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Earth's Continental Lithosphere Through Time

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Keywords

continental crust, continental mantle lithosphere, plate tectonics, preservation bias, early Earth

Abstract

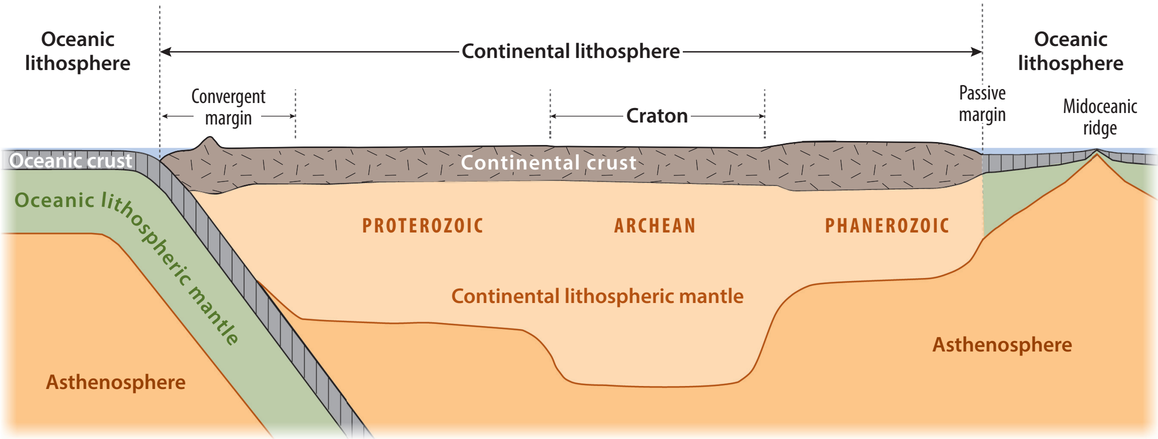
The record of the continental lithosphere is patchy and incomplete; no known rock is older than 4.02 Ga, and less than 5% of the rocks preserved are older than 3 Ga. In addition, there is no recognizable mantle lithosphere from before 3 Ga. We infer that there was lithosphere before 3 Ga and that ~3 Ga marks the stabilization of blocks of continental lithosphere that have since survived. This was linked to plate tectonics emerging as the dominant tectonic regime in response to thermal cooling, the development of a more rigid lithosphere, and the recycling of water, which may in turn have facilitated plate tectonics. A number of models, using different approaches, suggest that at 3 Ga the volume of continental crust was ~70% of its present-day volume and that this may be a minimum value. The continental crust before 3 Ga was on average more mafic than that generated subsequently, and this pre-3 Ga mafic new crust had fractionated Lu/Hf and Sm/Nd ratios as inferred for the sources of tonalite-trondhjemite-granodiorite and later granites. The more intermediate composition of new crust generated since 3 Ga is indicated by its higher Rb/Sr ratios. This change in composition was associated with an increase in crustal thickness, which resulted in more emergent crust available for weathering and erosion. This in turn led to an increase in the Sr isotope ratios of seawater and in the drawdown of CO_2. Since 3 Ga, the preserved record of the continental crust is marked by global cycles of peaks and troughs of U-Pb crystallization ages, with the peaks of ages appearing to match periods of supercontinent assembly. There is increasing evidence that the peaks of ages represent enhanced preservation of magmatic rocks in periods leading up to and including continental collision in the assembly of supercontinents. These are times of increased crustal growth because more of the crust that is generated is retained within the crust. The rates of generation of continental crust and mantle lithosphere may have remained relatively constant at least since 3 Ga, yet the rates of destruction of continental crust have changed with time. Only relatively small volumes of rock are preserved from before 3 Ga, and so it remains difficult to establish which of these are representative of global processes and the extent to which the rock record before 3 Ga is distorted by particular biases.

1. INTRODUCTION

The lithosphere is the rigid outer layer of Earth, and it is divisible into crust and upper mantle components (**Figure 1**). The formation and evolution of the lithosphere are driven by thermal energy from the planet's interior and modulated by solar radiation, which regulates the temperature at the surface of the lithosphere. The lithosphere records interactions with the underlying asthenospheric mantle and with the overlying hydrosphere and atmosphere. The lithosphere is the long-term archive of Earth's history, and yet our knowledge of this history is limited by variable but generally poor preservation and by difficulties in accessing its deep crustal and mantle components. The rock record is incomplete, and the lithosphere that is preserved may not be representative of the processes that formed it (Hawkesworth et al. 2016). The rocks that are readily accessible for sampling record long and complex cycles of generation and reworking, with most crustal rocks derived from preexisting crustal rocks, from "the ruins of mountains" (Hutton 1788, p. 215). Hence, it remains difficult to see back through to the composition and to the tectonic settings in which the continental lithosphere was generated. In this contribution, we consider the nature of the continental crust and its mantle lithosphere, focusing on the former, for which there are more data, and then discuss the interrelationship between the two and the tectonic processes that have controlled their evolution.

At the present, the distinction between the continental and oceanic lithosphere is marked. The oceanic lithosphere is relatively dense, and its crustal component is mafic and less than 200 Myr old (Cloos 1993, Kearey et al. 2009). In contrast, the continental lithosphere is more differentiated and buoyant and is harder to destroy; the crust preserves rocks that are up to 4.02 Ga and zircons that are up to 4.4 Ga in age (Bowring & Williams 1999, Reimink et al. 2016, Wilde et al. 2001). Half of the exposed rocks in the continental crust have formation ages (e.g., igneous crystallization ages for magmatic rocks, or deposition ages for sedimentary rocks, referred to here as geological ages) less than 600 Ma, and fewer than 5% have geological ages older than 3.0 Ga (Goodwin 1996). Before 3.0 Ga, more of Earth's crust appears to have been mafic, and the distinction between continental and oceanic crust was less clear cut. Lee et al. (2016) use the term mafic continental crust for continental crust older than 2.7 Ga. The continental crust evolved from a predominantly mafic composition to today, when it is andesitic and consists of three generalized layers: the upper, middle, and lower crust (Dhuime et al. 2015, Rudnick & Gao 2003, Tang et al. 2016). See Hacker et al. (2015) for an alternative two-layer crust.

Figure 2 summarizes the ages (circles) and age ranges (bars) of major events and cycles in Earth history. It highlights the nature of the geological record through the peaks of zircon crystallization



Crystallization age
Age of crystallization of a mineral or rock from a melt

Model age
Age at which new crust is generated from the mantle

Depleted mantle
Mantle depleted through extraction of one or more basaltic melts

Crust generation
Formation of new crust; extracted from the mantle

Crust recycling (and destruction)
Return of crust to the mantle

Crust reworking
Intracrustal remobilization, involving erosion and sedimentation, and/or (re)melting of preexisting crustal rocks

Growth of crust
The volume of new crust less the amount lost by recycling

Supercontinents
Assembly of large volumes of continental crust, i.e., redistribution of continental crust on Earth's surface

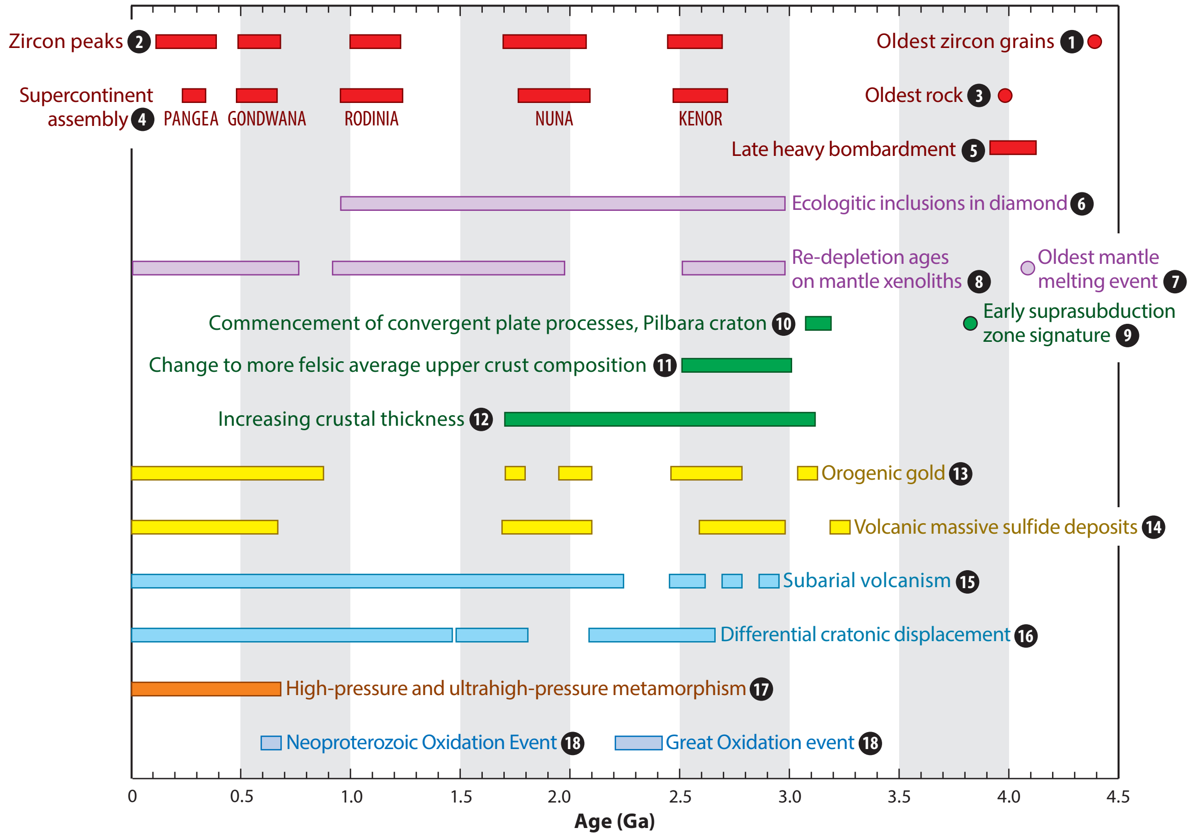
Figure 1

Schematic cross section of types of continental lithosphere emphasizing the thick stable nature of Archean cratons, accompanied by a glossary of some of the terms used. Thickness of lithosphere beneath Archean regions is of the order of 200–250 km, and that of the oceanic lithosphere is up to 100 km (after Cawood et al. 2013).

ages and the distribution of mantle depletion ages, the cyclical pattern of supercontinent formation, and the apparently intermittent nature of Au and volcanic massive sulfide mineralization. The early crust is not recorded, but it is assumed to have developed from the initial magma ocean and to have been mafic in composition (e.g., Elkins-Tanton 2008). The oldest zircon at 4.4 Ga indicates that there were felsic magmas at that time, and the elevated $\delta^{18}O$ values in some zircon crystals are taken to be evidence for the presence of surface water by 4.4 Ga (Valley et al. 2002, Wilde et al. 2001). Mafic and ultramafic volcanic rocks in greenstone belts are associated with tonalite-trondhjemite-granodiorite rocks in the classic granite-greenstone belt terrains of the Archean. The late Archean was characterized by tonalite-trondhjemite-granodiorite magmatism, the remelting of intermediate to felsic crust, the widespread generation of relatively potassic granites, and the stabilization of the continental crust and mantle lithosphere (Anhaeusser & Robb 1981, Carlson et al. 2005, Dewey & Windley 1981, Schoene et al. 2009).

