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GR Focus Review

Continental crustal volume, thickness and area, and their geodynamic implications

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ABSTRACT

Models of the volume of continental crust through Earth history vary significantly due to a range of assumptions and data sets; estimates for 3 Ga range from < 10 % to > 120 % of present day volume. We argue that continental area and thickness varied independently and increased at different rates and over different periods, in response to different tectonic processes, through Earth history. Crustal area increased steadily on a pre-plate tectonic Earth, prior to ca. 3 Ga. By 3 Ga the area of continental crust appears to have reached a dynamic equilibrium of around 40 % of the Earth's surface, and this was maintained in the plate tectonic world throughout the last 3 billion years. New continental crust was relatively thin and mafic from ca. 4-3 Ga but started to increase substantially with the inferred onset of plate tectonics at ca. 3 Ga, which also led to the sustained development of Earth's bimodal hypsometry. Integration of thickness and area data suggests continental volume increased from 4.5 Ga to 1.8 Ga, and that it remained relatively constant through Earth's middle age (1.8-0.8 Ga). Since the Neoproterozoic, the estimated crustal thickness, and by implication the volume of the continental crust, appears to have decreased by as much as 15 %. This decrease indicates that crust was destroyed more rapidly than it was generated. This is perhaps associated with the commencement of cold subduction, represented by low dT/dP metamorphic assemblages, resulting in higher rates of destruction of the continental crust through increased sediment subduction and subduction erosion.

1. Introduction

Formation and evolution of the Earth's crust and lithospheric mantle are driven by thermal energy from the planet's interior, modulated by external forces (e.g., solar radiation), which regulates the temperature in the exogenic envelope. The evolving thermal budget of the Earth has resulted in a multistage history of crustal evolution (Hawkesworth et al., 2017), recorded by progressive changes in the composition of the lithosphere, its differentiation into continental and oceanic variants, and its interactions with the mantle, hydrosphere, atmosphere and biosphere.

Understanding interactions within and between the different elements of the Earth system are limited by uncertainty in the proportion of continental to oceanic lithosphere, especially for the early Earth, due to increasing gaps in the preserved rock archive and uncertainty as to tectonic models of lithosphere generation (Cawood et al., 2013; Cawood et al., 2018; Lenardic, 2018; Stern, 2018). For example, whether the volume of continental crust was < 10 % or > 100 % of its current volume at 3 Ga (Fig. 1), and whether it was emergent or submerged, has significant implications for the surficial elements of the Earth system (oceans, atmosphere and biosphere) as well as for models of secular change in mantle temperature (e.g., Herzberg et al., 2010; Lee et al., 2016). This uncertainty reflects the different methods used to constrain the proportion of continental crust. In this paper, we argue that continental area and thickness increased at different rates and at different times, and by different mechanisms. We explore the implications of estimated crustal volumes, the thickness of crust at times of crust generation, the depths of melt generation of TTG (tonalite-trondhjemite-granodiorite) and potassic granites in the Archean, and the constraints from continental freeboard, to evaluate the area of continental crust, and when the dichotomy of oceanic and continental crust was established on Earth.

2. Crustal volume

Figure 1 shows a number of models for volume of continental crust through time and the estimated preserved present day distribution of exposed rock ages from Goodwin (1996). Temporal changes in the volume of crust between models reflects different data sets and assumptions used in their construction (Hawkesworth et al., 2018). All are determined relative to the present volume of the continental crust. However, some curves are based on present day distributions and hence do not account for the current paucity of the early rock record, which in part reflects selective preservation and recycling of material back into the mantle (Hawkesworth et al., 2009; Spencer et al., 2015). Such curves include those in which the volume of crust is based on present day age and thickness data (Artemieva, 2006), Nd isotope ratios in Australian shales (Allègre and Rousseau, 1984), and presently preserved volumes of rocks with different Nd or Hf crust formation ages (Condie and Aster, 2010). A second set of curves attempts to estimate the volumes of continental crust at different times in Earth history, independent of the relative volumes preserved today. These are based on the proportions of reworked and juvenile crust in the zircon record (Belousova et al., 2010; Dhuime et al., 2012; Roberts and Spencer, 2015), modelled secular evolution of atmospheric argon based on measurements of $^{40}Ar/^{36}Ar$ in fluid inclusions in 3.5 Ga quartz (Pujol et al., 2013), and Nb/U ratio of Archean basalts and komatiites derived from the mantle, along with modelled secular variations (Campbell, 2003). A limitation with all these approaches is that the curves are based on cumulative growth of the crust and sum to unity at the present day, and hence no curve can have a past volume greater than the current volume. Nonetheless, the agreement between the second group of curves, modelled on such different data sets, remains striking.

The curves of Fyfe (1978) and Armstrong (1981) represent a third group in which the curves are more schematic in character but they were amongst the first to highlight that growth not only involved the generation of continental crust through the extraction and crystallization of magma from the mantle, but also its recycling back into the mantle. The paucity of preserved old continental crust, as represented in the present day age distribution, reflects this recycling process.

3. Crustal Thickness

The present day thickness of the continental crust reflects a) its original thickness at the time it is generated; b) thickness at the time of its stabilization into the long term geological archive, and; c) the effects of post-stabilization compression, extension, and/or destruction (Fig. 2) (Kusky et al., 2014; Lee and Chin, 2014; Lee et al., 2011; Snyder et al., 2017; Zhu et al., 2012). The age gap between each of these phases can vary from millions to hundreds of millions of years or more. It may therefore be difficult to establish how the present day thickness of Archean and Proterozoic crust compares with the thickness closer to their time of crust generation.

Geophysical studies of Archean terranes indicate present day crustal thickness estimates of some 40 km (Fig. 3A, red line; Artemieva, 2006; Durrheim and Mooney, 1991; Galer and Mezger, 1998; Mooney et al., 1998; Taylor and McLennan, 1985). These estimates, along with isostatic arguments built around the preservation of a thin veneer of Archean sedimentary rocks on ancient shields, have been used to argue that the currently exposed crust has been at, or near, sea-level for much of its existence, and hence of relatively constant crustal thickness (Armstrong, 1981; Wise, 1974). Such approaches assume or imply that the composition of Archean and post-Archean crust is similar, and while that might be correct for late Archean terrains, there is increasing evidence that it may not be valid for the early to mid-Archean (Dhuime et al., 2015; Lee et al., 2016; Tang et al., 2016). The dominant component of Archean continental crust is TTG, and most models concur that these are generated by partial melting of hydrated basalt at depths of 35-60 km (Martin and Moyen, 2002; Moyen and Laurent, 2018; Moyen and Martin, 2012). It is not clear that depths of melting are a robust measure of the thickness of the overlying crust, although the shallowest depth estimate might be, but most authors agree that much of the overlying crust at this time is mafic in composition, such that Lee et al. (2016) used the term mafic continental crust.

