Geoscience Frontiers 8 (2017) 309–327

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Geoscience Frontiers

journal homepage: www.elsevier.com/locate/gsf

Research paper

Hadean Earth and primordial continents: The cradle of prebiotic life

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ARTICLE INFO

Article history:
Received 28 February 2016
Received in revised form
15 July 2016
Accepted 17 July 2016
Available online 12 August 2016

Keywords:
Earth-Moon system
Primordial continents
ABEL bombardment
Tectonic erosion
Hadean plate tectonics

ABSTRACT

The Hadean history of Earth is shrouded in mystery and it is considered that the planet was born dry with no water or atmosphere. The Earth-Moon system had many features in common during the birth stage. Solidification of the dry magma ocean at 4.53 Ga generated primordial continents with komatiite. We speculate that the upper crust was composed of fractionated gabbros and the middle felsic crust by anorthosite at ca. 21 km depth boundary, underlain by meta-anorthosite (grossular + kyanite + quartz) down to 50–60 km in depth. The thickness of the mafic KREEP basalt in the lower crust, separating it from the underlying upper mantle is not well-constrained and might have been up to ca. 100–200 km depending on the degree of fractionation and gravitational stability versus surrounding mantle density. The primordial continents must have been composed of the final residue of dry magma ocean and enriched in several critical elements including Ca, Mg, Fe, Mn, P, K, and Cl which were exposed on the surface of the dry Earth. Around 190 million years after the solidification of the magma ocean, "ABEL bombardment" delivered volatiles including H_2O, CO_2, N_2 as well as silicate components through the addition of icy asteroids. This event continued for 200 Myr with subordinate bombardments until 3.9 Ga, preparing the Earth for the prebiotic chemical evolution and as the cradle of first life. Due to vigorous convection arising from high mantle potential temperatures, the primordial continents disintegrated and were dragged down to the deep mantle, marking the onset of Hadean plate tectonics.

1. Introduction

The existence and composition of primordial crust on Earth after the consolidation of magma ocean is a topic of prime importance for not only understanding the early evolution of our planet, but also in gaining insights into the origin of early life. One of the remarkable achievements of the Apollo missions was the knowledge that we could gain on lunar geology, including the presence of primordial continents with variable thickness ranging from 30–75 km predominantly composed of anorthositic crust (Wood et al., 1970; Shearer and Papike, 2005, 2006). Since Moon and Earth shared similar history during the Hadean, it is possible that Earth also had an anorthositic crust, although there are no preserved remnants. The Archean and Proterozoic anorthosites now exposed on Earth are all thin (less than 1 km in thickness), and petrogenetic studies indicate that most of them formed under island arc in supra-subduction zone settings or in post-collisional environment fractionated from underplated basalts (e.g., Windley, 1984; Ram Mohan et al., 2013; Teng and Santosh, 2015), and not through consolidation of magma ocean. Moreover, there are no Hadean rocks on the Earth except for minor detrital zircons as reported from the Jack Hills in Western Australia (Wilde et al., 2001; Kemp et al., 2010). It is clear that there no rocks remained on the modern Earth as old as 4.53 Ga, the age of solidification of magma ocean on the Moon (Borg et al., 2015).

Earth is the only planet in Solar System characterized by plate tectonics involving the birth of oceanic lithosphere at the mid-ocean ridge, its horizontal flight to the trench, and subduction into the deep mantle, controlling the assembly, growth and disruption of continents and supercontinents, the opening and closing of ocean basins, and the impact of all these processes on

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Peer-review under responsibility of China University of Geosciences (Beijing).

http://dx.doi.org/10.1016/j.gsf.2016.07.005
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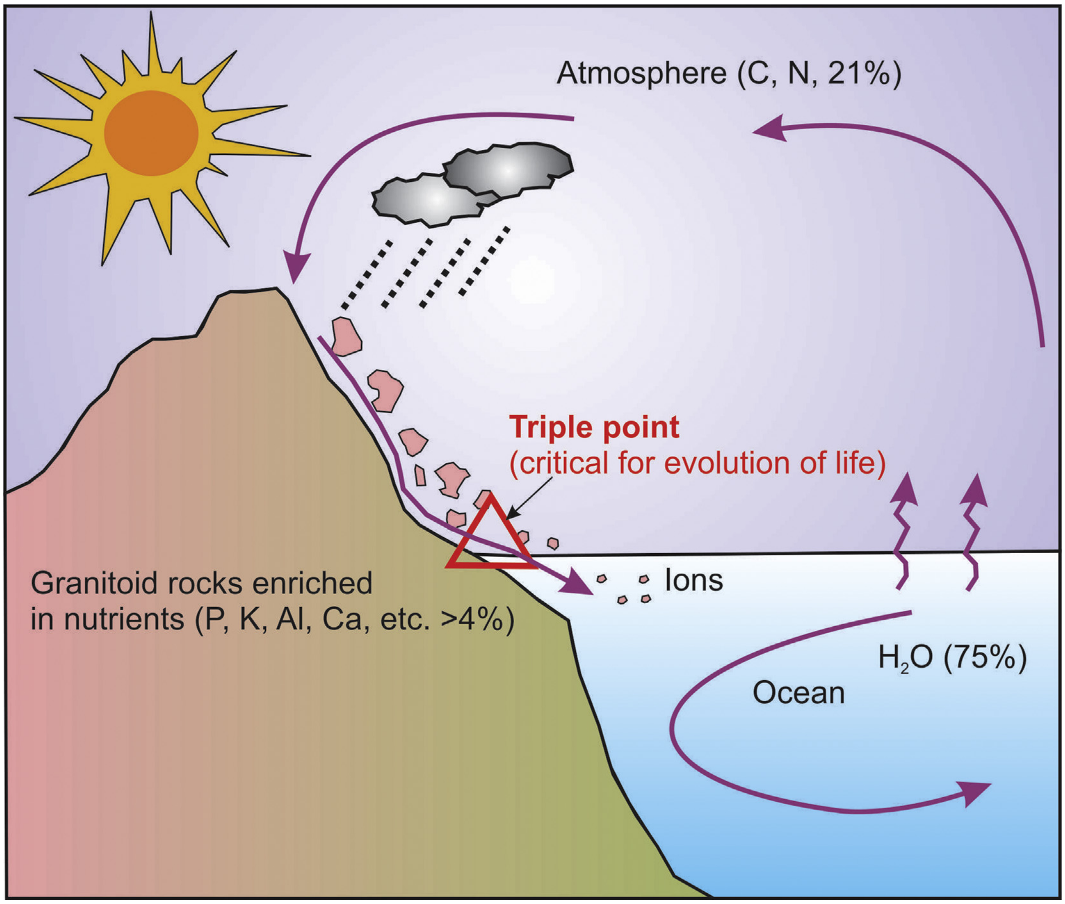