The lithosphere is described on the basis of its mechanical properties as Earth's rigid outer layer, which can support horizontal and vertical stresses. It is recognized as a thermal boundary layer that cools by conduction, as a seismic lithosphere marked by an abrupt decrease in shear wave velocity at its base (a low-velocity zone), and as an elastic layer whose thickness is determined from the response to surface loads, such as seamounts and mountain belts (Watts et al. 2013).

Geotherms are steeper in the continents, and so the thermal boundary layer is thicker beneath continents than beneath the oceans. The continental mantle lithosphere shows significant variations as surface heat flux values increase, and thickness decreases, with decreasing lithospheric age. Present-day heat flux values range from 30–40 mW/m^2 for Archean regions with lithosphere thicknesses of >150 km to >60–80 mW/m^2 and generally <100 km for Phanerozoic regions (Jaupart & Mareschal 2003, Jaupart et al. 1998, McLennan et al. 2005, Nyblade 1999, Pollack et al. 1993). Moreover, global versus seismic tomography models (Grand 2002, Shapiro & Ritzwoller 2002) and the global thermal model of Artemieva & Mooney (2001) yield consistent estimates of the lithospheric thickness beneath continents (Artemieva 2009).



1 Wilde et al. 2001
2 Voice et al. 2011 and references therein
3 Bowring & Williams 1999
4 Adapted from Campbell & Allen 2008
5 Marchi et al. 2014
6 Shirey & Richardson 2011
7 Malitch & Merkle 2004
8 Figure 5 and references
9 Jenner et al. 2009, O'Neil et al. 2011, and Turner et al. 2014
10 Smithies et al. 2007 and Van Kranendonk et al. 2007
11 Keller & Schoene 2012 and Tang et al. 2016
12 Dhuime et al. 2015
13 Goldfarb et al. 2001
14 Mosier et al. 2009
15 Kump & Barley 2007, Shields 2007, Cawood et al. 2013, and Flament et al. 2013
16 Cawood et al. 2006, Evans & Pisarevsky 2008
17 Brown 2006, 2007, 2014
18 Campbell & Squire 2010, Farquhar et al. 2013 and references therein

The continental lithospheric mantle beneath Archean-aged crust tends to be composed of dehydrated, highly depleted mantle peridotite (Boyd 1989, Boyd et al. 1997, Griffin et al. 2009, Pearson et al. 1995, Pollack 1986), resulting in it being intrinsically buoyant and strong, which counteracts the destabilizing effect of its cold thermal state (Jordan 1978, 1988). These regions of thick (> ~150 km) stable lithosphere are referred to as cratons (kratogen) (**Figure 1**) (Kober 1921), and owing to their association with Archean crust, as well as Archean ages from xenoliths of underlying mantle lithosphere (Pearson & Wittig 2013), they are largely thought to have formed at this time. From a geological perspective, cratons may be viewed as areas of stable continental lithosphere in which weakly deformed and little metamorphosed Proterozoic sedimentary rocks unconformably overlie a deformed and metamorphosed Archean basement (e.g., Moores & Twiss 1995). On the basis of recent interpretations of Rayleigh wave tomography data, McKenzie et al. (2015) show that thick lithosphere (>200 km) remains a feature of Archean terrains and that areas of 150–200-km thick lithosphere are present under the Proterozoic crust. McKenzie & Priestley (2008) refer to the thick lithosphere currently developed beneath Tibet as a proto-craton, implying the process of craton formation is in part related to continental collision and continues today.

2. OROGENIC AND GLOBAL CYCLES

Geological cycles operate on a range of spatial and temporal scales. James Hutton (1785, 1788) concluded that there had been innumerable cycles, involving deposition on the seabed, uplift with tilting and erosion, and submergence under water for additional layers to be deposited. Orogenic cycles provide a framework whereby periods of sediment deposition are followed by compression, crustal thickening, uplift, and erosion. They differ in spatial and temporal scales from supercontinent cycles, which are global features and are therefore more likely to have had a significant role in the evolution of the continental crust as a whole. Supercontinent cycles have punctuated the post-Archean history of Earth, and they involve the episodic assembly and breakup of large landmasses, with profound consequences for the geologic record of both the solid and surficial Earth (Nance et al. 2014, Worsley et al. 1986).

Zircon remains the key timepiece for dating events in the geological record, and hundreds of thousands of U-Pb zircon ages have been published (e.g., Voice et al. 2011). Zircon typically crystallizes from relatively felsic magmas, and it is preserved in detrital sediments that have sampled continental crust. Compilations of age data have long been known to yield episodic

Figure 2

Ages (*circles*) and age ranges (*bars*) of major events and cycles preserved in Earth history. Sources of the data are indicated in the figure key: (1) The oldest terrestrial fragments, zircons from the Jack Hills; (2) the age range of principal peaks in U-Pb crystallization ages; (3) the oldest rocks, Slave craton; (4) the approximate age range of supercontinent assembly; (5) the late heavy bombardment; (6) the age range of eclogitic inclusions in diamonds; (7) the oldest Os model ages from detrital osmiridium grains, Witwatersrand Basin; (8) Re-depletion age ranges for peridotite xenoliths (see also **Figure 5**); (9) the ~3.8 Ga Nuvvuagittuq greenstone belt, Superior craton, and 3.85 Ga metabasalts from Isua, Greenland, are similar to modern subduction-related islands; (10) the pre-3.2 Ga rock units from Pilbara craton are associated with vertical tectonics, and rocks younger than 3.1 Ga are inferred to have formed by plate subduction processes; (11) the shift from predominantly mafic crustal compositions prior to 3.0 Ga to increasingly felsic compositions by 2.5 Ga; (12) the secular increase in time-integrated Rb/Sr ratios between 3.1 Ga and 1.7 Ga, indicative of a doubling in average continental crustal thickness from ca. 20 km to ca. 40 km; (13) the main pulses of orogenic gold deposition; (14) the main pulses of volcanic-hosted massive sulfide deposits; (15) subaerial large igneous province magmatism increases from 3.0 to 2.5 Ga, corresponding with the increase in $^{87}Sr/^{86}Sr$ ratios of seawater consistent with the emergence and weathering of continental crust; (16) the significant difference between the relative paleopositions of the Kaapvaal and Superior cratons at 2.68 Ga and 2.07 Ga and Baltica and Australia between 1.77 Ga and 1.5 Ga; (17) the high-pressure and ultrahigh-pressure metamorphism, which are limited to rock units dated at <0.7 Ga; and (18) the increases in atmospheric oxygen values during the Great Oxidation Event and the Neoproterozoic Oxidation Event.

distributions (Gastil 1960, Stockwell 1961), but these were based on mineral systems that can be reset through younger tectonothermal events (e.g., K-Ar, Ar-Ar, Rb-Sr). However, this episodic distribution is also a characteristic of global compilations of physiochemical resistant zircon U-Pb ages (**Figure 3**). Such apparently cyclical patterns are unexpected in a plate tectonic Earth in which broadly similar volumes of magmas are likely to have been generated, and destroyed, each year (Scholl & von Huene 2009).

The peaks and troughs of zircon ages have been interpreted as primary signals of periods of high and low volumes of granitic magmas, perhaps related to mantle superplumes (Albarède 1998, Arndt & Davaille 2013, Condie 1998, Condie et al. 2015, Parman 2015, Rino et al. 2004). There are, however, a number of observations that question this interpretation. First, the distribution of ages is based on a global data set, and so it would appear necessary to invoke periods of increased plume activity on a large scale. Second, there are peaks of zircon crystallization ages at ~1 Ga, 0.6 Ga, and 0.3 Ga (**Figure 3**), corresponding with the Grenville, Pan-African, and Variscan-Alleghenian orogenies, which are widely regarded as periods of continental assembly and collision (e.g., Cawood & Buchan 2007, Cawood et al. 2016, Gower & Krogh 2002), rather than times of unusual volumes of magma generation. Third, the composition of the continental crust is dominated by minor and trace element features that are characteristic of subduction-related magmas rather than those generated in intraplate settings (Rudnick & Gao 2003).