Qualitative estimates for increasing thickness of crust through the Archean include the records from large igneous provinces, which change from predominantly submarine to subaerial in the late Archean (Arndt, 1999; Kump and Barley, 2007). Similarly, evidence for the accumulation of sedimentary rocks in terrestrial environments is poor prior to the late Archean (Eriksson et al., 2013).

An alternative approach is to evaluate the thickness of the crust at the time and place of crust generation. Dhuime et al. (2015) provided quantitative estimates on median crustal thickness at its time of generation using the estimated Rb/Sr ratios of new continental crust through time. Rb/Sr correlates with SiO_2 contents of crustal rocks, and in recent continental margins, such as the Andes, with the thickness of continental crust (Fig. 3A). They showed that new crust older than about 3 Ga has a low median Rb/Sr value (~0.03) that increases to a maximum of ~0.08 around 1.8 Ga where it plateaus until around 1.0-0.8 Ga from where it decreases to Recent values around 0.065. Thus, secular changes in Rb/Sr ratio suggest new continental crust was mafic (48 % SiO_2) with a thickness of 20 km or less prior to 3 Ga, before becoming increasingly of intermediate composition (57 % SiO_2) between 3 Ga and 1.8 Ga, by which time it had median thickness of 40 km before decreasing towards 30-35 km for the present day at SiO_2 values of around 55% (Dhuime et al, 2015). Other geochemical data sets also indicate a change in composition of at least the emergent continental crust from mafic to felsic in the period 3.2 Ga to 2.5 Ga (Keller and Schoene, 2012; Tang et al., 2016).

In summary, estimates of the thickness of continental crust incorporate a variety of data sets including present day depth to Moho, the overall low metamorphic grade of surface outcrops of Archean granite greenstone terranes, heat flow, and inferred depths of crustal melting required to produce preserved igneous bodies and crustal lithologic assemblages (e.g., Durrheim and Mooney, 1991; Galer and Mezger, 1998; Moyen and Martin, 2012). Thus, estimates of crustal thickness, especially for the Earth's early crust, commonly reflect the cumulative history of a variety of temporally discrete events (Fig. 2). To understand the evolution of the crust and its lithosphere it is important to segregate this history into its constituent events, and in particular to evaluate the thickness of the crust at the time of its initial generation.

4. Crustal Area

The current area of continental crust is 210.4 x 10^6 km^2 or about 40 % of the Earth's surface area with some 33 % above sea-level (Cogley, 1984). Estimates of how this value has varied through time are limited and largely qualitative. The continental freeboard model of Wise (1974) and the continental growth model of Armstrong (1981) used empirical observations to infer constant thickness, area and volume since at least the late Archean. Arndt (1999) noted that the widespread preservation of flood volcanism on continental platforms required a significant area of continental crust in the Precambrian. Estimates of continental volumes at 3 Ga of around 70 % of the present day volume (Fig. 1; Campbell, 2003; Dhuime et al., 2012; Dhuime et al., 2017; Pujol et al., 2013) along with inferred thickness of continental crust at around half that of today (Dhuime et al., 2015) imply an area of continental crust already similar to the modern Earth (Hawkesworth et al., 2016). We infer that the bimodal hypsometry of oceanic and continental crust at a ratio of 0.6:0.4 became an ingrained component of the Earth system from around ca. 3 Ga in association with the increasing volume of Archean continental crust related to the development and stabilization of the cratons. Undoubtedly, there were components of thick and thin crust prior to this time but the establishment of a sustained system of plate tectonics requires on-going subduction of oceanic lithosphere, which through interaction with the mantle wedge generates continental crust.

The time-integrated average composition of continental crust is calc-alkaline andesite and is widely thought to have been generated in convergent plate margin settings (e.g., Rudnick and Gao, 2003; Taylor, 1967). This link between continental composition and tectonics indicate that the similarities in continental area between the present day and 3 Ga reflect the operation of a plate tectonic regime on the Earth throughout this time period.

If, on an Earth-like planet controlled by plate tectonics the initial crust was entirely oceanic, with a unimodal hypsometry, then all subduction zones would be intra-oceanic. The resultant supra-subduction zone magmatism would produce continental crust of calc-alkaline andesite and hence the proportion of continental crust would increase. Conversely, if all crust was of continental type, then convergent boundaries would be collisional and subduction would cease. These relationships imply a feedback mechanism between continental surface area and convergent plate boundary type; production of calc-alkaline andesite at ocean-ocean boundaries will increase the proportion of continental crust whereas high proportions of continental crust will limit subduction of oceanic lithosphere resulting in a decrease in continental generation. Höning and Spohn (2016) argue that the interaction between the growth of continental areas and water contents in the mantle has an important role in the thermal evolution of Earth. They conclude that the present continental area of 40 % of the Earth’s surface is consistent with a mantle water content of 100 ppm, and most of the destructive plate margins being of ocean-continent type. They note that their models are more sensitive to the balance between water degassing and regassing than to the balance between continental erosion and production at the present day. The different strands of evidence discussed here highlights that the present continental area of 40 % of the Earth’s surface has been a feature for the last 3 Ga.

The concept of a steady-state value for continental area is further supported by modelling the effects of continental insulation above the convecting mantle and the equilibrium in gravitational potential energy of continental lithosphere and mid-ocean ridges. Lenardic et al. (2005) noted that continental growth will reach a critical point where increasing continental area and resultant continental insulation on the mantle causes lithospheric stress levels to approach their yield stress. This condition suggests a limiting condition on continental surface area and makes it difficult to maintain plate margins, which would lower subduction rates limiting the further growth of continents. Their modelling suggests that current continental area values of 0.4 maximizes global mantle heat flux. Sandiford (2010) pointed out the near equivalence of gravitational potential energy of the current area of continental lithosphere with that of mid-ocean ridges implying that on geological time-scales the continents are everywhere critically stressed such that feedbacks sustain the current surface area. Thus, thinner continental crust spread over a greater area of the current Earth’s surface would likely be under a compressional state of stress leading to thickening of the crust. Conversely, thicker crust with a reduced area would be under extension and likely to spread and become thinner. Furthermore, modelling variations in continental area, erosion of continental sediment into ocean basins, and mantle wedge melting in subduction zones can reproduce a continental area at 0.4 of the Earth’s surface area under a plate tectonic regime (Fig. 3; Höning and Spohn, 2016). Thus, the area of continental crust has likely remained around 40% of present day area since at least the initiation of plate tectonics at around 3 Ga.