Figure 1. Habitable trinity condition of ocean, landmass composed of primordial continents, and atmosphere on the Hadean Earth envisaged as essential for the origin of early life (after Dohm and Maruyama, 2015).

mineral systems and surface environment (e.g., Maruyama et al., 2007; Santosh et al., 2009; Nance et al., 2014; Pirajno, 2015). However, the timing of initiation of plate tectonics since the birth of Earth at ca. 4.56 Ga ago is highly disputed (Condie and Kröner, 2008). Some workers consider that modern style plate tectonics started only late in the Earth history by early Neoproterozoic (e.g., Stern, 2005), or even later at the onset of Phanerozoic, whereas others propose that plate tectonic processes were initiated soon after the birth of the Earth in Hadean times, immediately after the formation of ocean (e.g., Zahnle et al., 2007; Sleep, 2010). Between these two end-member models are several studies that identify plate tectonic processes broadly analogous to those of modern time in Archean and Paleoproterozoic terranes (e.g., Kranendonk et al., 2007; Smithies et al., 2007; Zhai and Santosh, 2011; Santosh et al., 2015a,b; Yang et al., 2016; among others).

Investigations on lunar samples procured through the Apollo missions indicate the presence of primordial continents on the surface of the Moon composed mainly of anorthosite which has been classified into pure, mafic and iron rich varieties (designated as PAN, MAN and FAN), among which the pure anorthosite dominates (Ohtake et al., 2009; Borg et al., 2015).

If anorthosite was absent on the Hadean Earth, then what were the possible rock types present is an important question. Speculations included water on the early Earth, but not on the Moon, as the giant impact removed all the volatiles on the Moon. It has also been believed that komatiite and related rocks covered the Hadean Earth when the magma ocean solidified (e.g., Synder et al., 1992; Elkins-Tanton, 2008). It also important to evaluate why primordial continents are necessary on the Hadean Earth because this is the fundamental question to understand the Hadean surface environment in relation to the source of nutrients for the birth and evolution of life. Recent models suggest that a 'habitable trinity' condition of ocean, landmass composed of primordial continents, and atmosphere are required on the Hadean Earth (Fig. 1) in order for the early Earth to bear primitive life supported by a balanced supply of nutrients (Dohm and Maruyama, 2015).

In this paper, we present a petrologic model of primordial continents on the Earth to evaluate the most probable lithological stratigraphy on the Hadean Earth. We also propose that after the advent of bio-elements during 4.37–4.20 Ga, the cradle of first life was formed on our planet.

2. Primordial continents: existing views

Among the near and rear (far) sides of the Moon in relation to Earth, thick continental crust of 30–50 km occurs on the near side facing the Earth. The rear side is also characterized by unusually thick (50 to 70 km) anorthositic continental crust (Ohtake et al., 2009). The near side of the Moon also shows extensive lunar basalts formed after the consolidation of magma ocean (Fig. 2). Selective bombardment has generated craters, some of which have diameter up to 400 km diameter on the nearside of Moon (Floran and Dence, 1976) although such features are nearly absent on the rear side suggesting the role of Earth as a 'gravitational lens' to deflect the path of the bombarders resulting in the cratering of

Moon. A comparison of the relative size of Moon, Mars, Venus and Earth, together with schematic diagrams showing the compositional cross sections of Moon and Mars are show in Fig. 3.

Borg et al. (2015) reviewed geochronological data from various isotopic methods on lunar highland samples and noted that the oldest reliable ages are no older than 4.40 Ga. The KREEP basalts are an important component of the lunar crust (Warren and Wasson, 1979) and their age is estimated as 4389–4353 Ma based on mare basalts. The peak Pb-Pb ages of lunar zircons at ca. 4340 Ma is closely comparable with the 4374 Ma of oldest terrestrial zircon age. Borg et al. (2015) suggested that the common range of 4.34 to 4.37 Ga ages in the lunar samples might reflect primordial solidification of a lunar magma ocean or a widespread secondary magmatic event on the lunar nearside.

Hopkins and Mojzsis (2015) reported new zircon U-Pb and trace element investigations from Apollo 14 lunar impact breccia that show ages of 4334–3953 Ma. Ti in zircon thermometry constraints the temperature range as 800–1200 °C. Based on a compilation of age data from Apollo 14 samples, the authors argued that zircons older than 4300 Ma formed by igneous processes associated with lunar crust formation. Ages from the youngest population of ca. 3950 have been assigned to the 'Late Heavy Bombardment (LHB)' although several other studies consider the LHB to have occurred much earlier (e.g., 4.5 Ga; Albarede, 2009).

Petrologically, the presence of unusually thick anorthositic crust indicates density crossover to generate the large amounts of plagioclase. Thus, the total melts must be equivalent to the entire size of the Moon. The question arises as to why the Moon was totally melted? Planetary scientists alert that Moon could not have been entirely melted because its size is too small and therefore the collision and accretion energy are insufficient to cause total melting. This is major reason behind the Giant Impact theory for the origin of Earth-Moon system (e.g., Canup and Asphaug, 2001).