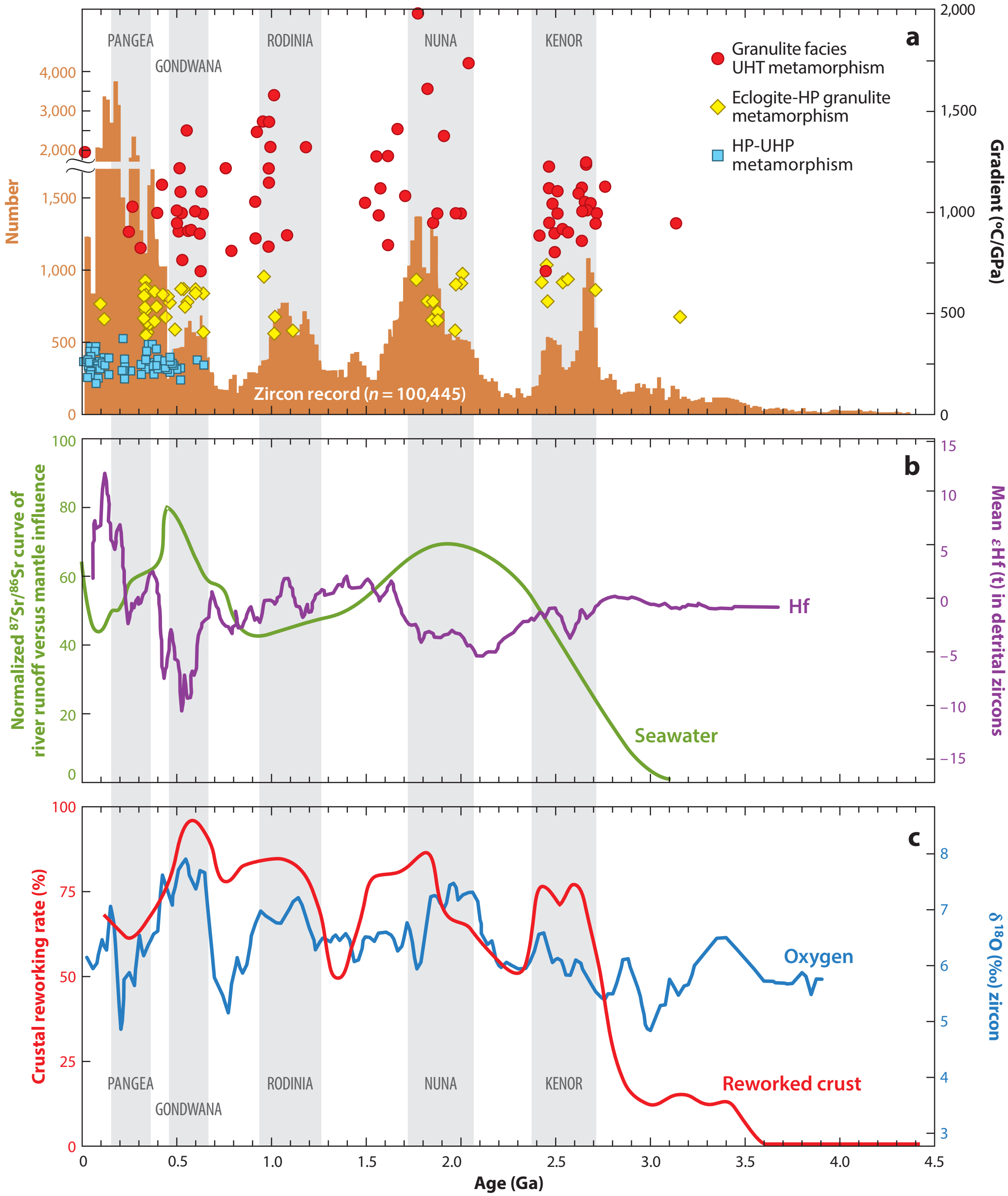
Large volumes of magma are generated above subduction zones, but in these settings, continental crust is also destroyed by erosion, subduction, and in some areas density foundering at rates similar to, or greater than, those at which new crust is generated (Clift et al. 2009; Scholl & von Huene 2007, 2009; Stern 2011). The existence and ages of different supercontinents are still matters for debate (e.g., Arndt & Davaille 2013, Bradley 2011, Campbell & Allen 2008), but there is a broad link between the peaks of U-Pb zircon ages, periods of crustal reworking as determined from zircon Hf and O isotopic records, and the inferred development of supercontinents (**Figure 3**). We therefore prefer models in which the peaks of zircon ages reflect the preferential preservation of magmatic rocks formed during the latter stages of ocean closure that mark the times of supercontinent assembly (Cawood et al. 2013, Hawkesworth et al. 2009). These would then be times in which the volume of stabilized continental crust would be expected to increase not because more new crust had been generated but because more of the crust that had been generated was retained within the continental crust. Overall the supercontinent cycle inherently biases the rock record both through selective isolation of material in continental cores during supercontinent assembly and through removal and recycling of material formed during stages of extension and convergence.

The assembly of supercontinents involves continental collision, and the associated thickening of the continental crust should result in increased amounts of crustal reworking. One proxy for

Figure 3

(*a*) Histogram of over 100,000 detrital zircon analyses showing peaks in U-Pb crystallization ages over the course of Earth history (Voice et al. 2011), which are similar to the ages of supercontinent assembly. Also shown is the apparent thermal gradient at the age of peak metamorphism for the three main types of extreme metamorphic belts (Brown 2007, 2014). (*b*) Normalized seawater $^{87}Sr/^{86}Sr$ curve (Shields 2007), and the running mean of initial εHf in ~7,000 detrital zircons from recent sediments (Cawood et al. 2013). (*c*) The red curve represents changes in the degrees of reworking of the continental crust through time, calculated from the distributions of the proportions of reworked crust and new crust, calculated from Hf isotope variations in zircons (Dhuime et al. 2012). The blue curve is the moving average distilled from compilation of ~3,300 $\delta^{18}O$ analyses of zircon versus U-Pb ages (Spencer et al. 2014). Abbreviations: HP, high-pressure; UHP, ultrahigh-pressure; UHT, ultrahigh-temperature. εHf is a measure of the deviation of the Hf isotope ratios of a sample from that of a chondritic reservoir, taken to be the bulk earth, as expressed in

$$\varepsilon\mathrm{Hf} = \left[\frac{(^{176}\mathrm{Hf}/^{177}\mathrm{Hf})_{\mathrm{sample}} - (^{176}\mathrm{Hf}/^{177}\mathrm{Hf})_{\mathrm{Chon}}}{(^{176}\mathrm{Hf}/^{177}\mathrm{Hf})_{\mathrm{Chon}}}\right] \times 10{,}000.$$


a
PANGEA
GONDWANA
RODINIA
NUNA
KENOR
Granulite facies UHT metamorphism
Eclogite-HP granulite metamorphism
HP-UHP metamorphism
Number
Gradient (°C/GPa)
Zircon record (n = 100,445)
b
Normalized 87Sr/86Sr curve of river runoff versus mantle influence
Mean εHf (t) in detrital zircons
Hf
Seawater
c
Crustal reworking rate (%)
δ18O (‰) zircon
Oxygen
Reworked crust
Age (Ga)

crustal reworking is the temporal distribution of crystallization ages of zircons with Hf model ages greater than their crystallization ages (Dhuime et al. 2012). Another is the distribution of the peaks in $\delta^{18}O$ values in zircons through time (**Figure 3**) (Bindeman et al. 2016, Dhuime et al. 2012, Roberts & Spencer 2014, Spencer et al. 2014). Elevated $\delta^{18}O$ values indicate reworking of supracrustal material, and this is most readily achieved in sections of crust thickened in response to continental collision. Both proxies indicate that the periods of increased crustal reworking are also those of supercontinent assembly (**Figure 3**). Thus, this is independent evidence that the peaks of U-Pb crystallization ages are associated with periods of crustal thickening and of continental collision, and the development of supercontinents.

3. SAMPLING THE CONTINENTAL CRUST

Zircons provide a record of the ages of rocks exposed on Earth's surface, and there is reasonable agreement on the composition of the upper and the bulk of the continental crust (Rudnick & Gao 2003; Taylor & McLennan 1985, 1995). The crust can be sampled by analyzing clastic sediments derived by weathering and erosion of different rock units in their source areas. The compositions of clastic sedimentary rocks provide reasonable estimates of the average minor and trace element contents of their source rocks, at least for less soluble elements (McLennan et al. 2005, Taylor & McLennan 1985). However for radiogenic isotopes, the ages of the different source rocks become important, and it is necessary to establish the extent to which source rocks with different ages contribute to the sediments analyzed. Specifically, it is necessary to determine whether the proportions of source rocks in the eroded terrain are similar to those estimated from bulk analysis of the resultant sediment (Allègre & Rousseau 1984). If not, it becomes critical to correct for the bias introduced during sedimentary processes to unravel the age profile of the upper continental crust.

Archean rocks within the drainage of the modern-day Frankland River in southwest Australia are, for example, underrepresented by a factor of approximately six in the sediments sampled at the river mouth (Cawood et al. 2003, Dhuime et al. 2011). The post-Archean rocks are exposed in areas of relatively high relief, and where the river contains larger volumes of water, and so they are likely to contribute disproportionately large volumes of detritus to the Frankland River, with the result that the Archean crust is underrepresented. **Figure 4** shows that, in general, crust with Archean Nd model ages tends to have lower average topographic relief and that the average Nd model age of samples taken from areas with more than 100 m of relief is ~1 Ga (Allan 2014). Given that denudation rates increase with topographic relief (Larsen et al. 2014) and that the average model age of at least the upper continental crust is thought to be ~1.8 Ga (Chauvel et al. 2014), it follows that old rocks are significantly underrepresented in the sedimentary record.

