustal xenoliths, Condie (1997*b*, 1999) discovered that about one third of the post-Archaean lower continental crust is composed of mafic rocks with mantle-plume signatures (oceanic plateaus) and the remainder are chiefly arc material. He concluded that the value of one third is a minimum for the plume component in the lower crust.

There are two fundamentally different models to explain the primary mechanisms that controlled the formation of continental crust. The first model is the **subduction model**, according to which new continents are formed at subduction zones with the result that >90% of the continental crust can be accounted for by convergent margin magmatism (Kay & Kay 1986; Rudnick *et al.* 1998). This model is based not only on field relations but also on trace element geochemical data using modern arc analogues. The results indicate that some Precambrian arcs formed through wet melting of upwelling asthenosphere at the initiation of oceanic subduction, others were generated from the mantle wedge during more mature stages of arc evolution, and yet others were generated through melting of young and/or warm slabs (Tarney & Jones 1994).

The second model is the **mantle plume model**, which states that new continents are formed by high degrees of partial melting within mantle plumes, giving rise to oceanic plateaus which are too thick and buoyant to subduct, and so accrete to and become part of a continent (Stein & Hofmann 1994; Abbott *et al.* 2000). The plume model is based on the calculation that oceanic plateaus thicker than 17 km are unsubductable (Cloos 1993). However, a variety of evidence suggests that parts or the whole of thick oceanic plateaus are able to subduct. For example, the Mesozoic Sanbagawa Belt of southern Japan contains a 2 km thick slab of an oceanic plateau originally 30 km thick, the Iratsu Body, which has been subducted to 90 km depth at 3 GPa and then raised by wedge extrusion as an eclogite into the accretionary orogen (Ota *et al.* 2002). Also, the lower ultramafic unit of the 1000 km long Sorachi Oceanic Plateau in northern Japan (Kimura *et al.* 1994), which attained a tectonized thickness of 30 km, is missing, presumed to have been subducted. Although a 4 km thick section of the 30 km thick Ontong Java Plateau is being obducted onto the Solomon Islands, three-dimensional tomographic inversion shows that a low-velocity root of the plateau has been subducted to 300 km depth (Klosko *et al.* 2001). Two-dimensional finite element modelling by van Hunen *et al.* (2002) shows that an 18 km thick oceanic plateau can subduct, causing development of a shallow-dipping or flat slab. Finally, Saunders *et al.* (1996) concluded that after 100 Ma a thick oceanic plateau may become potentially negatively buoyant and deeper zones will transform to eclogite, and so will be able to spontaneously subduct. With the calculation that oceanic plateaus <17 km in thickness are inherently subductable (Cloos 1993), and with the evidence of partial or complete subduction of thicker oceanic plateaus, it is not surprising that only five examples of obducted plateaus have been identified in the Mesozoic and Cenozoic geological record (Coffin & Eldholm 2001). Accordingly, the fact that modern oceanic plateaus can subduct, weakens the plume model for the growth of the continental crust.

The episodic growth of continents has been known for several decades from the episodic peaks of isotopic ages. In a recent re-evaluation, Condie (2000) concluded that the continents grew episodically, with major periods of growth (superevents) at 2.7 and 1.9 Ga, each superevent lasting only *c.* 800 Ma. The superevents are episodes of enhanced exchange between the lower and upper mantle, which replenish the upper mantle with juvenile trace elements, leading to rapid growth of the continents. Accordingly, Stein & Hofmann (1994) related these periods of major addition to mantle plumes that would produce oceanic plateaus. One result of a plume or superplume event would be increased seafloor spreading rates, as envisaged by Larson (1991), which would lead to increased subduction and a higher rate of island-arc production. Another result would be increased production of oceanic plateaus which, if not subducted, would accrete and increase the volume of continental crust (Abbott *et al.* 2000).

All the above data and models take on a new perspective when tomographic data are taken into account. They are interpreted to indicate that cold lithospheric slabs are subducted down to the D″core–mantle boundary, and some of that material is later incorporated into plumes or superplumes giving rise to oceanic plateaus, continental flood basalts and mafic dyke swarms. Condie (1998) related the superevents in the mantle at 2.7 and 1.9 Ga to catastrophic slab avalanching at the 660 km boundary.