Most scientists agree that the Earth was formed originally by asteroid aggregation with composition of CI chondrite that is enriched in volatile components. However, Apollo mission to the Moon revealed a dry planet without any hydrous silicates on the lunar crust. To explain this, petrologists speculated that the dry Moon was formed through the Giant Impact which removed all the volatile components into space.

However, isotope data suggest that the Earth-Moon system must have been derived from enstatite chondrite which is highly reduced meteorite with no volatiles (e.g., Javoy, 1995). The source meteorite of the Earth was addressed in earlier studies by

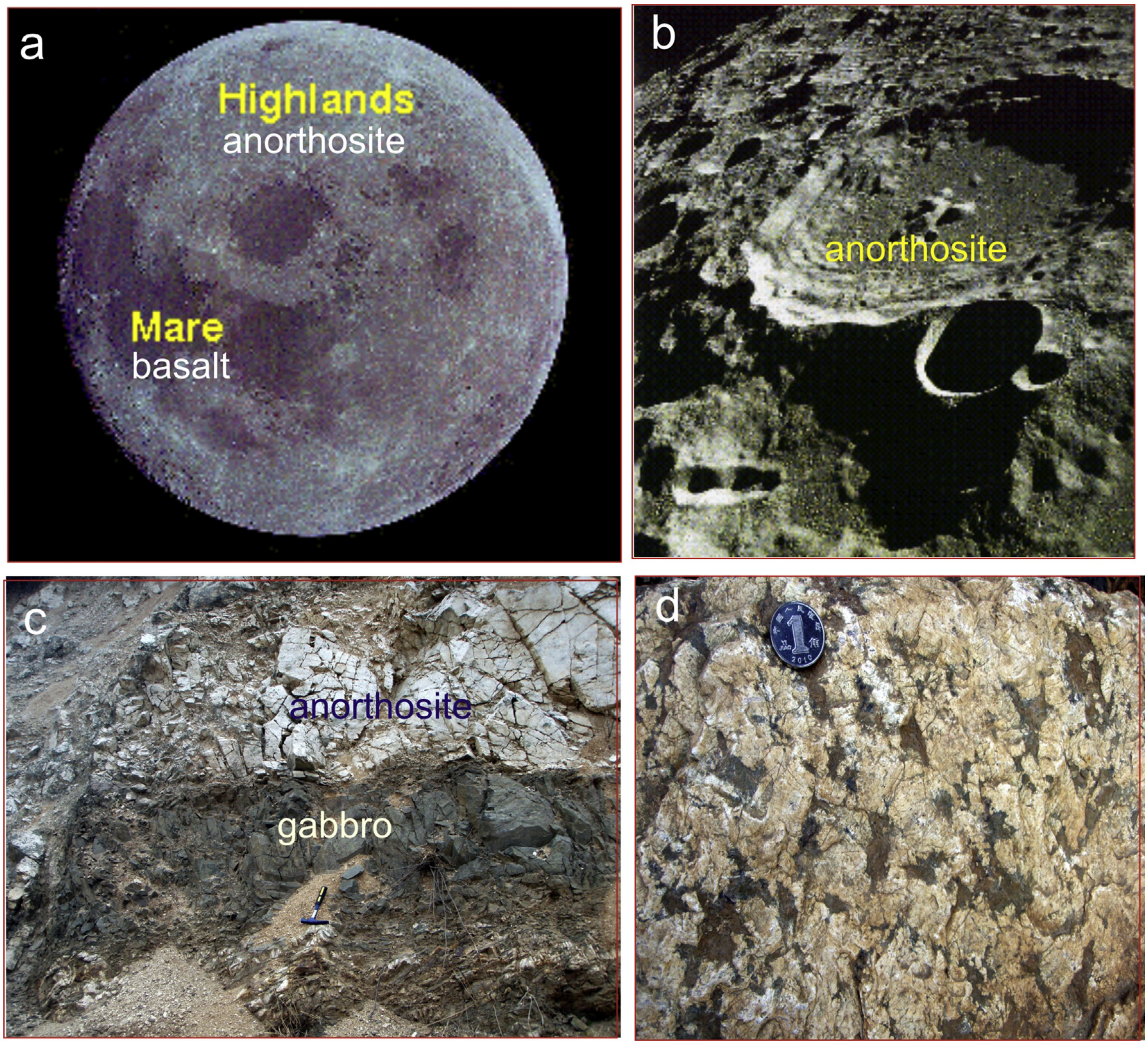


Figure 2. Photos of the lunar nearside (a) and far side (b) (from http://csep10.phys.utk.edu/astr161/lect/Moon/Moon_surface.html) showing the major geological formations of anorthosites and basalts. (c) and (d) Field photographs of terrestrial anorthosites with example from the Paleoproterozoic Damiao anorthosite-gabbro complex in the North China Craton. Note the coarse crystals of anorthitic plagioclase in (d).