Allègre & Rousseau (1984) addressed the issue of bias introduced by using detrital sediments to sample the continental crust. They defined an erosion parameter K to relate the model age of the sediments analyzed to the average model age of their source rocks. For a source terrain with two units of different model ages, K is the ratio of the material from the younger to the older terrain determined from analyses of the sediment divided by the ratio of the area of the younger relative to the older terrain present in the catchment area, i.e., estimated from a geological map. This has proven difficult to measure in natural systems, and values ranging between two and three have been assumed in previous studies (Allègre & Rousseau 1984, Garrels & Mackenzie 1971, Goldstein & Jacobsen 1988, Jacobsen 1988, Kramers 2002, Kramers & Tolstikhin 1997). Such values for K would suggest that the volume of continental crust at 3 Ga was ~25–30% of the present volume. However, using the value of $K = 15$, obtained from the sediments at the mouth

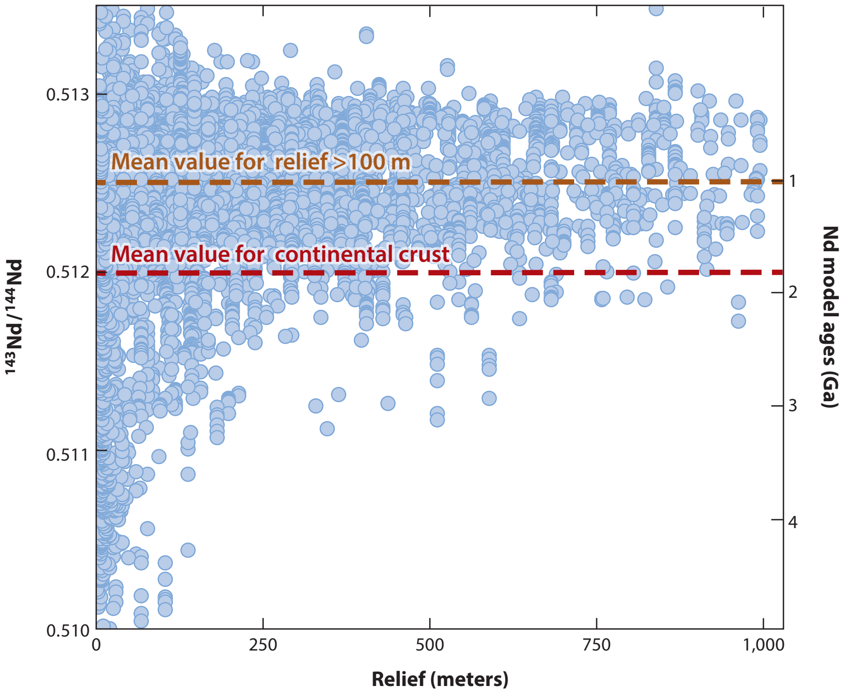


Figure 4

Present-day Nd isotope ratios for whole-rock samples plotted against topographic relief (Allan 2014). The average Nd model age for samples from areas of >100-m relief is ~1 Ga, whereas estimates of the average Nd model age for the continental crust is ~1.8 Ga from Condie & Aster (2010) and Chauvel et al. (2014).

of the Frankland River (Dhuime et al. 2011), suggests that the volume of continental crust present at 3 Ga was 65–70% of the present volume.

4. MANTLE LITHOSPHERE AND ITS RELATIONSHIP TO THE OVERLYING CRUST

The mantle lithosphere is the rigid portion of mantle attached to the overlying crust (**Figure 1**), and so it may preserve a record of upper mantle processes dating back to the formation of the oldest preserved continental crust. It can be mapped by geophysics (e.g., James et al. 2001, Lee & Nolet 1997, McKenzie et al. 2015, McKenzie & Priestley 2008, Mooney et al. 1998), and it is sampled at Earth's surface as fragments transported in younger kimberlites and alkali basalts (e.g., Carlson et al. 2005). Mineral thermobarometry on suites of samples of mantle lithosphere from adjacent localities has been used to reconstruct geotherms within the lithosphere. Typically, the lowest geotherms are found beneath stable cratons where temperatures at 100 km depth are between 750°C and 900°C (Carlson et al. 2005, Finnerty & Boyd 1987, Lee et al. 2011).

Many mantle peridotites have undergone alteration and/or metasomatism, and the Mg number of olivine is taken as the most reliable indication of the melting processes responsible for their initial bulk rock compositions (Pearson & Wittig 2013). The variation of the Mg number of olivine with depth in the lithosphere can be compared with experimental parameterizations of mantle-melting products under different conditions. Peridotite xenoliths from Archean crustal terrains tend to have olivines with higher Mg numbers, and hence, these are more residual in composition than those from massif peridotites, which are tectonically exhumed fragments of the upper mantle (Bodinier & Godard 2003), and those mantle xenoliths from off-craton localities. The xenoliths from Archean areas tend to have lower Fe contents, and hence, these are less dense and therefore more buoyant than those in younger areas. Many of these xenoliths also have elevated silica (Boyd 1989 and references therein, Carlson et al. 2005, Pearson & Wittig 2013). The Archean lithosphere is thicker than younger counterparts; it is less dense and hence harder to destroy than the more Fe-rich mantle, and so it may have had a role in the long-term survival of the Archean continental crust (Davidson & Arculus 2006).

The generation of low-Fe (high-Mg number) residual mantle in the Archean has been the subject of much debate. Walter (1998) demonstrated experimentally that a low-Fe residual mantle can result from the generation of komatiite magmas from fertile peridotite at depths >7 GPa (~250 km). However, Pearson & Wittig (2013) highlighted that the low heavy rare earth element (REE) content, and the fractionation of Lu from Yb, in most mantle xenoliths from cratonic areas can only be generated by extensive melting in the spinel peridotite stability field even though garnet is present as a residual phase in Walter's experiments. Herzberg & Rudnick (2012) invoked a model of non-arc magmatism with mantle potential temperatures of up to ~1,700°C, and melt fractions of 0.25–0.45, at the time of formation of cratonic mantle lithosphere. Variations in olivine Mg number with depth are consistent with models in which the peridotites represent residues after melting at high temperatures at shallow depths, as in a hot ridge setting, and there is relatively little evidence for the even higher Mg number residues than would be expected in a hot mantle plume setting (Herzberg & Rudnick 2012, Pearson & Wittig 2013). Intriguingly in the context of when plate tectonics might have developed, Simon et al. (2007) developed a model for the origins of low-temperature peridotite xenoliths from Kimberley and Lesotho involving shallow ridge melting, followed by subduction-related metasomatism probably during amalgamation of smaller preexisting terranes in the Late Archaean (~2.9 Ga).

Matsukage & Kawasaki (2014) argued that cratonic garnet peridotites could reflect melt depletion from almost anhydrous to a maximum of approximately 1% to 2% water content in the upper mantle at depths of 100 to 200 km. Such elevated water content would imply the presence of subduction, and eclogites are fragments of basalt, which are also interpreted as evidence for subduction. Eclogites have been dated back to 2.9 Ga (Tappe et al. 2011), and many have Os model ages of 2.7–3.0 Ga. Eclogite assemblages are recognized in inclusions in diamonds back to 3 Ga and not before (Shirey & Richardson 2011). Eclogite provides evidence of the return of basalt to the mantle, and although this is commonly attributed to the onset of subduction (Shirey & Richardson 2011), it might also happen through density foundering of crustal material.

Melt depletion events in mantle rocks may now be dated using the Re-Os decay scheme. Walker et al. (1989) defined the Re-depletion model age, T_{RD}, calculated by assuming that the Re/Os ratio of the analyzed samples was zero since the time of melt extraction. This in effect provides a minimum age for the depletion event, as not all the Re may have been removed.

Mantle xenoliths from cratonic areas tend to have older T_{RD} ages, with a peak at 2.8 Ga, and off-craton xenoliths have younger T_{RD} ages with peaks at 2.4 and 1.5 Ga. Spinel peridotites tend to occur in relatively young geological settings, typically in alkali basalts rather than kimberlites, and most T_{RD} ages from spinel peridotites are less than 1.5 Ga (**Figure 5**). It is tempting to assume that mantle depletion events reflect the time of formation of the continental mantle lithosphere, but that is not always the case as some segments of mantle within the asthenosphere also have depleted Os isotope signatures before they are incorporated into the lithosphere (Liu et al. 2015). Nonetheless, as noted by many studies (Carlson et al. 2005, Pearson & Wittig 2013, and references

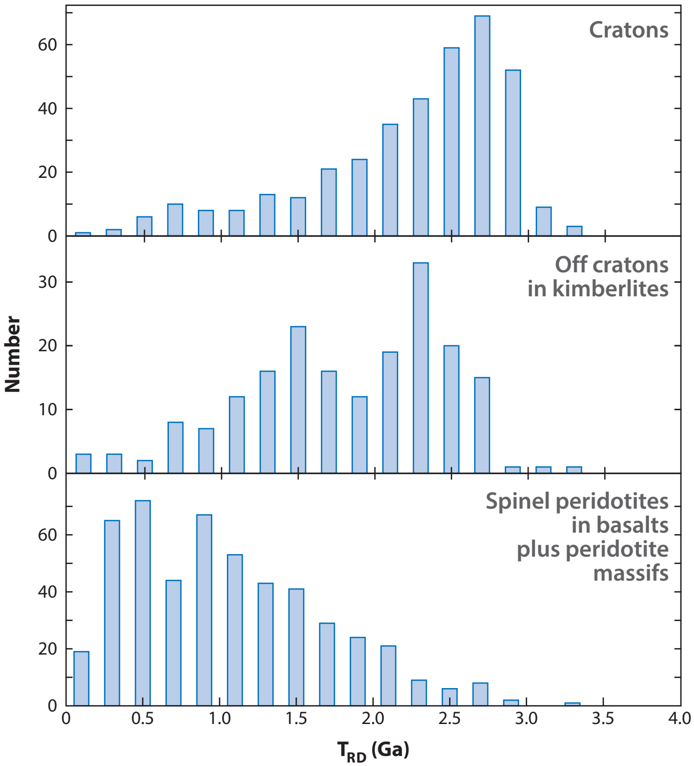


Figure 5

Histogram of Re-depletion (T_{RD}) ages from on and off craton and from spinel peridotites. The data for on craton peridotites are from Pearson & Wittig (2013 and the references therein), and the other sources of data are as follows: Becker et al. (2001, 2006); Burton et al. (2000); Carlson & Irving (1994); Carlson et al. (2004); Chesley et al. (1999); Choi et al. (2010); Gao et al. (2002); Handler et al. (1997, 2003, 2005); Janney et al. (2010); Lee et al. (2000); Liu et al. (2011, 2015, 2016); Luguet et al. (2009); Meisel et al. (2001); Pearson & Wittig (2013); Peslier et al. (2000a,b); Simon et al. (2007); Wu et al. (2003, 2006); Xu et al. (2008); H.-F. Zhang et al. (2009); and Y.-L. Zhang et al. (2011).

therein), mantle rocks with Archean T_{RD} ages predominate in areas of Archean crust, and those with younger T_{RD} ages generally occur beneath younger continental crust.