Conclusions

Consideration of the data, ideas and models reviewed in this paper leads to the following conclusions.

(1) Both plate tectonics and plume tectonics played important roles in controlling growth of the continents during the Proterozoic. The former gave rise to oceanic plate accretion and subduction, expressed as island arcs, oceanic crust and ophiolites. The latter gave rise to oceanic islands and plateaus, continental flood basalts and mafic dyke swarms.

(2) The oceanic parts of the above scenario are mostly preserved in accretionary orogens, the modern equivalents of which are in the western Pacific (Japan to Indonesia).

(3) The two main competing models to explain the formation of the continental crust [subduction model (island arcs) v. plume model (oceanic plateaus)] have strong advocates. Problems associated with these models in relation to the Proterozoic include the following: (a) examination of the geological record shows that island arcs are the principal component of Proterozoic accretionary orogens; (b) the plume/oceanic plateau model relies heavily on the calculation that thick oceanic plateaus are too buoyant to subduct, and therefore must accrete and contribute substantially to crustal growth. However, only a few oceanic plateaus have been found in Proterozoic (and Archaean) orogens, and only five in the Mesozoic–Cenozoic geological record, and in the modern accretionary orogens of the western Pacific substantial parts of oceanic plateaus have been subducted. What is being seen today is only the remnants that have accreted to the trenches and the accretionary orogens.

(4) Calculation of crustal production rates and crustal net growth rates of Proterozoic crust are premature, because it is not known how much of the continental crust has been eroded and recycled or subducted, what parts of and how much of the components of the oceanic crust have been subducted, and so whether or not the preserved parts played an important or minor role in the accretion and growth of the continents. Finally there are still insufficient seismic surveys to confirm preliminary data which suggests that the crust in accretionary and collisional orogens has an imbricated structure, which implies that juvenile rocks at the present surface may not continue down to the Moho.

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References

ABBOTT, D., SPARKS, D., HERZBERG, C., MOONEY, W., NIKISHIN, A. & ZHANG, Y. S. 2000. Quantifying Precambrian crustal extraction: the root is the answer. *Tectonophysics*, **322**, 163–190.

ABOUCHAMI, W., BOHER, M., MICHARD, A. & ALBAREDE, F. 1990. A major 2.1 Ga event of mafic magmatism in West Africa: an early stage of crustal accretion. *Journal of Geophysical Research*, **95**, 17605–17629.

ÅHÄLL, K. I. & DALY, J. S. 1989. Age, tectonic setting and provenance of Östfold–Marstrand belt supracrustals: westward crustal growth of the Baltic Shield at 1760 Ma. *Precambrian Research*, **45**, 45–61.

AMELIN, Y. V. & SEMENOV, V. S. 1996. Nd and Sr isotopic geochemistry of mafic layered intrusions in the eastern Baltic shield: implications for the evolution of Paleoproterozoic continental mafic magmas. *Contributions to Mineralogy and Petrology*, **124**, 255–272.

AMELIN, Y. V., HEAMAN, L. M. & SEMENOV, V. S. 1995. U–Pb geochronology of layered mafic intrusions in the eastern Baltic Shield: implications for the timing and duration of Paleoproterozoic continental rifting. *Precambrian Research*, **75**, 31–46.

ANONYMOUS. 1972. Penrose Field Conference Report on Ophiolites. *Geotimes*, **17**, 24–25.

BENNETT, V. C. & DEPAOLO, D. J. 1987. Proterozoic crustal history of the western United States as determined by neodymium isotopic mapping. *Geological Society of America Bulletin*, **99**, 674–685.

BERHE, S. M. 1997. The Arabian–Nubian Shield. *In:* DE WIT, M. & ASHWAL, L. D. (eds) *Greenstone Belts.* Clarendon Press, Oxford, 761–771.

BEVINS, R. E., PHARAOH, T. C., COPE, J. C. W. & BREWER, T. S. 1995. Geochemical character of Neoproterozoic volcanic rocks in southwest Wales. *Geological Magazine*, **132**, 339–349.