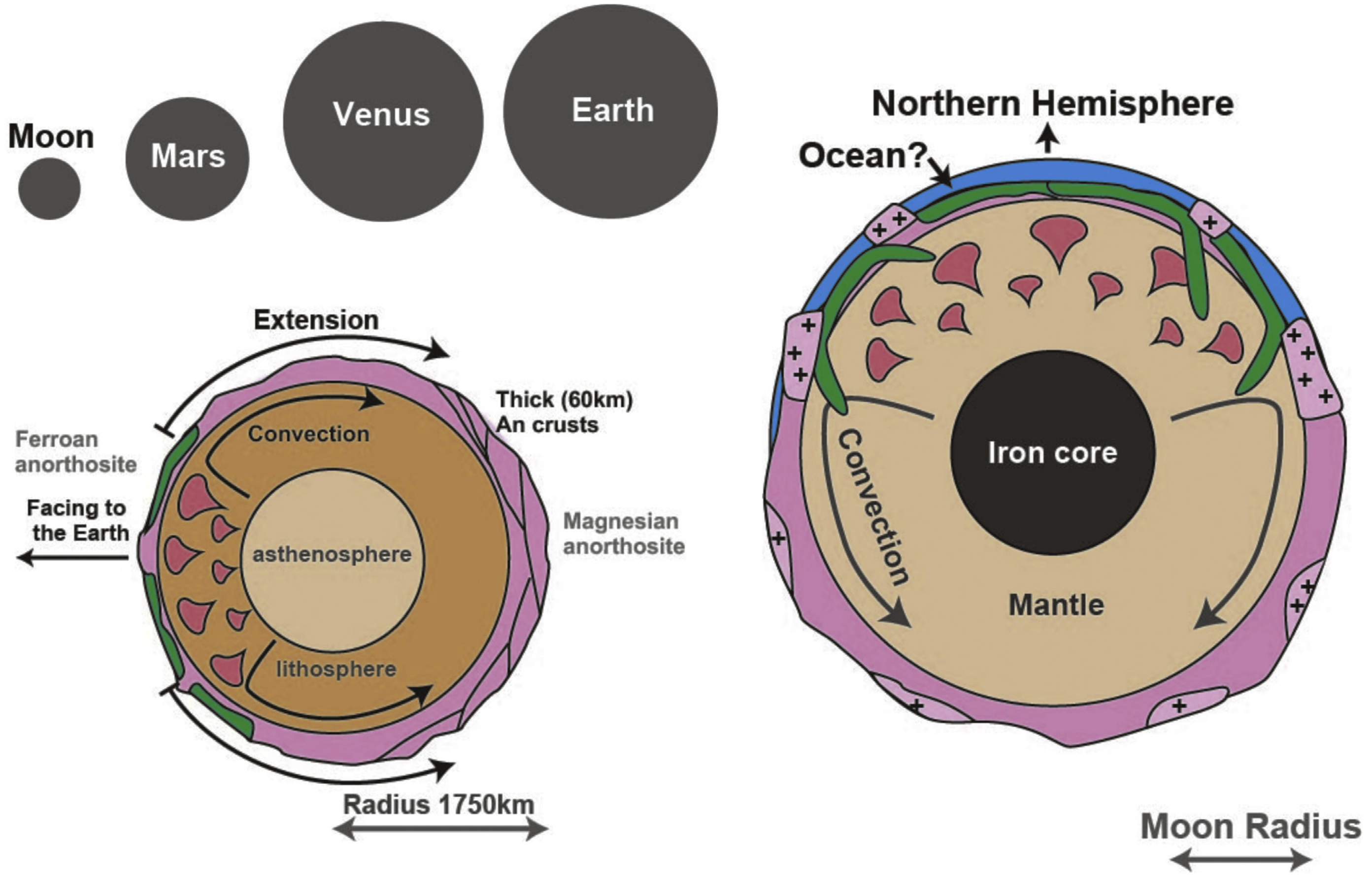


Figure 3. A comparison of the relative size of Moon, Mars, Venus and Earth (top panel). Mars is almost double the size of Moon, whereas Venus and Earth have similar size. The left side figure shows a schematic cross section of Moon and the right side figure shows that of Mars. The nearside of Moon facing the Earth is composed dominantly of basalt with thin (20–30 km) anorthositic crust, with mantle upwellings, whereas the far side has a thicker (average 60 km) anorthositic continental crust (marked as An crusts in the figure), possibly due to horizontal shortening. In Mars, the northern hemisphere is dominantly oceanic crust, covered extensively covered by sediments, chemically equivalent to andesite.

Wanke and Dreibus (1982), and Ringwood (1990). It is considered that the Earth-Moon system was born dry, followed by the later addition of volatiles. The D/H ratios of Earth and Moon are remarkably similar to those of CI chondrites, whereas the values are different from the H comets and solar hydrogen (e.g., Saal et al., 2013). Albarede (2009), summarizing chronological data on early history of Moon, inferred the Late Heavy Bombardment to be at 4.4 Ga and to have lasted over 50 Myr based on U-Pb and I-Xe ages. In summary, the naked planet Earth was formed through aggregation of dry enstatite chondrite at 4.56 Ga followed by the formation of Moon through Giant Impact. Subsequently, hydrous asteroid bombardment occurred delivering the primordial atmosphere and ocean.

The Giant Impact theory envisages the impactor to be of Mars size to form Moon through extrusion of magma from the Earth. Thus, the bulk chemistry of Moon must not be markedly different from that of the Earth. The metallic core of the Moon is approximately 300 to 400 km in radius and is remarkably small as compared to the radius of the planet (1737 km). In contrast, the Earth's metallic core comprises half of the planet's radius. Thus, the bulk composition of the Moon is clearly different from that of the Earth. The abundance and distribution of iron on the Moon (ca. 3 wt.% in the lunar highlands and 7–8 wt.% in the far side; Lucey et al., 1995) confirm the notion that much of the lunar crust was derived from a magma ocean, and also indicate increasing mafic composition of the lunar crust with depth. The data also indicate that the bulk composition of the Moon differs from that of the Earth's mantle (Lucey et al., 1995). In spite of the difference in bulk chemical composition, critical observations suggest that the Earth and Moon share many common features. For example, data from oxygen isotope studies (Pillinger et al., 2014) show that 'Terrestrial Fractional Line' is nearly identical for both Earth and Moon, comparable to that of enstatite chondrite which is dry, and different from other types of chondrites including carbonaceous chondrite. The oxygen isotope data, together with other elemental characteristics including those of Ti, Si, Cr, Rb or Sr and others indicate similarity between Moon and Earth (e.g., Young et al., 2016), particularly with respect to the fact that the source chondrite was dry. This discovery suggests that the Earth was also born dry as naked planet (Maruyama et al., 2013), even though it was covered by magma ocean.

The dominant occurrence of craters on Moon, Mars and Mercury and several other asteroid belts indicate extensive bombardment on the surface. The timing of the major bombardment on the Moon was previous roughly estimated as 3.8 Ga, although more recent estimates based on precise isotope dating techniques consider that the bombardment occurred earlier, at ca 4.2 Ga (Hopkins and Mojzsis, 2015). The consolidation of lunar magma ocean has been recently estimated as 4.37–4.46 Ga using zircons in mare basalts and those from the orange soils (lunar glass) (Borg et al., 2015, and references therein). The late heavy bombardment must be younger than this event, with workers suggesting it as around 4.4 Ga (Albarede, 2009).