5. Discussion – Evolving crustal growth and the Earth System

5.1. Model of crustal growth

The spatial and temporal distribution of sedimentary, igneous and metamorphic rock types, along with available but limited paleomagnetic data document the development of rigid lithospheric plates and their independent relative motion, requiring significant changes in Earth behaviour in the mid- to late Archaean, between 3.2 Ga and 2.5 Ga (Cawood et al., 2018). These changes are consistent with a transition from a non-plate tectonic mode to the onset of sustained, globally distributed system of plate tectonics. For simplicity in the following discussion, we use a figure of ca. 3 Ga for the onset of sustainable plate tectonics but recognize that the geological record indicates a range of Archean ages, typically from different areas, for stabilization of cratonic lithosphere and changing tectonic mode.

Estimates of the changes in the thickness of new continental crust and of crustal area outlined above are linked to the operation of plate tectonics. Accepting the geological evidence for the initiation of this tectonic mode to be around 3 Ga enables speculation on an integrated model of continental growth (Fig. 4) and the geodynamic implications for the Earth system. It is inferred that the area of continental crustal increased rapidly after formation of the Earth’s initial mafic crust and magma ocean until ca. 3 Ga when it reached present day values of around 40 % of total surface area, which has been maintained since that time. The thickness of juvenile crust and associated lithosphere on the early Earth was relatively thin, and mainly mafic in composition with a largely unimodal hypsometry. Continental crustal thickness increased through the late Archean associated with the stabilization of cratons, and then through the Paleoproterozoic associated with convergent plate interaction and collisional assembly of the cratons along orogenic belts to form the supercontinent Nuna. New crust generated during the Mesoproterozoic was relatively thick but has decreased since the Neoproterozoic, in part overlapping with the break-up of the Rodinian supercontinent.

5.2. Bimodal hypsometry of crust, and the development of plate tectonics

The bimodal hypsometry of oceanic and continental crust is a striking feature of the present day Earth (Fig. 5), as is plate tectonics as the dominant tectonic mode controlling the Earth system. At issue is the extent to which they are related, the mechanisms involved, and when they may have developed.

Bimodal hypsometry on Earth reflects the contrasting physical/chemical properties of relatively thin, dense, mafic oceanic crust and thick, less dense, felsic continental crust. We envisage a number of possible ways bimodal hypsometry may have developed and its relationship to tectonic mode (Fig. 5). We assume that Earth's crust post-solidification of the magma ocean was, like other rocky planets in the solar system that lack plate tectonics, mafic and unimodal (Fig. 6A). High mantle potential temperatures for the early Earth suggest high degrees of adiabatic decompression melting associated with fertile mantle upwelling to solidus conditions resulting in mafic crust some 20 km thick or more (Sleep and Windley, 1982; van Thienen et al., 2004). The subsequent formation of Archean cratons through the generation of TTG's marked the development of the continental crust and the beginning of deviation away from a unimodal hypsometry.

The magmatic record preserved in Archean cratons indicates a protracted evolution. It consists of early high-level largely mafic magmatism and associated sedimentation, represented by greenstone belts, followed by a protracted history of more felsic TTG magmatism emplaced into the greenstone belt crust. The temporal record of TTG emplacement extends up to hundreds of millions of years, and is followed by a relative short pulse, ranging over only a few million to tens of millions of years, of more potassic granite activity (Laurent et al., 2014, and references therein). The overall duration of cratonic magmatic activity and the timing of its termination vary from craton to craton. The TTGs are derived from melting of mafic crust altered to amphibolite and they record secular trends with a decrease in combined Na+Ca and Sr from early to late Archean time (Martin and Moyen, 2002). High-Sr TTG appear progressively with time, whereas low-Sr TTG (i.e., low pressure melting in the presence of plagioclase) are present throughout the Archean (Moyen and Laurent, 2018; Moyen and Martin, 2012). Thus, the spectrum of TTG compositions, and hence, presumably, the variety of depth of melting and geodynamic environments, progressively widens during the Archean. The increasing importance of high-Sr TTG demands a more prominent role of melting in the presence of larger amounts of garnet (i.e., deeper, higher-pressure melting for a given source composition) towards both the late Archean and the latter stages in the evolution of each craton (Fig. 6B). Overall melting depths for the TTG suite are estimated to range from ca. 5 kbar to > 20 kbar (Moyen and Laurent, 2018). The increase in melting depth with time suggests relatively thick cratonic crust by the late Archean, and in Figure 6 we have taken that to be 50 km. The presence of potassic and peralkaline plutons during the final stages of Archean craton magmatism is consistent with this record of overall thickening and melting of the crust (Nebel et al., 2018). Much of the preserved Archean continental lithosphere is characterized by a thick, low density, mantle component (Griffin et al., 2009), facilitating its long term preservation in the geological archive. Given that Archean age crust makes up less than 15 % of the present day continental volume (Fig. 1), but is inferred to constitute at least some 75 % of continental volumes by 3 Ga (Fig. 3; Dhuime et al., 2012; Pujol et al., 2013), then significant destruction and recycling of material must have occurred and that is presumed to have occurred through plate tectonics

(Chowdhury et al., 2017; Dhuime et al., 2018). This relationship, together with the variable depths of crustal melting of amphibole required to produce the spectrum of Na+Ca and Sr compositional trends within TTG, suggests a range of Archean continental crustal thicknesses and associated hypsometry (Fig. 5).

The relatively thin nature of new continental crust at its time of generation in the Archean (Fig. 3A, blue curve) contrasts with the present day thickness of Archean cratonic crust (Fig. 3A, red line). This present day thickness was likely achieved by the end of the Archean, by thickening, melting and felsic magmatism, based on the near coincidence of sea level and erosion base level for cratons with present day sea level (Armstrong, 1981). This relationship implies at least a doubling in the thickness of Archean crust since its initial generation and is consistent with geological observations of the cratons requiring thick, rigid crust by the end Archean (Cawood et al., 2018). Increasing crustal thickness is also consistent with models of TTG genesis, which require melting of mafic amphibolitic crust at up to 20 kbars or more (Fig. 6B)(Moyen and Laurent, 2018; Nebel et al., 2018).

From the end of the Archean, the production of continental crust within an established plate tectonic regime would have developed on, and alongside, this basement of older Archean crust of variable thickness. Subduction would likely have preferentially recycled the thinner Archean crust that did not form a basement for arc magmatism (Chowdhury et al., 2017; Dhuime et al., 2018), thus leading to the development of bimodal hypsometry through the late Archean and early Proterozoic (Fig. 5).