BÉZIAT, D., BOURGES, F., DEBAT, P., LOMPO, M., MARTIN, F. & TOLLON, F. 2000. A Paleoproterozoic ultramafic–mafic assemblage and associated volcanic rocks of the Boromo greenstone belt: fractionates originating from island-arc volcanic activity in the West African craton. *Precambrian Research*, **101**, 25–47.

BINGEN, B., DEMAIFFE, D. & VAN BREEMEN, O. 1998. The 616Ma-old Egersund basaltic dike swarm, SW Norway, and Late Neoproterozoic opening of the Iapetus Ocean. *Journal of Geology*, **106**, 565–574.

BLASLUND, B., WHITE, S., BROOIJMANS, P., DE BROODER, H. & VISSER, W. 2000. Late Proterozoic extensional collapse in the Arabian–Nubian Shield. *Journal of the Geological Society, London*, **157**, 615–628.

BOHER, M., ABOUCHAMI, A., MICHARD, F., ALBAREDE, F. & ARNDT, N. T. 1992. Crustal growth in West Africa at 2.1 Ga, *Journal of Geophysical Research*, **97**, 347–369.

BREWER, T. S., DALY, J. S. & ÅHÄLL, K. I. 1998. Contrasting magmatic arcs in the Palaeoproterozoic of the south-western Baltic Shield. *Precambrian Research*, **92**, 297–315.

BUCHAN, C., CUNNINGHAN, D., WINDLEY, B. F. & TOMURHUU, D. 2001. Structural and lithological characteristics of the Bayankhongor ophiolite zone, Central Mongolia. *Journal of the Geological Society, London*, **158**, 445–460.

CAPDEVILA, R., ARNDT, N., LETENDRE, J. & SAUVAGE, J-F. 1999. Diamonds in volcaniclastic komatiite from French Guiana. *Nature*, **399**, 456–458.

CLOOS, M. 1993. Lithospheric buoyancy and collisional orogenesis: subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts. *Geological Society of America Bulletin*, **105**, 715–737.

CLOWES, R. M., CLAVERT, A. J., EATON, D. W., HAJNAL, Z., HALL, J. & ROSS, G. M. 1996. LITHOPROBE reflection studies of Archean and Proterozoic crust in Canada. *Tectonophysics*, **264**, 65–88.

COFFIN, M. E. & ELDHOLM, O. 2001. Large igneous provinces: progenitors of some ophiolites? *Geological Society of America, Special Paper*, **353**, 59–70.

CONDIE, K. C. 1997*a*. Sources of Proterozoic mafic dyke swarms: constraints from Th/Ta and La/Yb ratios. *Precambrian Research*, **81**, 3–14.

CONDIE, K. C. 1997*b*. Contrasting sources for upper and lower continental crust: the greenstone connection. *Journal of Geology*, **105**, 729–736.

CONDIE, K. C. 1998. Episodic continental growth and supercontinents: a mantle avalanche connection? *Earth and Planetary Science Letters*, **163**, 97–108.

CONDIE, K. C. 1999. Mafic crustal xenoliths and the origin of the lower continental crust. *Lithos*, **46**, 95–101.

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

CONDIE, K. C. 2001. Continental growth during formation of Rodinia at 1.35–0.9Ga. *Gondwana Research*, **4**, 5–16.

DANN, J. C. & BOWRING, S. A. 1997. The Payson ophiolite and Yavapai–Mazatzal orogenic belt, central Arizona. *In:* DE WIT, M. & ASHWAL, L. D. (eds) *Greenstone Belts.* Clarendon Press, Oxford, 781–790.

DARRACOTT, B. W. 1975. The interpretation of the gravity anomaly over the Barberton Mountain Land, South Africa. *Transactions of the Geological Society of South Africa*, **78**, 123–128.

DUPUY, C., MICHARD, A., DOSTAL, J., DAUTEL, D. & BARAGAR, W. R. A. 1992. Proterozoic flood basalts from the Coppermine River area, Northwest Territories: isotope and trace element geochemistry. *Canadian Journal of Earth Sciences*, **29**, 1937–1943.