5. CONSTRAINING THE VOLUME OF CONTINENTAL CRUST AND HOW THIS HAS CHANGED WITH TIME

The issue of the volume of continental crust through time has been a long-standing debate (Armstrong 1981, Gastil 1960, Moorbath 1977, O'Nions et al. 1979). Models of the progressive accretion of mobile belts onto older cratonic cores may have implied linear crustal growth through time (e.g., Dana 1873), and although early compilations of geochronological data appeared to support this concept (e.g., Hurley et al. 1962), in practice they reflected reworking of older crust by younger orogenic events. Other models were based on arguments that Earth has progressively cooled, and so a relatively large volume of crust was generated early and was subsequently destroyed and generated at rates similar to those observed today (Armstrong 1981). However, little of that old crust is preserved (**Figure 6**), and so the debate became linked to the

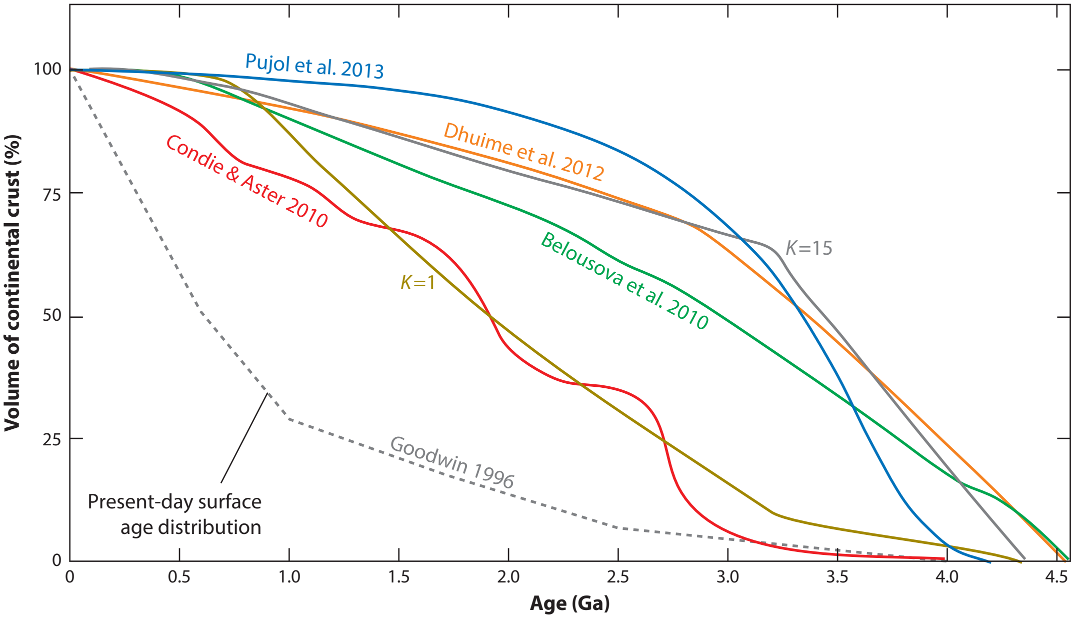


Figure 6

Mean curves from the continental growth models of Belousova et al. (2010), Dhuime et al. (2012), and Pujol et al. (2013) are compared with the distribution of geological ages in presently preserved crust from Goodwin (1996). The Condie & Aster (2010) curve reflects the proportion of juvenile crust of different ages preserved today. The Belousova et al. (2010) and Dhuime et al. (2012) curves are based on the proportions of reworked and juvenile crust in the zircon record. The $K = 1$ curve is the continental growth curve for the Gondwana supercontinent calculated from the Nd isotope data for Australian shales (Allègre & Rousseau 1984), assuming an erosion parameter $K = 1$ (i.e., no preferential erosion of the different lithologies producing the sediment). A K value of unity would imply that 30% of the continental crust was generated by the end of the Archean, but this increases to 75% if $K = 15$ (curve from Cawood et al. 2013). The analyses of detrital zircons (Belousova et al. 2010, Dhuime et al. 2012) are not sensitive to values of K. K influences the proportions of old and young zircons available for analysis, but the crustal growth curves of Belousova et al. (2010) and Dhuime et al. (2012) are based on the proportions of reworked and juvenile crust irrespective of the age distribution of the zircons analyzed.

ease with which continental crust might be destroyed. The advent of Nd and then Hf isotope measurements allowed the age of formation of different segments of the continental crust to be investigated (Allègre & Rousseau 1984, Belousova et al. 2010, Dhuime et al. 2012, McCulloch & Wasserburg 1978, O'Nions et al. 1983). Yet many of the early Nd isotope studies relied on data from shales, and as discussed above in Section 3, older geological terrains are significantly underrepresented in the compositions of terrigenous sediments (**Figure 6**).

The distributions of crust formation ages (**Figure 3*a***) and of the proportions of reworked crust and how that has changed with time (**Figure 3*c***) do not indicate the volume of crust from different periods. Initial models regarded the continental crust as the complementary reservoir to the depleted upper mantle, and investigators sought to ensure that the volume of crust generated at different times would result in a residual depleted mantle reservoir with what was regarded as a reasonable average age of 2.0–1.8 Ga (Allègre et al. 1983, Jacobsen & Wasserburg 1979, Kramers & Tolstikhin 1997, O'Nions et al. 1979). Recent models have focused on the records available from the crust itself; they are anchored by the present-day volume of the crust and then projected back in time based on estimated ratios of the proportions of new and reworked crust for different time slices until the inferred time of initial formation of the continental crust (Belousova et al. 2010, Dhuime et al. 2012).

The proportions of new and reworked crust can be estimated from (*a*) the balance between zircons with Hf model ages close to their crystallization ages and those with model ages significantly older than their crystallization ages (Belousova et al. 2010, Dhuime et al. 2012), and from (*b*) Nd isotope ratios in clastic sediments (Allègre & Rousseau 1984, Dhuime et al. 2011). Approaches based on Nd and Hf model ages involve a number of uncertainties linked to how well the formation age of new continental crust is constrained by the calculation of model ages (Arndt & Goldstein 1987). There are also concerns that for detrital zircons of Archean age it remains difficult to establish whether they have suffered Pb loss, which would result in shifts to younger ages and thence spuriously old model ages (Vervoort & Kemp 2016). Nonetheless, two independent approaches, one based on Hf isotopes in zircons and one based on variations in the Nd isotope ratios of shales, yield similar curves for the changes in the volumes of continental crust through time (the curves labeled Dhuime et al. 2012 and $K = 15$ in **Figure 6**). The distinction is that the approaches of Belousova et al. (2010) and Dhuime et al. (2012) are concerned with the volumes of new crust that were present at different times in the past, whereas the Allègre & Rousseau (1984) model evaluates the proportion of crust with different model ages at the present day and infers that it is a fair reflection of the volumes of new crust that were present at different times in the past. Campbell (2003) used secular changes in Nb/U in mafic rocks as a nonisotope approach to constraining crustal volumes and concluded that by ~3 Ga the volume of continental crust was ~70% of the present volume of the continental crust. Pujol et al. (2013) reached a similar conclusion based on estimates of Archean atmospheric $^{40}Ar/^{36}Ar$ from measurements of Ar in 3.5-Ga hydrothermal quartz. Such values of ~70% of the crust at 3 Ga are significantly higher than the volume of such crust preserved at the present day [<5% from the Condie & Aster (2010) curve in **Figure 6**], which in turn highlights the question of how and when such volumes of old crust were destroyed (see Section 6.3).

6. SYNTHESIS

Data from field studies along with paleomagnetic, geochemical, and ore deposit studies indicate a change in the character, composition, and thickness of the continental lithosphere in the latter half of the Archean from ~3.2 Ga (**Figure 2**) (Cawood et al. 2006, Dhuime et al. 2012, Keller & Schoene 2012, Tang et al. 2016). There are no significant peaks of zircon crystallization ages older than 3 Ga (**Figure 3**), and the development of potassic granites from 3.1 to 2.6 Ga and main peak of old T_{RD} ages (3.0–2.7 Ga) (**Figure 5**) are considered to mark the stabilization of the continental lithosphere (Anhaeusser & Robb 1981, Carlson et al. 2005, Schoene et al. 2009). Recent curves for the volume of continental crust through time (**Figure 6**) are also marked by a change in slope at approximately 3 Ga, reflecting a change in the estimated growth rate of the continental crust. Such changes presumably happened over a period of time and at different times in different places. Nonetheless, 3 Ga is a useful time marker, and here, we discuss some of the differences that characterize geological terrains older and younger than 3 Ga.

6.1. Before 3 Ga

The oldest terrestrial zircons, from the Jack Hills area in Western Australia, indicate that there were felsic magmas and surface water from ~4.4 Ga (Valley et al. 2002, Wilde et al. 2001). There is, however, much controversy over the setting in which these Hadean zircons were generated and thus their significance for continental crustal growth (e.g., Harrison 2009, Kemp et al. 2010).

The oldest known terrestrial rocks, the Acasta gneisses, have components dated at ~4.02 Ga (Bowring & Williams 1999). At least in some Archean areas, pre-3.2 Ga crust is dominated by dome and basin granite-greenstone terranes, thought to be indicative of vertical tectonics (Bédard et al. 2013, Van Kranendonk et al. 2007). Willbold et al. (2009) and Reimink et al. (2014) noted that Archean tonalitic gneisses are similar in composition to intermediate lavas generated in Iceland at the present day and invoked similar models linked to mantle upwelling. Pre-3.1 Ga upper crust also appears to have been more mafic than that generated subsequently (Tang et al. 2016, Taylor & McLennan 1985). Taylor & McLennan (1985) reported a sharp increase in Th/Sc in fine-grained sedimentary rocks at the end of the Archean. Tang et al. (2016) demonstrated that Archean shales and glacial diamictites have strikingly high Ni/Co and Cr/Zr ratios compared to younger sedimentary rocks. Both ratios correlate with MgO contents in igneous rocks, and Tang et al. (2016) concluded that the average upper crust, as sampled by these sediments, had ∼15% MgO at 3.2 Ga and that this had dropped to ∼4% by 2.6 Ga.