The concept of Late Heavy Bombardment (Ringwood, 1990) envisages the timing after consolidation of magma ocean. This concept is based on the abundance of PGE (platinum group elements) in mantle peridotites on the Earth as studied in xenoliths brought up by strongly alkaline magmas such as kimberlite, lamproite, and carbonatite. The original idea proposed by Ringwood (1990) envisaged that when the Earth was formed, strongly siderophile elements must have all entered in to the core. Nevertheless, the mantle xenoliths show high PGE element abundance and therefore it was speculated that the upper mantle must have received extensive asteroid bombardment later. The late heavy bombardment added the PGE to the superficial part of mantle. It is more reasonable to consider that not only PGE elements, but other components including volatiles were also added together, during the CI chondrite-dominated asteroid bombardment at 4.4 Ga. In the terrestrial scenario, the hydrogen isotope features of ocean water suggest the source of water on Earth as carbonaceous chondrites (Maruyama et al., 2013).

Abundant silicate-melt inclusions have been observed in lunar samples which provide strong indication for late stage silicate-liquid immiscibility and the presence of traces of volatiles including water vapor (Roedder and Weiblen, 1970). The presence of water has also been inferred from melt and fluid inclusions trapped within olivine crystals from the mare basalts (Saal et al., 2013) and based on FTIR spectroscopy of plagioclase grains in anorthosites (Hui et al., 2013). Although debate surrounds the origin water on the Moon as to whether it indicates a more prolonged crystallization of the lunar magma ocean or a dry Moon, the more plausible scenario is that Late Heavy Bombardment delivered the water, since most of the older surface rocks do not carry any hydrous phase. Moreover, seismicity in the Moon is restricted to mantle depths of approximately 1000 km, suggesting the presence of water or other volatiles in deep mantle (Zhao et al., 2012). Direct evidence also comes from the transient illuminations or ‘flashing discharges’ on the Moon (e.g., Dollfus, 2000). This phenomenon involves the simultaneous increase of brightness and polarization suggesting outgassing from the lunar interior, and levitation of soil particles from the surface.

3. Formation of primordial continents on Earth during the cooling of magma ocean at 4.56–4.53 Ga

3.1. Petrological composition of the Hadean Earth

In this section, we propose a new model for the primordial continents on the Hadean Earth. We envisage that magma ocean was present on the Earth through the process of collision and amalgamation processes from planetesimals grown in solar nebula. The final giant impact caused extensive melting of the Earth and Moon. Numerical simulation of these processes clearly supports the idea of magma ocean (Melosh, 1989). Geological and geophysical studies support the nearly whole melting of the Moon, because the total amount of plagioclase in the primordial continents ca. 70 km thick on the Moon is nearly equivalent to 10% of the modal amount of lherzolitic mantle (bulk silicate Moon).

Many workers consider that the Hadean Earth was covered by nearly 25 km thick komatiitic crust or fractionated basaltic crust (e.g., Synder et al., 1992; Elkins-Tanton, 2008) and no anorthositic rocks were present. Using the bulk chemical composition, and in conjunction with Bulk Silicate Earth model, pyrolite model, or most primitive lherozolitic peridotite model as analog for the magma ocean composition, Synder et al., 1992 calculated the fractionation of magma ocean, which was recently revised by Elkins-Tanton (2008). The results are shown in Fig. 4. Both models are broadly similar, with the distinction arising from different pyroxene phase diagrams and complete or incomplete olivine separation during the solidification of magma ocean. Snyder et al. (1992) considered that pigeonite appears at high temperature after orthopyroxene than Ca-rich clinopyroxene and anorthosite at lower temperature. On the other hand, Elkins-Tanton (2008) did not differentiate the clinopyroxenes. Natural occurrence of Al-depleted komatiites such as the 3.5 Ga Barberton komatiites supports the interpretation of Snyder et al. (1992). Owing to high-potential mantle temperature of Al-depleted komatiite at 1600 °C, the thickness of primordial crust reaches up to 25 km thick, or could have been even thicker. Elkins-Tanton (2008) speculated a thickness of 25 km. The extent of magma ocean is assumed to be two-thirds of the mantle thickness. The bottom 1000 km remained in solid state. Applying the crystallization sequence on the left figure in Fig. 4, the lower mantle (660–2000 km) composition is dominantly olivine (cumulate) which breaks down into perovskite + wüstite. The mantle transition zone (ca. 400–600 km depth range) consists of upper 40% β-olivine +35% majorite +25% clinopyroxene and lower 45% γ-olivine +55% majorite. More majorite in the lower mantle transition zone is reflected by wider stability filed of majorite. The upper mantle 65–400 km depth range consists of 50% olivine +20% orthopyroxene +20% clinopyroxene +10% garnet. The depth range 25–65 km is occupied by rocks composed of 50% olivine +25% clinopyroxene +20% orthopyroxene +5% spinel. The top surface is low-Al komatiite composed of 50% olivine +25% clinopyroxene + 20% orthopyroxene +5% plagioclase (Fig. 4 right panel).

According to the Elkins-Tanton (2008) model, no anorthositic primordial continents are formed, although plagioclase appears (up to 5 modal %) in the komatiitic rocks. The reason is that if pressure is higher than 10 GPa (the approximate boundary between upper and lower mantle), majorite is stable and would act as the sink for Al leading to residual melts being depleted in aluminum. Therefore, the thickness of anorthosite, even if present, would be less than 1 km (Herzberg, 1983).