5.3. Implications of evolving crustal growth to the Earth system

The tectonic mode, and hence the processes of crust generation, prior to the initiation of sustained plate tectonics remain difficult to establish, and they are dependent in large part on the nature of the thermal conditions of the Archean mantle. The high mantle potential temperature of the early Earth (Herzberg et al., 2010) resulted in thin, rheologically weak lithosphere, hindering the propagation and focusing of stress, and hence the development of plate boundaries. Most models of the pre-plate tectonic mode highlight the development of lithospheric instabilities resulting in an early regime involving vertical asthenospheric up-welling, melting and mafic crustal generation counter-balanced by lithospheric down-welling and recycling (Bédard, 2018; Fischer and Gerya, 2016; Johnson et al., 2014). These processes, together with mantle convection induced lithospheric imbrication (Beall et al., 2018; Nebel et al., 2018), resulted in zones of crustal thickening and melting leading to the widespread emplacement of TTG, as discussed above.

Integration of area and thickness curves implies an increase in continental volume from ca. 4.5 to 1.8 Ga, relatively constant volume during Earth's middle age and then a decreasing volume from the Neoproterozoic to the present day (Fig. 4). The increase in crustal volume from the Hadean to the Paleoproterozoic was driven initially by an increase in continental area associated with a non-plate tectonic mode, and since approximately 3 Ga and the initiation of plate tectonics in response to the development of stronger lithosphere. Given the calc-alkaline andesite bulk composition of the crust the increase in thickness was likely due to convergent plate margin magmatism. Inferred constant continental volume in the period 1.8-1.0 Ga corresponds with stability in crustal thickness and area but thickness, and presumably volume, have decreased since then (Fig. 4).

A constant continental area for the last 3 Ga has implications for interactions between crust and mantle with respect to heat flux and convection (cf., Lenardic et al., 2005) as well as between the crust and the surficial components of the Earth system including hydrosphere, atmosphere and biosphere. Proxies of seawater and source rock composition through time, including Sr, Zn and Ti isotopes, along with evidence from Hf and O isotopes in zircon, imply emergence, erosion and reworking of at least an upper continental crust of intermediate to felsic composition since 3.0-2.5 Ga (Bindeman et al., 2018; Dhuime et al., 2017, and references therein; Greber et al., 2017; Satkoski et al., 2016; Smit and Mezger, 2017; Viehmann et al., 2014). Evolution of the Sr seawater curve from mantle values in the late Archean to evolved values similar to those measured today, and the shift in the triple-oxygen isotope composition of shales to modern values around 2.5 Ga, is consistent with emergence of continental crust due to its increased thickness and felsic composition (Cawood et al., 2013; Dhuime et al., 2017) and an exposed area comparable to that of the present day (Bindeman et al., 2018). Similarly, the evolving thickness of constant area crust for the last 3 Ga, has implications for biological productivity through input into the oceans from continental exposure and weathering of bio-essential elements such as P (Cox et al., 2018; Reinhard et al., 2017).

The thickness of new crust appears to have continued to increase throughout the Paleoproterozoic reaching a maximum value at ca. 1.8 Ga, and maintaining a constant thick value until the early Neoproterozoic around 1.0-0.8 Ga (Fig. 4). This time period (1.8-0.8 Ga) is characterized by preponderance of certain rock types (e.g., massif anorthosites) but a paucity of other rock types and mineral deposit types (e.g., banded iron formations, glacial deposits, orogenic gold and volcanic hosted massive sulfides) and is referred to as the “boring billion” or Earth’s middle age (Cawood and Hawkesworth, 2014; Holland, 2006; Roberts, 2013). The presence or absence of rock associations that characterize this period relative to their distribution in preceding and succeeding periods may link to processes recorded in generating the observed new continental crustal thickness curve and/or continental paleogeography. This time period is bookended by assembly of the major components of the Nuna supercontinent at ca. 1.8 Ga and the commencement of the breakup of the succeeding supercontinent of Rodinia commencing around 0.8 Ga. The transition from Nuna to Rodinia was not marked by a major phase of supercontinent breakup, as evidenced by the lack of preserved passive margins in this time period (Bradley, 2008), but rather by the addition of continental fragments (e.g., Australia-Antarctica and Amazonia) onto a stable cratonic core of Laurentia, Baltica and Siberia (see also Cawood et al., 2016). A long-lived convergent plate margin existed along the Laurentian and Baltic margin from the late Paleoproterozoic and throughout the Mesoproterozoic (ca. 1.8-1.0 Ga; Åhäll et al., 2000; Bingen et al., 2008; Brewer et al., 2004; Cawood and Pisarevsky, 2017; Dickin et al., 2015; Dickin et al., 2010; Slagstad et al., 2017).

This long ranging and stable continental configuration may have provided a primary constraint controlling the lithological associations and environmental conditions that characterize this period. Massif anorthosites and related igneous rocks are focussed, particularly in terms of overall area of anorthosite, in the period 1.8-0.8 Ga, but with minor examples extend the overall range from ca. 2.6-0.5 Ga (Ashwal and Bybee, 2017). They are spatially and temporally linked to overall convergent continental plate margins, occurring in syn-subduction environments but well inboard of the inferred plate margin, or in post-subduction, collisional orogenic settings (Ashwal and Bybee, 2017; McLelland et al., 2010; Whitmeyer and Karlstrom, 2007). Massif anorthosite formation is related to the ponding of magma of basaltic composition at depths of 30-40 km, at the base of the crust, where plagioclase rich mushes accumulate through flotation separation, and which then rise over protracted periods through the crust, were variably contaminated, prior to emplacement at mid-crustal levels (Ashwal and Bybee, 2017, and references therein). Thus, the relatively thick new continental crust during Earth’s middle age (Fig. 3A, blue curve) would have facilitated massif anorthosite formation. Ashwal and Bybee (2017) note that massif anorthosites pre-dating middle age and extending back to 2.6 Ga may relate to secular cooling, which resulted in increased lithospheric strength and crustal thickness (Fig. 3A, red line) suitable for the ponding and slow crystallization of basaltic magmas at the base of the thick stabilized Archean cratonic crust.