Geochemical studies on more mafic lithologies suggest that many of the older rock suites from Barberton, Pilbara, and West Greenland have affinity with modern within-plate magmas (Campbell 2003, Jenner et al. 2013, Næraa et al. 2012, Smithies et al. 2007, Van Kranendonk et al. 2007). Old mafic rocks with chemical similarities to recent subduction-related suites have also been reported, and at least for the period 3.85–3.2 Ga, they appear confined to West Greenland and Canada (Jenner et al. 2009, O'Neil et al. 2011). Thus, the magmatic records appear different in different places. In southern Africa and the Pilbara, there appears to be a shift from early within-plate-like magmas and an emphasis on vertical tectonics (Campbell 2003, Sandiford et al. 2004, Smithies et al. 2007, Van Kranendonk et al. 2007) to magmatic rocks with subduction-related signatures after ∼3.1 Ga (Smithies et al. 2007). In the Superior province, the older rocks have subduction-related affinities, and in West Greenland there are >3.8 Ga rocks with both subduction-related and within-plate geochemical signatures (Jenner et al. 2009, 2013). At present it appears that Archean areas with the classic dome and basin structures, as in the East Pilbara and southern Africa, are more likely to have within-plate magmatic rocks, associated with dominantly vertical tectonic regimes (e.g., Sandiford et al. 2004).

Very few rocks, and even crystals of zircon, have survived from before ∼3.9 Ga, a time that marks the end of the period of heavy meteorite bombardment (**Figure 2**). Yet Earth would have had a crust before this, and it is likely that it would have been thick enough to have remelted and so developed a felsic component (Kamber 2015, Kemp et al. 2010). Marchi et al. (2014) rescaled the lunar meteorite flux to Earth and argued that the surface of the Hadean Earth was widely reprocessed by impacts through mixing and burial by impact-generated melt. Hansen (2015) took the argument further and contrasted Archean granite-greenstone terrains on Earth with crustal plateaus now recognized on Venus. In Hansen's (2015) model, Archean cratons are formed in response to a large bolide-piercing thin Archean lithosphere, causing localized high-temperature, high-fraction partial melting in the sublithospheric mantle and the melt rises, forming an igneous province that evolves to form a granite-greenstone terrain, whereas the melt residue develops a complementary mantle lithosphere.

Such models remain difficult to test, but the available evidence suggests that there was both subduction-related and what is typically termed within-plate magmatism at similar times in different terranes in the early Archean, before subduction-related magmatism became more dominant after ∼3.2 Ga.

6.2. Changes About 3 Ga

The geological record is characterized by peaks and troughs in the distribution of U-Pb zircon crystallization ages from 2.7 Ga to the present (**Figure 3**). The peaks of ages are taken to reflect periods of continental collision and crustal thickening in the development of supercontinents, even though the age and perhaps even the existence of the oldest suggested supercontinent remains a matter of debate (Bleeker 2003). Continental collision requires convergence and a role for plate tectonics (but see Bédard et al. 2013 for an alternative mechanism), and a similar story is structurally available. Rectilinear, thrust-bound belts characterize late Archean terranes, and in the Pilbara, they date from ∼3.1 Ga (Smithies et al. 2007, Van Kranendonk et al. 2007). Large-scale thrust faults, reflecting horizontal compression and dated approximately 2.7 Ga, have been mapped in West Greenland (Friend & Nutman 2005), and a seismic reflection image of an inferred paleosubduction zone has been ascribed to the 2.7 Ga docking of the northern Abitibi granite-greenstone and Opatica plutonic belts, Superior craton Canada (Calvert 1995), both of which are consistent with the commencement of plate tectonics by this time (Cawood et al. 2006).

Geochemical studies offer insights into the conditions of melt generation and the tectonic setting in which new crust was generated. Mafic rocks show a marked increase in La/Yb at ∼2.7 Ga, a secular decrease in compatible element contents, and an increase in incompatible element contents that reflect a decrease in the degrees of mantle melting through time (Keller & Schoene 2012). Archean terrains are marked by a strongly bimodal distribution in silica (Hawkesworth et al. 2016, Kamber 2015), and this is less well developed in younger areas. Archean felsic rocks are characterized by higher Na_2O/K_2O and La/Yb and less marked Eu anomalies than younger rocks (see Taylor & McLennan 1985). These differences are attributed to partial melting of mafic source rocks at depths in the stability field of garnet rather than plagioclase, likely in response to over thickening of the mafic crust (Kamber 2015, Kemp et al. 2010). Collerson & Kamber (1999) noted an increase in Nb/U and Nb/Th in depleted mantle-derived rocks at ∼3 Ga, which they linked with the onset of subduction at that time and the extraction of low Nb/U and Nb/Th material.

An alternative approach is to reconstruct the composition of juvenile continental crust using the initial isotope ratios of preserved continental rocks of different ages. Dhuime et al. (2015) note a marked increase in the Rb/Sr ratio of inferred new crust, and hence of its silica content, at about 3 Ga (**Figure 7**). Prior to 3 Ga, the Rb/Sr ratios of new continental crust are highly scattered, but the median is ∼0.03, and it increases to a maximum value of ∼0.08 from 3.0 to 1.7 Ga before decreasing to values of ∼0.065 in approximately the last 1 Ga. There is a strong positive correlation between Rb/Sr and the SiO_2 contents of crustal rocks and in recent geological settings with the thickness of the continental crust (**Figure 7**) (Dhuime et al. 2015). It was concluded that the SiO_2 contents of new continental crust increased from ∼48% before 3.0 Ga to more intermediate compositions in younger terrains with SiO_2 up to ∼57%.

The Rb/Sr ratios, and SiO_2 contents, of recent magmatic rocks increase with crustal thickness in Central and South America, and the increase in Rb/Sr appears to be linked to increased differentiation in areas of thicker continental crust (Dhuime et al. 2015). Although the rate of increase of Rb/Sr with crustal thickness may have been different earlier in Earth history, the progressive increase in the time-integrated Rb/Sr ratios of the new crust from 3 Ga to 1 Ga (**Figure 7**) is taken to reflect at least the relative thickening of the continental crust through time from ∼20 km at 3 Ga to ∼40 km by 1 Ga, before it decreases to ∼30 km toward the present. Assuming that thickness is a proxy for crustal volume, the decrease in inferred crustal volumes presumably reflects crust destruction rates being higher than crust generation rates, perhaps since the onset of cold subduction (**Figure 3*a***) (Brown 2007, 2014). The inferred increase in crustal thickness from 3 Ga to 1 Ga is accompanied by a global increase in the sedimentary contribution to the magmatic record reflected in higher $\delta^{18}O$ since the late Archean (Valley et al. 2005) and by increasing $^{87}Sr/^{86}Sr$ isotope ratios in seawater from 3 to 2 Ga (Shields 2007). It therefore indicates when significant volumes of continental crust emerged (Arndt 1999; Flament et al. 2008, 2013; Pons et al. 2013; Rey & Coltice 2008) and were available for surficial weathering with the associated drawdown of CO_2 and modification of the Earth's atmosphere (Kramers 2002, Lee et al. 2016).

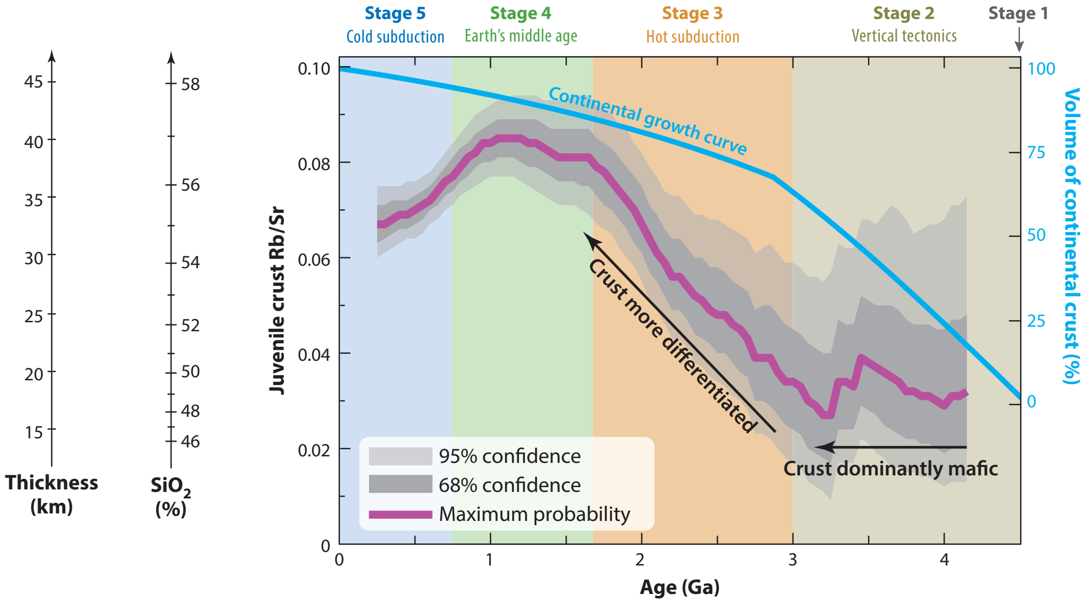


Figure 7

The estimated Rb/Sr ratios of juvenile continental crust plotted against the age of crust formation for ∼13,000 whole-rock analyses (Dhuime et al. 2015). Rb/Sr increases with both whole-rock silica content and crustal thickness at the site of magma generation. The calculated volumes of continental crust from Dhuime et al. (2012) are presented for comparison. Stages of Earth evolution: Stage 1 is the early accretion of Earth, which may have persisted for 5–10 Myr (Elkins-Tanton 2008); stage 2 is dominated by vertical tectonics; stage 3 is dominated by hot subduction; stage 4 is Earth's middle age; and stage 5 is characterized by cold subduction.