We further evaluate this in Fig. 5 where the Ol-Cpx-Grt tripe points are shown from 5 to 13 GPa (after Herzberg, 1992) with respect to the crystallization of terrestrial magma ocean. The triple point moves to olivine corner with increasing pressure, indicating larger stability field of majoritic garnet. Assuming the bulk chemical composition of bulk silicate Earth at the point marked by orange filled circle in Fig. 3, the residual melt would be depleted in plagioclase, at pressures higher than 10 GPa, with majoritic garnet as the first phase to crystallize in magma ocean which settles down to the bottom depleting the Ca and Al contents in the residual melt (Fig. 3). Due to this process, the uppermost mantle is depleted in Al and Ca, and therefore it is difficult to form thick anorthositic crust on the surface (Fig. 5).

3.2. Fractional crystallization of peridotite melt (magma ocean)

Modeling the fractionation of magma ocean using MELTS programme (Fig. 6) shows that olivine starts to crystallize at 1640 °C, followed by Opx at 1615 °C, Plag at 1291 °C and finally Cpx at 1234 °C, near the surface after the consolidation of nearly 85% of the magma. The final 15% residue is the most fractionated FeO-enriched KREEP rocks with ores including U often associated with P- and REE-bearing minerals.

Fig. 6a shows solidification of lunar magma ocean (% solidified), vs. pressure (GPa). The early, and hence the first liquidus phase is olivine that crystalizes at $T = 1840$ °C, at 23 GPa (ca. 1500 km depth). Olivine is the only phase to continuously precipitate and move to the bottom of the magma ocean to form dunite cumulate until 40% of the total volume of magma ocean is consolidated at around 1815 °C at 1.2 GPa. At this depth and temperature, the next liquidus phase orthopyroxene will crystallize together with olivine until 1291 °C at 3 kbar, leaving 75% solidified cumulates (Ol-Opx biminerallic rocks above the dunite cumulates at the bottom of magma ocean). Then anorthite ($CaAl_2Si_2O_8$) joins with Ol + Opx

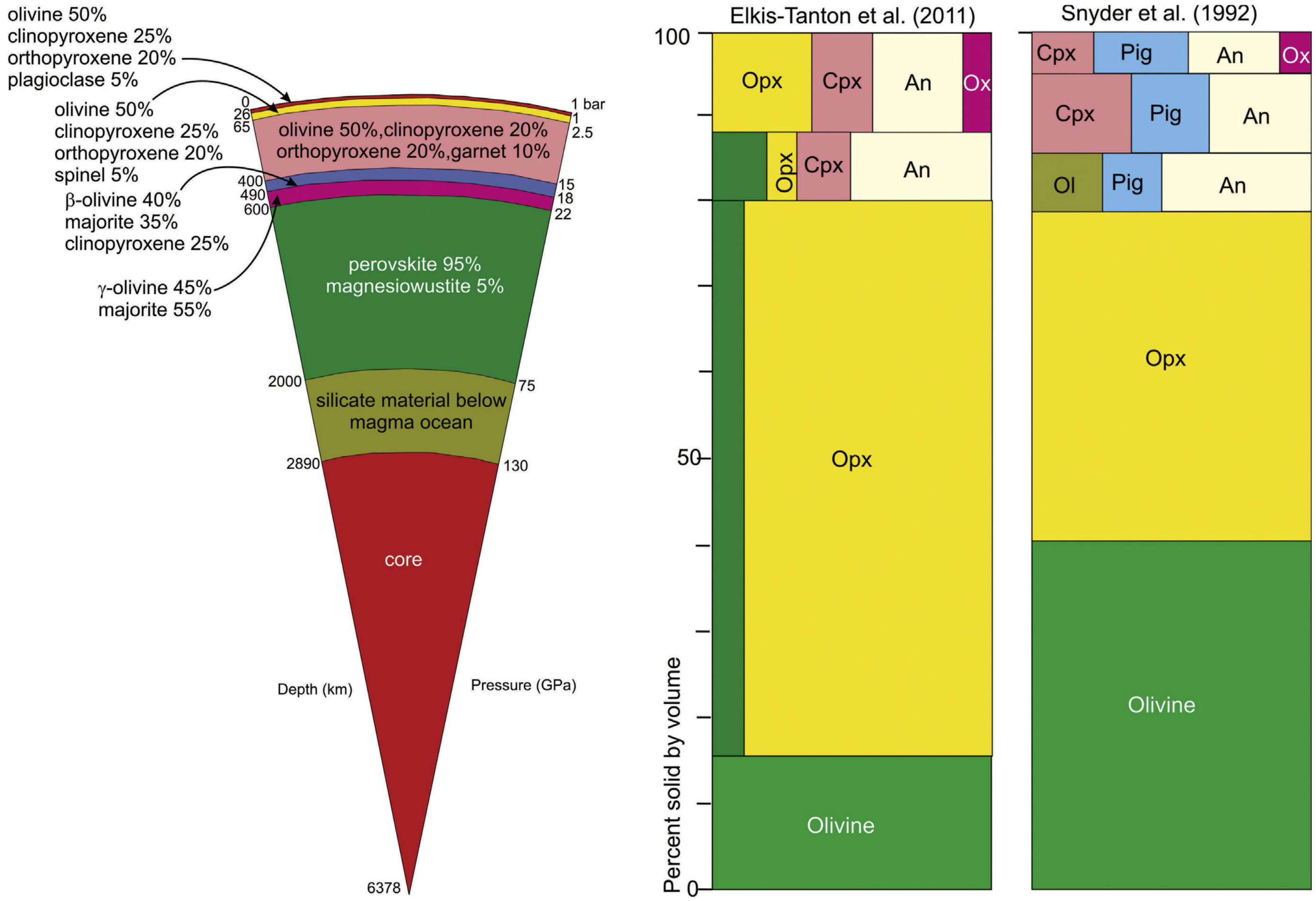


Figure 4. Left panel: Mineral assemblages solidifying from a 2000-km deep terrestrial magma ocean as proposed by Elikins-Tanton (2008). Lunar cumulate stratigraphy with solidifying mantle mineral assemblages as compared by Elkins-Tanton et al. (2011) (right panel) with their study and those from Snyder et al. (1992) (middle panel). Solidifying mantle Opx = orthopyroxene; Cpx = clinopyroxene; Pig = pigeonite; Ox = oxides, including ilmentite; An = anorthite feldspar (plagioclase). The fraction of well-mixed phases is represented by the width of mineral boxes. Neither models envisage primordial anorthositic continental crust.

which generating gabbro until 1234 °C at 2 kbar. From there, no olivine appears, instead, clinopyroxene appears along with Ca-plagioclase. The remaining 12% melt of the magma ocean is enriched in incompatible elements such as K, P, REE, Ti, Zr, Mn, Mo, U, Th, Nb, and others. The extent of fractionation depends on mineral separation in the remaining residual magma. Because of this uncertainty, no calculation is performed for this factor. It can be estimated that ca.12% of the total magma ocean must have occupied the top-most domain.