The absence of orogenic gold and VHMS from the interval ca. 1.8-0.8 Ga is more enigmatic. These deposit types are associated with convergent plate margin settings (Bierlein et al., 2009; Groves and Bierlein, 2007) and hence Earth's middle age should have been a fertile period for their genesis. Goldfarb et al. (2001) suggest the paucity of orogenic gold in this interval may reflect the lack of preservation of shallow to mid-crustal sections, in which such deposits tend to occur. However, these is no direct evidence that orogenic belts from this time interval display deeper crustal levels than preceding or succeeding intervals, although Brown and Johnson (2018) note that the thermal gradients associated with high d*T*/d*P* metamorphism rose to a maximum at this time, which they suggested reflected insulation of the mantle beneath the lithosphere of Nuna. The absence of VHMS deposits during middle age has been related to limited slab rollback and hence back-arc extension basins (Huston et al., 2010), which constitute prime sites for this type of mineralization along the margins of Rodinia. The thick nature of new continental crust generated is perhaps consistent with this proposal, but the widespread development of broad accretionary orogens along the eastern margin of Laurentia and Baltica (Karlstrom et al., 2001; Whitmeyer and Karlstrom, 2007) require an ocean plate and trench that were retreating from the supercontinent and thus likely placing the Rodinian upper plate in extension. Part of the problem in explaining the absence of certain deposit types from Earth's middle age is the assumption that the VHMS and orogenic gold deposits in older and younger time periods formed by the same process. For example, orogenic gold deposits are characterized by auriferous-bearing quartz veins formed during brittle fracturing of host rocks associated with regional deformation and metamorphism. However, the nature of the host rocks and their tectonic setting vary with Archean examples associated with greenstone belts, which may have formed in hot spot-related environments, whereas the younger deposits are associated with metasedimentary packages along convergent plate margins. Similarly, there are differences in the nature of subduction between the two time periods. Subducton of Archean lithosphere occurred into a hotter mantle than the Phanerozoic, resulting in shallow slab breakoff and likely episodic subduction (Cawood et al., 2018; Hawkesworth et al., 2016; Moyen and van Hunen, 2012). In contrast, cooler mantle temperatures during Phanerozoic subduction enabled deep sustained subduction of oceanic lithosphere (Sizova et al., 2014). Differences in the oxygenation of the oceans between the two periods is also significant in controlling the solubility of gold in the deep oceans (Evans and Tomkins, 2011; Tomkins, 2013). Similarly, there are significant differences between Archean and Phanerozoic VHMS deposits, with the more oxygenated environment of the latter as well as the more evolved composition of the crust in contrast to the more saline nature of the early Earth oceans (Huston et al., 2010).

The decrease in the thickness of juvenile crust since the Neoproterozoic corresponds with the disappearance for massif anorthosites from the rock record, implying a link related to changing process. This decrease in thickness would imply a corresponding decrease in crustal volume assuming no change in crustal area. A volume decrease is consistent with estimates of modern day crustal generation and recycling (Fig. 3D), in which overall global rates of crustal loss tend to be greater than magmatic addition by at least 1 km^3/a (Clift et al., 2009; Scholl and von Huene, 2009; Stern, 2011). Extrapolation of this rate of continental crust removal over 0.8 Ga results in a ca. 15 % reduction of continental volume. The Neoproterozoic marks the appearance in the rock record of suites of rocks considered the hallmarks of modern cold subduction such as the appearance of high- to ultrahigh-pressure metamorphic rocks (Brown and Johnson, 2018, and references therein), along with ophiolites and other rock types associated with modern cold subduction zones (Stern, 2005). The appearance of coesite and diamond-bearing ultra-high pressure (UHP) assemblages in the

Neoproterozoic and Phanerozoic are modelled within a cooler mantle environment (less than 100 ^{o}C higher than present day mantle potential temperatures), which enables a deeper level of detachment of the oceanic plate lithosphere from the lower continental lithosphere during continent-continent collision (Sizova et al., 2014). This facilitates both underplating of the cold lower plate continental crust down to the stability field for UHP metamorphism and its subsequent rapid exhumation. In addition to the incoming of UHP assemblages into the rock record, this increased burial and associated reworking of continental material in collision zones since the Neoproterozoic is recorded in pronounced evolved Hf and O zircon values associated with Gondwana assembly (Dhuime et al., 2012; Gardiner et al., 2016; Spencer et al., 2014). The corresponding increase in seawater Sr values (Shields and Veizer, 2002; Shields, 2007) likely reflects extensive exhumation and erosion of the continental crust, and the consequent increase in contributions to the oceans from riverine runoff. Thus, secular cooling of the mantle enabled more efficient recycling of the continental crust accounting for the inferred decrease in volume since the Neoproterozoic.

6. Conclusions

The Earth is a dynamic, evolving system in which the surficial and solid components of the planet interact through a series of cycles and at a variety of scales in response to energy supplied from internal heat and from external sources within the solar system. The grand challenge for the Earth Sciences is to unravel the feedbacks between the deep and surficial Earth, the record of which is preserved in the continental crust. A first order pre-condition is understanding the spatial and temporal evolution of the volume, thickness and area of the continental crust. We realize that all approaches to determine these amounts involve varying degrees of uncertainty. Unlike previous studies that have often focussed on estimates of continental volume (Fig. 1), we have attempted to document temporal variations in area and thickness, to derive variations in volume, and argue that these are related to the tectonic mode in which the Earth operates. But we also note that calculations of continental volume that are independent of the preserved present day volume display a similar overall form (Fig. 1, blue and green curves) and yield values for the continental crust at 3 Ga of around 70 % of present day volume. Combined with estimates for the thickness of new formed continental crust determined from secular variations in Rb/Sr contents of igneous rocks at this time, this suggests a continental area similar to today. We consider that a global plate tectonic regime has operated since ca. 3 Ga and this mode of crust formation controls the thickness, composition, and area of continental crust. Prior to ca. 3 Ga, crustal growth was associated with production of mafic crust, which underwent thickening, leading to burial and melting of amphibolite at a range of depths, which in turn lead to a range of TTG compositions with a resulting hypsometry showing a continuum of values. By 3 Ga the crust was sufficiently rigid to enable a sustained plate tectonic mode, which continues to the present day. Once established, the interplay of the convergent plate margin types, intra-oceanic, continental margin and continental collision, may have reached an equilibrium that results in a crustal area of 60 % oceanic and 40% continental that has remained relatively constant throughout this period along with the associated bimodal hypsometry. Generation of new continental crust at convergent plate margins was initially associated with increasing the thickness of crust reaching a steady-state thickness from ca. 1.8-0.8 Ga with an overall hypsometry similar to the present day (Fig. 5). This thick crust facilitated the production of massif anorthosites, which characterize this interval. They occur in locations inboard of stable long-lived convergent margins, such as that along eastern Laurentia and southern Baltica. This stability may have facilitated the long-term ponding of basaltic magma at the base of the crust, flotation of plagioclase within the chamber, and its rise to the mid-crust.