It is noteworthy that magmatism with within-plate and subduction-like chemical signatures occurred at different times in different Archean areas, and a temporal shift from within-plate to subduction-related magmatism is better developed in some areas than others. However, the changes in the rates of growth of continental crust (**Figure 6**), the ages of the oldest mantle eclogitic samples (**Figure 2**), and the increasing emergence of the continental crust (**Figures 3** and **7**) are all taken to reflect the dominant role of plate tectonics by ∼3 Ga. Subduction-related magmatism occurred before 3 Ga, and new continental crust has been generated in within-plate settings since 3 Ga, such that the bulk composition of the continental crust may contain a ∼20% contribution of material generated in intraplate settings (see discussion in Rudnick 1995). The first preserved record of a subaerial, large igneous province-related magmatism occurs approximately 3 Ga and becomes widespread by 2.5 Ga (Kump & Barley 2007).

6.3. Preliminary Model for a Two-Phase Evolution of the Continental Crust

Key aspects of the model for the evolution of the continental crust outlined in this contribution are that (*a*) the composition of new continental crust (Dhuime et al. 2015) and the upper crust (Tang et al. 2016) was more mafic before 3 Ga than subsequently; (*b*) at ∼3 Ga the volume of continental crust was ∼70% of its present volume; (*c*) there was a significant change in the growth rate of the continental crust at ∼3 Ga, which has been attributed to the onset of plate tectonics

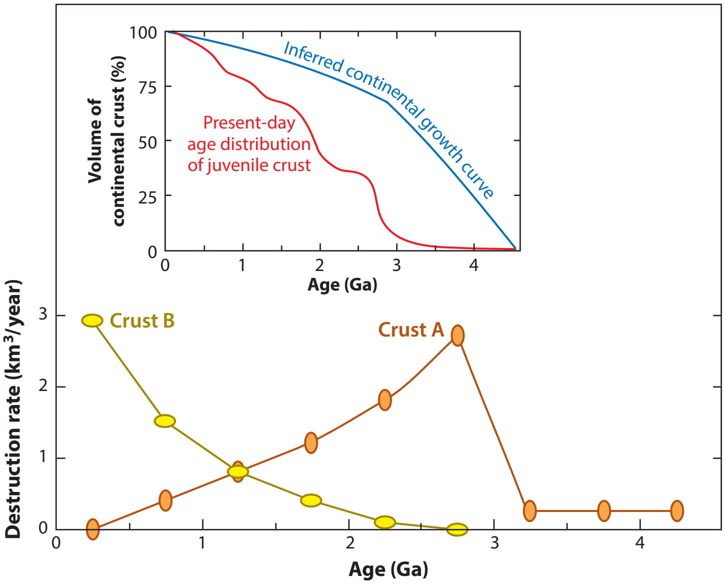


Figure 8

Schematic illustration of a preliminary two-part model for the generation and evolution of the continental crust. Crust A is the (more mafic) crust generated before 3 Ga (**Figure 7**), and crust B is the more differentiated crust generated subsequently. The simple box model starts with the continental crust growth curve (inset, *blue curve*; Dhuime et al. 2012), and crust destruction rates for crusts A and B were calculated for 500-Myr increments to match the present-day age distribution of juvenile crust (*red curve*; Condie & Aster 2010). The crust generation rates are held constant, and the calculated changes in crust destruction rates for crust A and B are shown.

as the dominant tectonic regime; and (*d*) less than 5% of the present volume of the continental crust consists of rocks older than 3 Ga.

These observations are incorporated in a simple two-part model of evolution for the continental crust (**Figure 8**) (Dhuime et al. 2016). This model is not unique, but it illustrates some of the implications that need to be considered. The continental crust generation rates were taken to be constant (e.g., Dhuime et al. 2012), but the composition of the crust changed with time. Juvenile continental crust, prior to 3 Ga, was dominantly mafic (crust A), and from ~3 Ga, juvenile crust became more differentiated in composition (crust B), as shown by the inferred increase in the SiO_2 content of juvenile crust plotted in **Figure 7**. This change is linked to the global onset of plate tectonics as the dominant setting of crust generation and the transition to a modern style of more differentiated continental crust associated with convergent margin settings. In the current model, the crust destruction rates before 3 Ga are taken to be ~10% of those at the present day, and these increased once plate tectonics became the dominant setting of crust generation at 3 Ga. The model has a high-destruction rate of crust A between 3 and 2.5 Ga to account for the low volume of pre-3 Ga crust preserved now, which is consistent with the predominantly mafic and hence relatively dense composition of the pre-3 Ga crust. This is followed by a rapid decrease in the destruction rates of that early crust and an increase in the destruction rates of crust B progressively from the Archean-Proterozoic transition to reach a maximum value at the present day (**Figure 8**). The volume of crust destroyed to account for the crustal volume at 3.0 Ga being

70% of the present-day value, and <5% of crust older than 3.0 Ga preserved at the present day, indicates that about twice the present volume of the continental crust has been recycled back into the mantle since the onset of plate tectonics (Hawkesworth et al. 2016).

6.4. Earth's Continental Lithosphere Through Time

The geological record preserved today is fragmentary, biased by tectonic and sedimentary processes, and many segments from old oceanic to early Archean lithosphere are barely represented. There are peaks of mantle depletion (T_{RD}) and zircon U-Pb crystallization ages at ~2.8 Ga and younger (**Figures 3** and **5**). Older ages are recorded in diamonds and in crustal rocks and zircons (**Figure 2**), but there is simply not the record for a meaningful comparison of crustal and mantle processes before ~3 Ga. It is argued that plate tectonics was the dominant tectonic process involved in the generation of the continental crust by ~3 Ga as evidenced in the reduction of crustal growth rate, which is attributed to the increased destruction of continental crust along destructive plate margins; the ages of the oldest eclogitic fragments; and changes from more mafic to more felsic compositions of new continental crust. The supercontinent cycle involving dispersal and assembly of continental crust through the opening and closing of ocean basins, at least as currently understood for Pangea and Gondwana, is driven by plate tectonics. Thus, if the early continental record preserves rocks that predate the start of plate tectonics, it is inferred that these rocks should not display the preservation bias associated with the development of supercontinents.

Based on changes in the character of the preserved rock record through time along with constraints provided by numerical modeling and the secular evolution of the mantle's thermal history, we envisage five main stages in the evolution of Earth. The boundary between each stage corresponds with a change in the character of the lithosphere and the tectonic processes controlling its development (Hawkesworth et al. 2016). These five stages are (*a*) initial accretion of Earth; (*b*) generation of crust prior to 3.0 Ga in a preplate tectonic regime; (*c*) early plate tectonics involving hot subduction at relatively high geothermal gradients and extending until approximately 1.7 Ga; (*d*) Earth's middle age from 1.7 to 0.75 Ga, characterized by apparent lithospheric, environmental, and evolutionary stability; and (*e*) initiation of modern cold subduction (i.e., at relatively low geothermal gradients) at ~0.75 Ga. The change from stage 2 to 3 is also associated with the change in the composition of new continental crust (**Figure 7**).

Stage 1 included the initial accretion of Earth, the differentiation of the core and mantle, the development of a magma ocean, and development of an undifferentiated mafic protocrust. It may have persisted for 5–10 Myr (Elkins-Tanton 2008). The second stage extended until approximately 3 Ga. Little lithospheric record remains of the first 600 Myr owing to widespread meteorite bombardment, including collision between the proto-Earth and a body inferred to be approximately the size of Mars and sometimes referred to as Theia, which resulted in the formation of the Moon (Jacobson et al. 2014, Young et al. 2016). The significant scale and frequency of meteorite impact terminated with what is referred to as the end of the late heavy bombardment at approximately 3.9 Ga (Gomes et al. 2005, Marchi et al. 2014). Potential mantle temperatures are inferred to have been ~250°C hotter than the present-day average convecting asthenospheric temperature of ~1,350°C at the end of the late heavy bombardment of meteorites (Herzberg et al. 2010, Korenaga 2013). Numerical modeling suggests widespread asthenospheric melting and impregnation of the lithosphere at such temperatures, reducing lithospheric rigidity and inhibiting, and perhaps preventing, development of sustained deep-level lithospheric subduction (Sizova et al. 2010, van Hunen & van den Berg 2008). Zones of mantle upwelling and melting constituted sites of lithospheric generation, with recycling back into the mantle inferred to occur through lithospheric delamination (Fischer & Gerya 2016, Johnson et al. 2014, Sizova et al. 2015).

The lack of a garnet signature in the heavy REE variations in residual peridotites and the profiles of olivine composition versus depth for different cratonic lithosphere sections in which the Mg number tends to decrease with increasing depth are taken to reflect melting of a relatively hot mantle at shallow levels (e.g., Canil 2004, Carlson et al. 2005, Helmstaedt & Schulze 1989, Herzberg & Rudnick 2012, Kelemen et al. 1998, Lee et al. 2011, Simon et al. 2007). This combination of high temperatures and shallow depths suggests that melting took place beneath the oceans, that the complementary oceanic crust was destroyed, and that convergent margin processes were involved in the accretion of the residual mantle material to the continental lithosphere, presumably at the time of, or soon after, the stabilization of the overlying crust. It further suggests that there may be no simple relationship between the mantle lithosphere and the overlying crust, at least at this time (see also Sizova et al. 2015).

It is striking that the degrees of melting invoked to generate the residual mantle lithosphere (25–40%) (Herzberg & Rudnick 2012, Walter 1998) are significantly greater than those required to generate the fractionated Lu/Hf and Sm/Nd ratios of mafic continental crust (as subsequently sampled in the magmas from which zircons crystallized), which were ~15% for a simple batch melting model. We infer that the higher degrees of melting took place under relatively thin lithospheric lids in the oceans, whereas the lower degrees of melting involved in the generation of juvenile mafic continental crust reflect melt generation beneath thicker lids in continental areas. The lack of significant Rb/Sr fractionation between the mantle and new continental crust, while accompanied by 15–25% fractionation of Lu/Hf and Sm/Nd, appears to have continued until 3 Ga, whereupon the composition of new crust became more differentiated (**Figure 7**), and the degrees of melting involved in the generation of new continental crust decreased.