ELMING, S.-Å. & MATTSSON, H. 2001. Post-Jotnian basic intrusions in the Fennoscandian shield and the break-up of Baltica from Laurentia: a palaeomagnetic and AMS study. *Precambrian Research*, **108**, 215–236.

ERNST, R. E. & BUCHAN, K. L. 2001*a*. Large mafic magmatic events through time and links to mantle plume heads. *In:* ERNST, R. E. & BUCHAN, K. L. (eds) *Mantle Plumes: their identification through time*. Geological Society of America, Special Paper **352**, 483–575.

ERNST, R. E. & BUCHAN, K. L. 2001*b*. The use of mafic dike swarms in identifying and locating mantle plumes. *In*: ERNST, R. E. & BUCHAN, K. L. (eds) *Mantle Plumes: their Identification through Time*. Geological Society of America, Special Paper, **352**, 247–265.

FERNÁNDEZ-SUÁREZ, J., GUTIÉRREZ-ALONSO, G., JENNER, G. A. & TUBRETT, M. N. 2000. New ideas on the Proterozoic–Early Palaeozoic evolution of NW Iberia: insights from U–Pb detrital zircon ages. *Precambrian Research*, **102**, 185–206.

GLEBOVITSKY, V., MARKER, M., ALEXEJEV, N., BRIDGWATER, D., SEDOVA, I., SALNIKOVA, E. & BEREZHNAYA, N. 2001. Age, evolution and regional setting of the Palaeoproterozoic Umba igneous suite in the Kolvitsa–Umba zone, Kola Peninsula: constraints from new geological, geochemical and U–Pb zircon data. *Precambrian Research*, **105**, 247–267.

GRISELIN, M., ARNDT, N. & BARAGAR, W. R. A. 1997. Plume–lithosphere interaction and crustal contamination during formation of Coppermine River basalts, Northwest Territories, Canada. *Canadian Journal of Earth Sciences*, **34**, 958–975.

HEAMAN, L. M. 1997. Global mafic magmatism at 2.45 Ga: remnants of an ancient large igneous province? *Geology*, **25**, 299–302.

HOFFMAN, P. F. 1988. United plates of America, the birth of a craton: early Proterozoic assembly and growth of Laurentia. *Annual Reviews of Earth and Planetary Sciences*, **16**, 543–603.

ILG, B. R., KARLSTROM, K. E., HAWKINS, D. P. & WILLIAMS, M. L. 1996. Tectonic evolution of Paleoproterozoic rocks in the Grand Canyon: insights into middle-crustal processes. *Geological Society of America Bulletin*, **108**, 1149–1166.

JOHNSON, S. P. & OLIVER, G. J. H. 2000. Mesoproterozoic oceanic subduction, island-arc formation and the initiation of back-arc spreading in the Kibaran Belt of central, southern Africa: evidence from the Ophiolite Terrane, Chewore Inliers, northern Zimbabwe. *Precambrian Research*, **103**, 125–146.

KAY, R. W. & KAY, S. M. 1986. Petrology and geochemistry of the lower continental crust: an overview. *In*: DAWSON, J. B., CARSWELL, D. A., HALL, J. & WEDEPOHL, K. H. (eds) *The Nature of the Lower Continental Crust*: Geological Society of London, Special Publication, **27**, 147–159.

KEMPTON, P. D., DOWNES, H., NEYMARK, L. A., WARTHO, J. A., ZARTMAN, R. E. & SHARKOV, E. V. 2001. Garnet granulite xenoliths from the northern Baltic Shield – the underplated lower crust of a Palaeoproterozoic Large Igneous Province. *Journal of Petrology*, **42**, 731–763.

KHAIN, V. E., GUSEV, G. S., KHAIN, E. V., VERNIKOVSKY, V. A. & VOLOBUYEV, M. I. 1997. Circum-Siberian Neoproterozoic ophiolite belt. *Ofioliti*, **22**, 195–200.