The absence of certain mineral deposit types from this time period (1.8-0.8 Ga), such as orogenic gold and VHMS, is difficult to explain but corresponds to changing lithospheric and environmental conditions including nature of host rocks, and hence tectonic mode of their generation, and oxygenation, salinity and sulphate content of the oceans, which affects mineral solubility. Since 0.8 Ga crustal thickness has decreased. This decrease corresponds with the beginning of the widespread preservation of high P/low T metamorphic assemblages. This may reflect increased continental subduction and recycling of the crust.

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Figure Captions

Figure 1. Crustal growth models of Artemieva (2006), Allègre and Rousseau (1984), Condie and Aster (2010), Belousova et al. (2010), Dhuime et al. (2012), Roberts and Spencer (2015), Campbell (2003), Pujoi et al. (2013), Armstrong (1981) and Fyfe (1978) compared to the age distribution of presently preserved crust from Goodwin (1996).

Figure 2. Schematic representation of changing thickness of crust from its time of generation through extraction of melt from the mantle, to its stabilization and incorporation into the long term rock archive, as well as late stage reworking of the crust.

Figure 3. A. Rb/Sr ratios of juvenile continental crust plotted against the age of crust formation for ~13,000 whole-rock analyses from Dhuime et al. (2015). Rb/Sr increases with both whole-rock silica content and crustal thickness at the site of magma generation. Solid red lines are average present day crustal thicknesses for specified time periods from Mooney et al (1998). B. Relative length of ocean-ocean, ocean-continent and continent convergent plate boundary types scaled to 1 (Höning and Spohn, 2016). C. Range (shaded area) of modelled evolution curves for continental area (see Höning and Spohn, 2016, for details of model parameters). D. Schematic cross section of convergent, collisional, and extensional plate boundaries showing estimated amounts (in km^3/a) of continental addition (numbers in blue above Earth surface) and removal (numbers in red below surface). Data are from Scholl and von Huene (2009), Clift et al., (2009) and Stern (2011).

Figure 4. Schematic curves for the evolution of continental crustal area and thickness adapted from Höning and Spohn (2016) and Dhuime et al. (2015), respectively.

Figure 5. Earth's present day hypsometry along with schematic curves for the early, mid- and late Archean, and the late Paleoproterozoic to early Neoproterozoic (middle age). Horizontal axis represents proportional area. Red dotted lines show mean maximum continental and ocean elevations, which increase and decrease respectively, with time.

Figure 6. Schematic cross sections of: A. Post-magma ocean mafic crust. B. Mid- to Late Archean cratonic crust.

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Cawood & Hawkesworth biography's

Peter Cawood

Peter Cawood's research has focused on the origin of the Earth's continental lithosphere (crust and upper mantle) and the processes of its generation, stabilization and reworking. He integrates direct field observations with leading laboratory techniques, and has worked in regions from Archean cratons to modern and active margins, and at scales ranging from global to microscopic. His work aims to resolve the range of tectonic processes involved in lithosphere formation and the feedbacks with the rest of the Earth system. His major research contributions include: demonstrating that the archive of Earth history is not simply a record of the processes of crustal generation but markedly biased by the supercontinent cycle; innovative studies on the early stages of collisional orogenesis that link ophiolite generation to its emplacement; a model for the deformation and stabilization of accretionary orogens, and temporal relations to collisional orogenesis; the role and timing for initiation of plate tectonics on the early Earth; and the application of microanalytical techniques to unravel the provenance history and palaeogeography of sedimentary basins and orogenic belts. Peter obtained undergraduate and PhD degrees from the University of Sydney and has held academic positions in Australia, New Zealand, Canada, and the UK. He is currently ARC Laureate Fellow at Monash University and is an elected Fellow of the Australian Academy of Sciences and the Royal Society of Edinburgh.

Chris Hawkesworth Bio

Chris Hawkesworth applies isotopic systems to unraveling how and when the continental crust was generated and the links between geochemistry and tectonics. He has contributed to our understanding of the generation of subduction-related magmas, continental flood basalts, timescales through U-series isotopes, the development of base metal deposits, and environmental changes throughout time. He has used zircons as an archive of the evolution of the continental crust, and investigated how its composition has changed from the Hadean to the present. He is interested in constraining rates of natural processes, and placing regional studies in a global context. Chris is a graduate of Trinity College Dublin, and obtained his D.Phil at Oxford working with Ron Oxburgh in the Tauern Window of the Eastern Alps. He set up an isotope research group at the Open University before moving to Bristol in 2000. He was Deputy Principal at the University of St Andrews from 2009 to 2014, and he has served as President of the EAG and as a Director of the Geochemical Society. Chris is an elected Fellow of a number of societies including the Royal Society and the Royal Society of Edinburgh, and holds Emeritus positions at the Universities of Bristol and St Andrews.

Highlights

Continental area and thickness varied independently and increased at different rates and over different periods, in response to different tectonic processes, through Earth history.

Crustal area increased steadily on a pre-plate tectonic Earth and by 3 Ga had reached a dynamic equilibrium of around 40 % of the Earth's surface.

New continental crust was thin and mafic from ca. 4-3 Ga and started to increase with the inferred onset of plate tectonics at ca. 3 Ga.

Earth's bimodal hypsometry developed at this time.

Continental volume increased from 4.5 Ga to 1.8 Ga, and remained relatively constant through Earth's middle age (1.8-0.8 Ga), and has decreased since the Neoproterozoic.

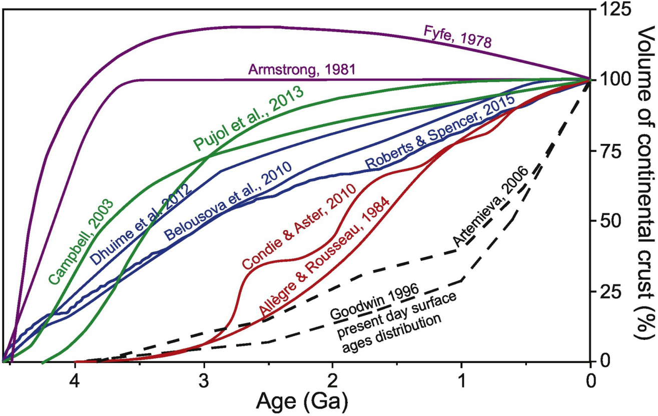
Fyfe, 1978
Armstrong, 1981
Pujol et al., 2013
Roberts & Spencer, 2015
Belousova et al., 2010
Dhuime et al. 2012
Campbell, 2003
Condie & Aster, 2010
Allègre & Rousseau, 1984
Artemieva, 2006
Goodwin 1996
present day surface
ages distribution
4
3
2
1
0
Age (Ga)
125
100
75
50
25
0
Volume of continental crust (%)