Fig. 6c shows the chemical variation trend of major oxides, SiO_2, Al_2O_3, FeO, MgO CaO and Na_2O, during fractional crystallization. The SiO_2 remains nearly constant from 45.5 to 48 wt.%, corresponding from dunite, through troctolite to gabbro. The Al_2O_3 content shows a marked change from 4 to 15 wt.% in the gabbro. CaO follows a similar trend from 3 to 10–15 wt.% for the gabbroic rocks. FeO increases from 8 to 15 wt.% for gabbro, and even more at final stage although no calculation was performed for the final 15 wt.% residual melt.

To compensate the FeO trend, MgO decreases rapidly from 38 to 10–8 wt.% at the final stage in the KREEP melt. Na shows only slight change (from 0.1 to 1.0 wt.%) because the initial value is small. Fig. 6d shows the $Mg^{\#}$ value (molar ratio of mineral, Mg/(Mg + Fe)). The value of $Mg^{\#}$ of melt rapidly changes from 0.89 to 0.45 during the fractional crystallization.

3.3. *Formation of primordial continents on the Moon*

Here we evaluate, based on MELTS program calculation, the origin of anorthositic primordial continents by incorporating the gravitational effect of plagioclase over coexisting melt to account for the floatation of plagioclase.

Due to the density crossover at 7 kbar, corresponding to 120 km depth on the Moon, Ca-plagioclase floats at 120 km depth, but sinks leading to the formation of anorthosite monominerallic layer (Fig. 7 left panel). With increasing crystallization, the residual melt becomes enriched in FeO, hence the density-crossover point moves to the surface, indicating that ca. 100 km thick anorthosite layer would finally float on the surface. During the upward movement of anorthosite layer, gabbroic layer with Ca-plagioclase and orthopyroxene with minor olivine, and a top surface of komatiitic chilled margin may also be present.

Thus, upper komatiite-gabbro mafic layer (<10 km), underplated by anorthosite layer (100 km thick) could be the major structure of primordial continental crust. The final residue formed during the fractional crystallization of lunar magma ocean must be KREEP magma and/or much more fractionated TiO_2- and FeO-enriched KREEP (FeO = 20–25 wt.%). These melts occupy 10–15% for the extremely fractionated KREEP, and are underplated below the anorthositic felsic crust. If so, the thickness of lunar continental