KIMURA, G., SAKAKIBARA, M. & OKAMURA, M. 1994. Plumes in central Panthalassa? Deductions from accreted oceanic fragments in Japan. *Tectonics*, **13**, 905–916.

KLOSKO, E. R., RUSSO, R. M., OKAL, E. A. & RICHARDSON, W. P. 2001. Evidence for a rheologically strong chemical mantle root beneath the Ontong Java plateau. *Earth and Planetary Science Letters*, **186**, 347–361.

KONTINEN, A. 1987. An Early Proterozoic ophiolite – the Jormua mafic–ultramafic complex, northeastern Finland. *Precambrian Research*, **35**, 313–341.

KORJA, A. & HEIKKINEN, P. J. 1995. Proterozoic extensional tectonics of the central Fennoscandian shield: results from the Baltic and Bothnian Echoes from the Lithosphere experiment. *Tectonics*, **14**, 504–517.

LARSON, R. L. 1991. Geological consequences of superplumes. *Geology*, **19**, 963–966.

LEGAULT, F., FRANCIS, D., HYNES, A. & BUDKEWITSCH, P. 1994. Proterozoic continental volcanism in the Belcher Islands: implications for the evolution of the Circum Ungava fold belt. *Canadian Journal of Earth Sciences*, **31**, 1536–1549.

LEWRY, J. F. & STAUFFER, M. R. 1997. The Reindeer zone of the Trans-Hudson orogen, Canada. *In:* DE WIT, M. & ASHWAL, L. D. (eds) *Greenstone Belts*. Clarendon Press, Oxford, 739–745.

LEYBOURNE, M. I., VAN WAGONER, N. A. & AYRES, L. D. 1997. Chemical stratigraphy and petrogenesis of the Early Proterozoic Amisk Lake volcanic sequence, Flin Flon–Snow Lake greenstone belt, Canada. *Journal of Petrology*, **38**, 1541–1564.

LUCAS, S. B., ST-ONGE, M. R., PARRISH, R. R. & DUNPHY, J. M. 1992. Long-lived continent–ocean interaction in the Early Proterozoic Ungava orogen, northern Quebec, Canada. *Geology*, **20**, 113–116.

LUCAS, S. B., STERN, R. A., SYME, E. C., REILLY, B. A. & THOMAS, D. J. 1996. Intraoceanic tectonics and the development of continental crust: 1.92–1.84 Ga evolution of the Flin Flon belt, Canada. *Geological Society of America Bulletin*, **108**, 602–629.

LUOSTO, U. 1997. Structure of the Earth's crust in Fennoscandia as revealed from refraction and wide-angle reflection studies. *Geophysica*, **33**, 3–16.

MELEZHIK, V. A. & STURT, B. A. 1994. General geology and evolutionary history of the early Proterozoic Polmak–Pasvik–Pechenga–Imandra/Varzuga–Ust'Po noy greenstone belt in the northeastern Baltic Shield. *Earth Science Reviews*, **36**, 205–241.

MYERS, J. S., SAHW, R. D. & TYLER, I. M. 1996. Tectonic evolution of Proterozoic Australia. *Tectonics*, **15**, 1431–1446.

OLIVER, G. J. H., JOHNSON, S. P., WILLIAMS, I. S. & HERD, D. A. 1998. Relict 1.4 Ga oceanic crust in the Zambezi Valley, northern Zimbabwe: evidence for Mesoproterozoic supercontinental fragmentation. *Geology*, **26**, 571–573.

ORTEGA-GUTIERREZ, F., RUIZ, J. & CENTENO-GARCIA, E. 1995. Oaxaguia, a Proterozoic microcontinent accreted to North America during the late Paleozoic. *Geology*, **23**, 1127–1130.

OTA, T., TERABAYASHI, M. & KATAYAMA, I. 2002. Thermobaric structure and metamorphic evolution of the Iratsu eclogite body in the Sanbagawa belt, central Shikoku, Japan. *Lithos*, in press.

PARK, J. K., BUCHAN, K. L. & HARLAN, S. S. 1995. A proposed giant radiating dyke swarm fragmented by the separation of Laurentia and Australia based on paleomagnetism of ca. 780 Ma mafic intrusions in western North Australia. *Earth and Planetary Science Letters*, **132**, 129–139.