The presence of surface water on Earth before 4 Ga has been inferred from slightly elevated $\delta^{18}O$ values in Hadean detrital zircon (Cavosie et al. 2005, Harrison 2009, Valley et al. 2002, Wilde et al. 2001). Water-lain sediments in the Isua greenstone belt (Greenland) attest to the presence of oceans by at least 3.7 Ga (Bolhar et al. 2004). Marine carbonates deposited at 3.48 Ga in the Pilbara craton (Western Australia) preserve evidence for cyanobacteria (Van Kranendonk et al. 2008), which implies photosynthesis and hence start of oxygenation of the atmosphere.

The stabilization of Archean cratons and the change in crustal growth rates at about 3.0 Ga are taken to mark the onset of stage 3, which extended until approximately 1.7 Ga. Mantle temperatures had cooled sufficiently to enable subduction of coherent semirigid plates (Sizova et al. 2010); water was returned to the mantle, which may in turn have facilitated plate tectonics (Korenaga 2013); and the composition of juvenile crust had become more intermediate in composition (**Figure 7**). Plate tectonics became the dominant regime in which new continental crust was generated (**Figure 9**), the continental crust became thicker and more evolved, supercontinent cycles developed, and there was increased erosion to the oceans due to thickening and subaerial exposure of continental crust resulting in an increase in Sr isotope ratios in seawater (**Figure 3**). If ~70% of the present volume of the continental crust was present at 3.0 Ga, and <5% are preserved at the present day (**Figure 6**), it follows that large volumes of continental crust have been destroyed, and yet the overall net growth of crust requires the rate of crust generation to exceed that of recycling (**Figure 8**). The change from a nonplate tectonic Earth to one in which plate tectonics dominated was unlikely to have been abrupt (Section 6.2). If (*a*) the inferred changes in composition and thickness of the lithosphere and (*b*) the linked changes in the surficial Earth such as ocean composition are correct and correspond to secular mantle cooling, then the transition from a predominantly nonplate tectonic Earth is likely to have extended over hundreds of millions of years about 3.0 Ga. Furthermore, rock relations

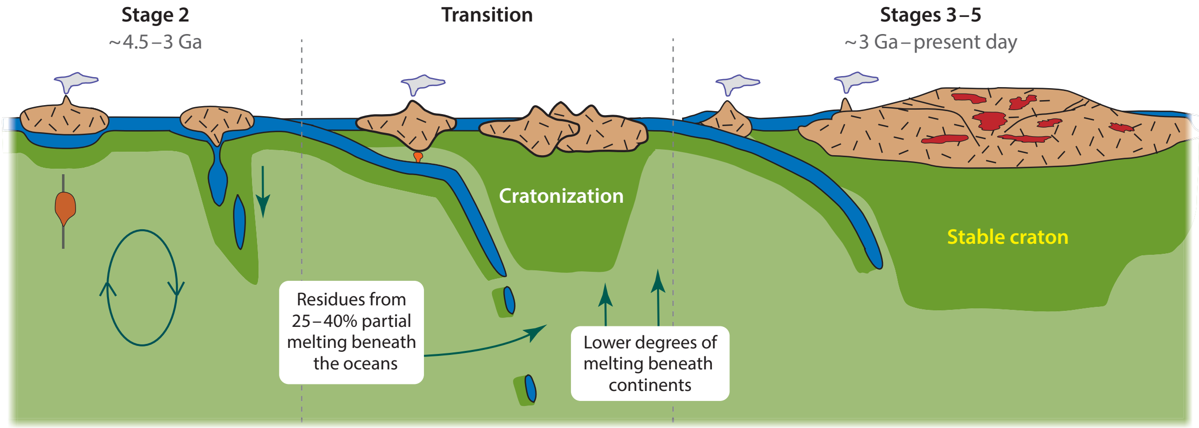


Figure 9

Schematic depiction of changing tectonic processes controlling the evolution of the lithosphere from an early Earth dominated by nonplate processes (stage 2) to one in which plate tectonics is the main mechanism for the generation and recycling of lithosphere (stages 3–5). These changes are a response to the secular cooling of the mantle and the consequent increase in lithospheric strength and rigidity. The mantle xenolith record suggests that they represent residues (from relatively shallow partial melting of hot mantle beneath the oceans), which subsequently accreted beneath continents. In contrast, the mafic crust that is the source of Archean tonalite-trondhjemite-granodiorites has relatively low Lu/Hf and Sm/Nd ratios, implying lower degrees of partial melting.

similar to those in modern convergent plate margins in some regions suggest that at least localized subduction commenced even earlier (Jenner et al. 2009, O'Neil et al. 2011, Turner et al. 2014).

The onset of plate tectonics resulted in significant changes in the nature of Earth processes preserved in the rock archive during the remainder of stage 3. Oxygen levels in the atmosphere increased significantly in the early Proterozoic during the Great Oxidation Event (Holland 2006, Kump 2008 and references therein). A rise in the atmospheric oxygen level followed on from significant increases in continental lithosphere volume and thickness, the change to a more felsic continental crustal composition, and widespread exposure of the continental crust. This is consistent with a causative relationship, and recently Lee et al. (2016) have linked these changes to a decrease in the oxidative efficiency of Earth's surface, allowing a rise in O_2 levels. Global compilations of $\delta^{18}O$ values (**Figure 3**) in zircon display a pattern of peaks and troughs coincident with zircon U-Pb age distribution patterns, except for the ~2.7 Ga age peak, which has only slightly elevated $\delta^{18}O$ values (Dhuime et al. 2012, Spencer et al. 2014). The deviation of $\delta^{18}O$ values from mantle values is taken to reflect increased incorporation of high $\delta^{18}O$ sediment into igneous zircons during crustal reworking and thickening, associated with periods of continental collision and supercontinent assembly, especially since the beginning of the Proterozoic.

The fourth stage is from 1.7–0.75 Ga, recently referred to as Earth's middle age (Cawood & Hawkesworth 2014) and before that as the boring billion (Holland 2006). It is characterized by an absence of significant anomalies in the paleoseawater Sr isotope record and in Hf isotopes in detrital zircon (**Figure 3**), indicating that it was a period of environmental and lithospheric stability (Brasier & Lindsay 1998), and there is a paucity of preserved passive margins (Bradley 2008). This stability may have been linked to the persistence of a relatively stable continental assemblage initiated during assembly of the Nuna supercontinent by ~1.7 Ga and until breakup of its successor, Rodinia, approximately 0.75 Ga (Cawood & Hawkesworth 2014, Cawood et al.

2016). This in turn marks the onset of stage 5 and the development of cold subduction recognized by the first appearance of high- to ultrahigh-pressure metamorphic rocks (Brown 2006, 2014). Falling mantle temperatures, to no more than 80–100°C greater than today, enabled deeper levels of slab break-off in collision zones and the resultant greater depths to which continental crust was subducted prior to exhumation, allowing the development of ultrahigh-pressure metamorphic assemblages (Brown 2006, Sizova et al. 2014). Stage 5 is marked by a strongly episodic distribution of ages linked to the supercontinent cycles of Gondwana and Pangea (**Figure 3**). The estimated thickness, and by analogy the volume, of continental crust decreased perhaps because crust was more efficiently destroyed after the Neoproterozoic transition to the modern plate tectonics regime at relatively low geothermal gradients (Brown 2007). Oxygen levels in both the atmosphere and deep oceans increased, and phosphate and evaporite deposits became widespread, providing a major stimulus for metazoans evolution.

7. WAYS FORWARD

Major developments in geochemical techniques have resulted in a large number of high-quality in situ analyses for radiogenic and stable isotopes in mineral archives that yield precise crystallization ages. High-quality U-Pb zircon analyses, together with T_{RD} Os isotope ages, have allowed for improved statistical approaches in the analysis of variations in the age and composition of the continental crust and its mantle lithosphere and provided insight into the secular evolution of processes involved in the generation and recycling of the lithosphere. There are, however, concerns over biases in the composition and the ages of the rocks preserved, and this highlights the long-standing issue of how best to evaluate the global context of more regional case studies. This remains of particular concern for studies of Archean geology given the relatively small volume of rock that is preserved, but one way to put regional case studies in perspective is in the development of global models of, for example, rates of crustal growth. There is evidence for subduction-related magmatism before 3 Ga, and clearly for within-plate magmatism subsequently, yet the inferred change in global crustal growth rates can be used to argue that plate tectonics became the dominant framework for the generation of continental crust from ∼3 Ga. Geodynamic models may also have a particular role as they allow us to make general predictions that are difficult to derive from the very limited (and potentially unrepresentative) rock record.

Other aspects to highlight include improved methods to evaluate the volume of less-differentiated continental crust, i.e., that contains little or no zircon, and the composition and the tectonic setting of new continental crust, for example, by evaluation of the Rb/Sr and the U/Pb ratios in the sources of crustal melts (Delavault et al. 2016, Dhuime et al. 2015). Given the concerns of how representative the rock record may be, and the biases present in the sedimentary record, there is a clear need for better constraints on a global value for the erosion parameter K (Allègre & Rousseau 1984) and for improved ways to sample well-mixed reservoirs, such as the upper mantle and the oceans. Improved resolution of mantle anisotropy offers new insights into possible structural differences between different cratonic areas and perhaps the timing of the development of such fabric in the upper mantle (e.g., Adam & Lebedev 2012, Kaczmarek et al. 2016). These approaches may determine the extent to which there are tectonic links between the continental mantle lithosphere and the overlying crust (Griffin et al. 2009, Pearson & Wittig 2013). There is increasing agreement that at least some of the rock record reflects conditions before plate tectonics was the dominant tectonic setting in which new continental crust was generated and reflects the nature of that transition. This too appears to be a fertile area to integrate geodynamic modeling and well-constrained data from the rock record.

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