Figure 1

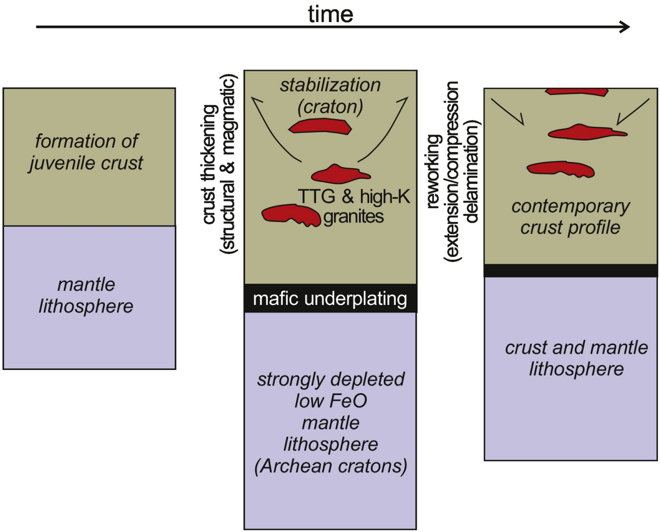
time
formation of juvenile crust
mantle lithosphere
crust thickening (structural & magmatic)
stabilization (craton)
TTG & high-K granites
mafic underplating
strongly depleted low FeO mantle lithosphere (Archean cratons)
reworking (extension/compression delamination)
contemporary crust profile
crust and mantle lithosphere

Figure 2

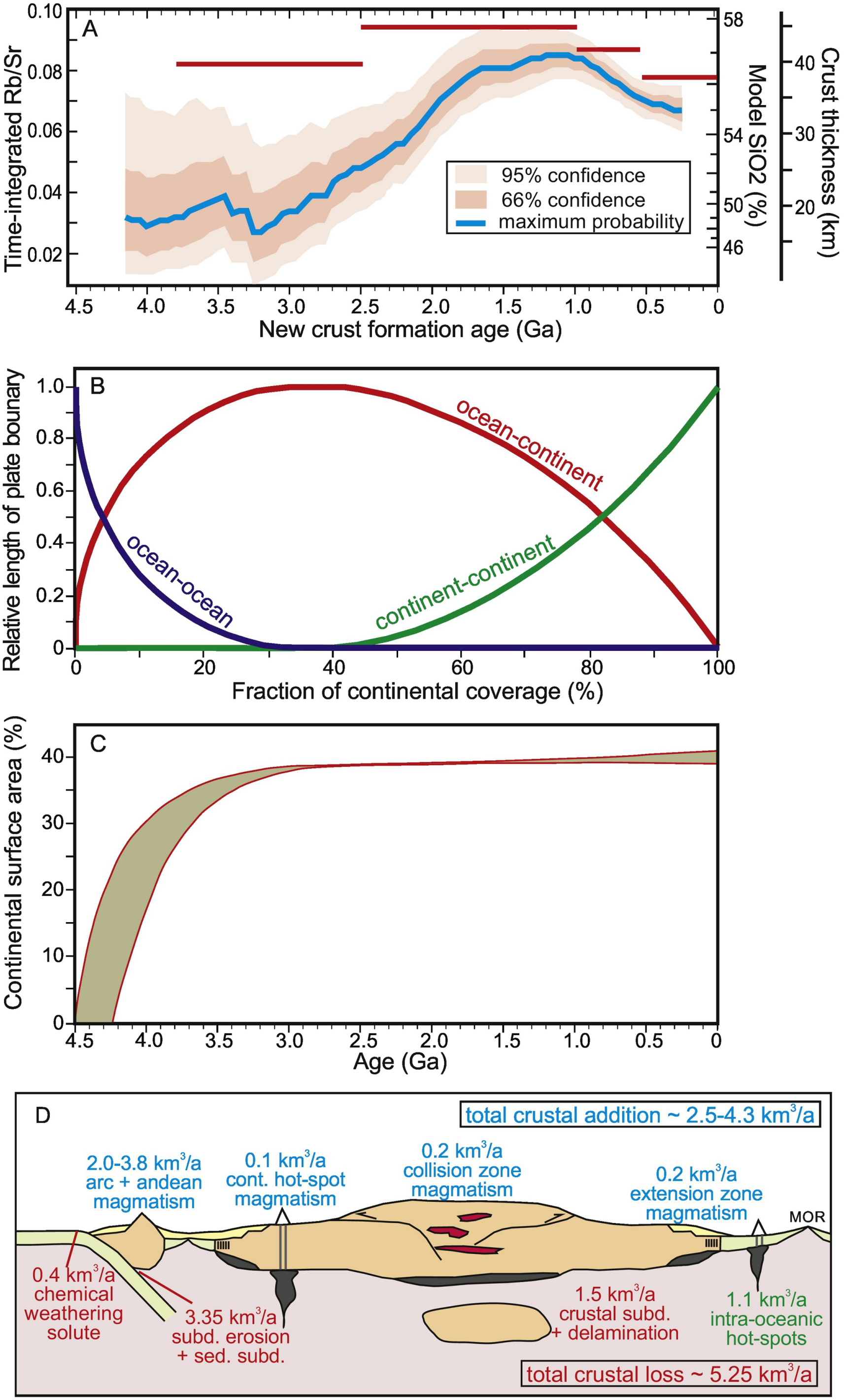
A
Time-integrated Rb/Sr
0.10
0.08
0.06
0.04
0.02
Model SiO2 (%)
58
54
50
46
Crust thickness (km)
40
30
20
95% confidence
66% confidence
maximum probability
4.5
4.0
3.5
3.0
2.5
2.0
1.5
1.0
0.5
0
New crust formation age (Ga)
B
Relative length of plate bounary
1.0
0.8
0.6
0.4
0.2
0
ocean-continent
ocean-ocean
continent-continent
0
20
40
60
80
100
Fraction of continental coverage (%)
C
Continental surface area (%)
40
30
20
10
0
4.5
4.0
3.5
3.0
2.5
2.0
1.5
1.0
0.5
0
Age (Ga)
D
total crustal addition ~ 2.5-4.3 km^3/a
2.0-3.8 km^3/a arc + andean magmatism
0.1 km^3/a cont. hot-spot magmatism
0.2 km^3/a collision zone magmatism
0.2 km^3/a extension zone magmatism
MOR
0.4 km^3/a chemical weathering solute
3.35 km^3/a subd. erosion + sed. subd.
1.5 km^3/a crustal subd. + delamination
1.1 km^3/a intra-oceanic hot-spots
total crustal loss ~ 5.25 km^3/a