PARK, R. G. 1994. Early Proterozoic tectonic overview of the northern British Isles and neighbouring terrains in Laurentia and Baltica. *Precambrian Research*, **68**, 65–79.

PARK, R. G., TARNEY, J. & CONNELLY, J. N. 2001. The Loch Maree Group: Palaeoproterozoic subduction–accretion complex in the Lewisian of NW Scotland. *Precambrian Research*, **105**, 205–226.

PATCHETT, P. J. & ARNDT, N. T. 1986. Nd isotopes and tectonics of 1.9–1.7 Ga crustal genesis. *Earth and Planetary Science Letters*, **78**, 329–338.

PATCHETT, P. J. & RUIZ, J. 1989. Nd isotopes and the origin of Grenville-age rocks in Texas: implication for Proterozoic evolution of the United States Mid-continent region. *Journal of Geology*, **97**, 685–695.

PELTONEN, P., KONTINEN, A. & HUHMA, H. 1996. Petrology and geochemistry of metabasalts from the 1.95 Ga Jormua ophiolite, northeastern Finland. *Journal of Petrology*, **37**, 1359–1383.

PERCIVAL, J. A. & WEST, G. F. 1994. The Kapuskasing Uplift: a geological and geophysical synthesis. *Canadian Journal of Earth Sciences*, **31**, 1256–1286.

PUCHTEL, I. S., ARNDT, N.T., HOFMANN, A. W *ET AL.* 1998. Petrology of mafic lavas within the Onega plateau, central Karelia: evidence for 2.0 Ga plume-related continental crustal growth in the Baltic Shield. *Contributions to Mineralogy and Petrology*, **130**, 134–153.

PUCHTEL, I. S., HAASE, K. M., HOFMANN, A. W., CHAUVEL, C., KULIKOV, V. S., GARBE-SCHÖNBERG, C.-D. & NEMCHIN, A. A. 1997. Petrology and geochemistry of crustally contaminated komatiitic basalts from the

Vetreny belt, southeastern Baltic Shield: evidence for an early Proterozoic mantle plume beneath rifted Archean continental lithosphere. *Geochimica et Cosmochimica Acta*, **61**, 1205–1222.

REYMER, A. & SCHUBERT, G. 1986. Rapid growth of some major segments of continental crust. *Geology*, **14**, 299–302.

RUDNICK, R. L., MCDONAUGH, W. P. & O'CONNELL, R. J. 1998. Thermal structure, thickness, and composition of continental lithosphere. *Chemical Geology*, **145**, 395–411.

SAUNDERS, A. D., TARNEY, J., KERR, A. C. & KENT, R. W. 1996. The formation and fate of large igneous provinces. *Lithos*, **37**, 81–95.

SCOTT, D. J., HELMSTAEDT, H. & BICKLE, M. J. 1992. Purtuniq ophiolite, Cape Smith belt, northern Quebec, Canada: a reconstructed section of Early Proterozoic oceanic crust. *Geology*, **20**, 173–176.

SENGÖR, A. M. C., NATAL'IN, B. A. & BURTMAN, V. S. 1993. Evolution of the Altaid tectonic collage and Palaeozoic crustal growth in Eurasia. *Nature*, **364**, 299–307.

SEYMOUR, K. ST. & KUMARAPELI, P. S. 1995. Geochemistry of the Grenville dyke swarm: role of plume-source mantle in magma genesis. *Contributions to Mineralogy and Petrology*, **120**, 29–41.

SNYDER, D. B., LUCAS, S. B. & MCBRIDE, J. H. 1996. Crustal and mantle reflectors from Paleoproterozoic orogens and their relation to arc–continent collisions. *In*: BREWER, T. S. (ed.) *Precambrian Crustal Evolution of the North Atlantic Regions*. Geological Society of London, Special Publication **112**, 1–23.

STEIN, M. & GOLDSTEIN, S. L. 1996. From plume head to continental lithosphere in the Arabian–Nubian shield. *Nature*, **382**, 773–778.