Figure 3

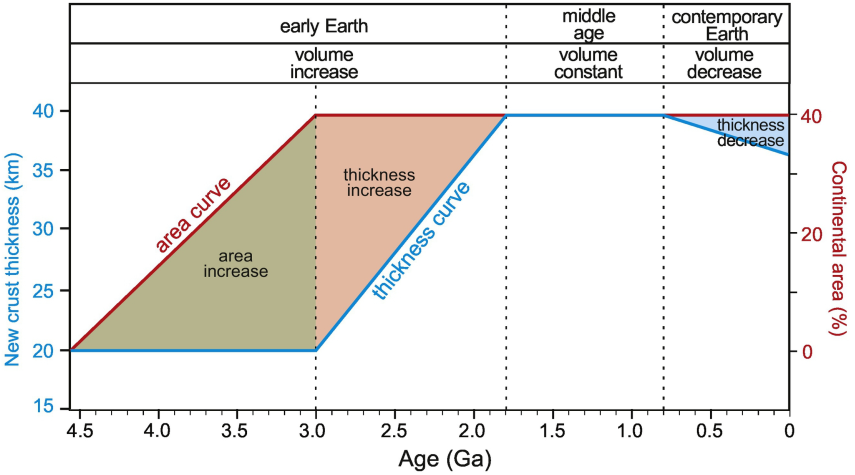
early Earth
middle age
contemporary Earth
volume increase
volume constant
volume decrease
area curve
area increase
thickness increase
thickness curve
thickness decrease
New crust thickness (km)
15
20
25
30
35
40
Continental area (%)
0
20
40
4.5
4.0
3.5
3.0
2.5
2.0
1.5
1.0
0.5
0
Age (Ga)

Figure 4

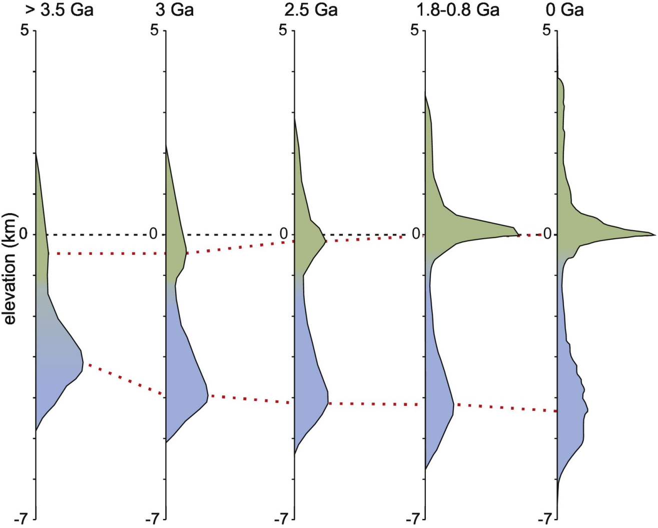

Figure 5

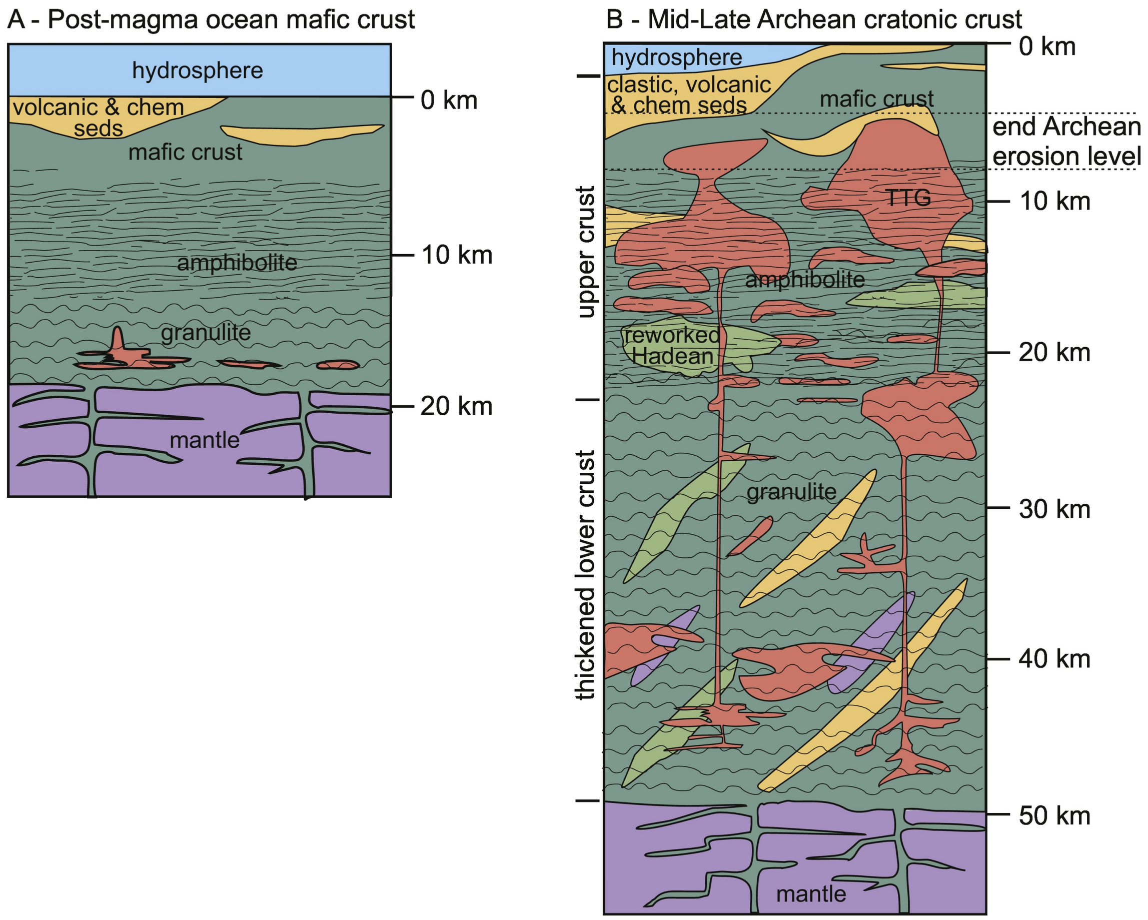
A - Post-magma ocean mafic crust
hydrosphere
volcanic & chem seds
mafic crust
amphibolite
granulite
mantle
0 km
10 km
20 km
B - Mid-Late Archean cratonic crust
hydrosphere
clastic, volcanic & chem seds
mafic crust
end Archean erosion level
TTG
amphibolite
reworked Hadean
granulite
mantle
upper crust
thickened lower crust
0 km
10 km
20 km
30 km
40 km
50 km

Figure 6