STEIN, M. & HOFMANN, A.W. 1994. Mantle plumes and episodic crustal growth. *Nature*, **372**, 63–68.

STERN, R. A., SYME, E. C. & LUCAS, S. B. 1995. Geochemistry of 1.9 Ga MORB- and OIB-like basalts from the Amisk collage, Flin Flon belt, Canada: evidence for an intra-oceanic origin. *Geochimica et Cosmochimica Acta*, **59**, 3131–3154.

ST-ONGE, M. R., LUCAS, S. B. & SCOTT, D. J. 1997. The Ungava orogen and the Cape Smith thrust belt. *In:* DE WIT, M. & ASHWAL, L. D. (eds) *Greenstone Belts.* Clarendon Press, Oxford, 772–780.

TARNEY, J. & JONES, C. E. 1994. Trace element geochemistry of orogenic igneous rocks and crustal growth models. *Journal of Geological Society, London*, **151**, 855–868.

VAN HUNEN, J., VAN DER BERG, A. & VLAAR, N. J. 2002. On the role of subducting oceanic plateaus in the development of shallow flat subduction. *Proceedings of an International Workshop: Role of Superplumes in the Earth System.* Tokyo Institute of Technology, Tokyo, Japan, Abstract volume, 120–123.

VANDERHAEGHE, O., LEDRU, P., THIÉBLEMONT, D., EGAL, E., COCHERIE, A., TEGYEY, M. & MILÈSI, J-P. 1998. Contrasting mechanism of crustal growth; geodynamic evolution of the Paleoproterozoic granite–greenstone belts of French Guiana. *Precambrian Research*, **92**, 165–193.

VOGEL, D. C., VUOLLO, J. I., ALAPIETI, T. T. & JAMES, R. S. 1998. Tectonic, stratigraphic and geochemical comparisons between ca. 2500–2440 Ma mafic igneous events in the Canadian and Fennoscandian Shields. *Precambrian Research*, **92**, 89–116.

WALKER, R. J., MOGAN, J. W., HANSKI, E .J. & SMOLKIN, V. F. 1997. Re–Os systematics of Early Proterozoic ferropicrites, Pechenga Complex, northwestern Russia: evidence for ancient ^{187}Os-enriched plumes. *Geochimica et Cosmochimica Acta*, **61**, 3145–3160.

WHITEHOUSE, M. J. & BRIDGWATER, D. 2001. Geochronological constraints on Paleoproterozoic crustal evolution and regional correlations of the northern Outer Hebridean Lewisian complex, Scotland. *Precambrian Research*, **105**, 227–245.

WHITEHOUSE, M. J., KALSBEEK, F. & NUTMAN, A. P. 1998*a*. Crustal growth and crustal recycling in the Nagssugtoqidian orogen of West Greenland: constraints from radiogenic isotope systematics and U–Pb zircon geochronology. *Precambrian Research*, **91**, 365–381.

WHITEHOUSE, M. J., WINDLEY, B. F., BA-BTTAT, F. A. O., FANNING, C. M. & REX, D. C. 1998*b*. Crustal evolution and terrane correlation in the eastern Arabian Shield, Yemen: geochronological constraints. *Journal of the Geological Society, London*, **155**, 281–295.

WINDLEY, B. F. 1992. Proterozoic collisional and accretionary orogens. *In*: CONDIE, K. C. (ed) *Proterozoic Crustal Evolution*. Elsevier, Amsterdam, 419–446.

WINDLEY, B. F. 1995. *The Evolving Continents* (3rd edition). Wiley, Chichester.

WINDLEY, B. F. WHITEHOUSE, M. J. & BA-BTTAT, M. A. O. 1996. Early Precambrian gneiss terranes and Pan-African island arcs in Yemen: crustal accretion of the eastern Arabian shield. *Geology*, **24**, 131–134.

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**, 1235–1259.

YAKUBCHUK, A. S., NIKISHIN, A. M. & ISHIWATARI, A. 1994. A Late Proterozoic ophiolite pulse. *Proceedings of the 29th International Geological Congress*, part D, 273–286.

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.