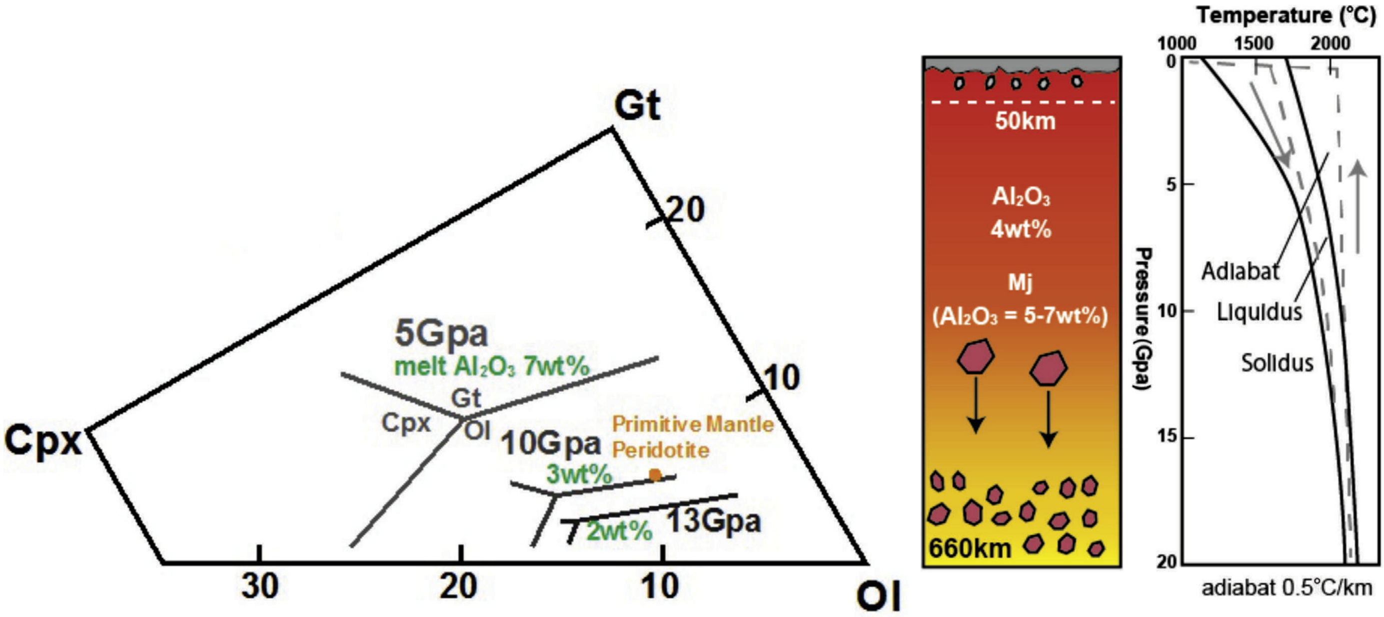


Figure 5. Phase diagram of garnet-clinopyroxene and olivine (after Herzberg, 1992). Pressure dependence of triple point Gt-Cpx-Ol from 5 Gpa to 13 GPa prevents anorthite component in depleted residual melt, because majorite garnet crystallizes in deeper mantle. Stability field of garnet/majorite expands at high pressures. The triple point shift toward the olivine apex with increasing pressure, with garnet stable at higher pressure. Primitive mantle peridotite composition is shown by orange spot. Under high pressure, even higher than 10 Gpa, garnet crystallizes as the early phase (denoted as Mj-majorite in the right hand side figure) and forms the sink for Al and Ca. According to this model, the residual melt would be depleted in plagioclase, making it difficult to form thick anorthositic crust on the surface. Right hand side figure shows liquidus and solidus of mantle peridotite and mantle adiabat down to 20 GPa, with the top portion showing the solidification at deeper mantle. Hence, anorthite component is selectively lost at deeper mantle by crystallization of majoritic garnet.

crust is over 200 km or higher. Seismological observations however, do not support this interpretation.

The most probable interpretation is that the final KREEP residue might have been removed down to the bottom of mantle, either during the fractional crystallization stage (Shearer et al., 2006), or after the solidification of magma ocean, by mantle convection.

We estimate that 40% volume of magma ocean must be monominerallic cumulate of dunite resting on the bottom. Over this layer occurs 40% of biminerallic cumulate of Ol + Opx, and the remaining 10% is plagioclase-bearing Opx + Cpx gabbroic rock. The final residue of 10% is FeO-rich KREEP. Thus, the possible lithologic structure is composed of 80% mantle peridotite, 10% anorthositic primordial continent, and 10% FeO-enriched KREEP (which may or may not be underplated below anorthositic upper crust).

3.4. Geology of the Moon

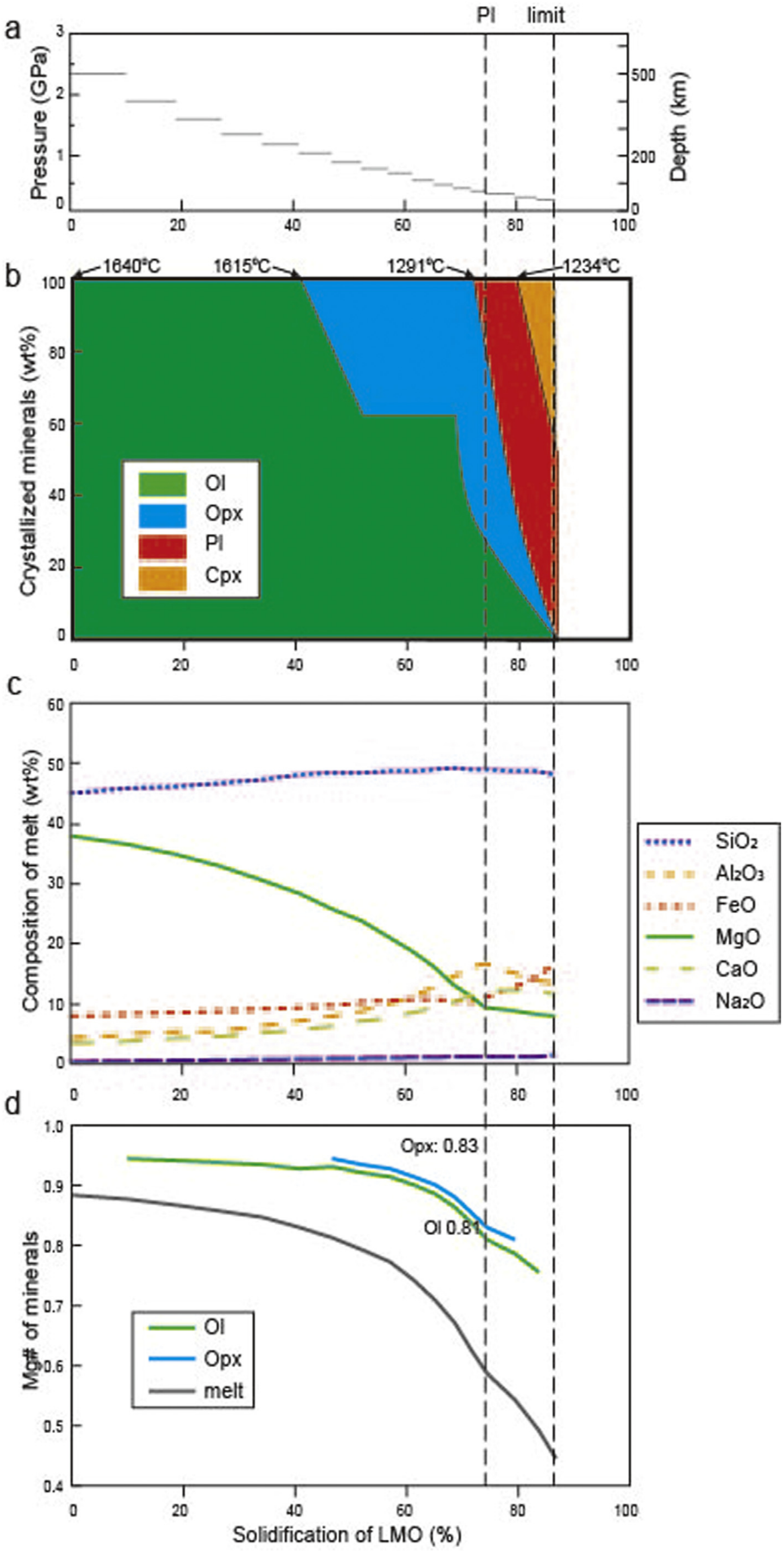
Moon is covered by about 10 m thick regolith layer formed during asteroid impact defined in a recent study as the ‘ABEL [advent of bio elements] bombardment’ by Maruyama and Ebisuzaki, 2016) which occurred at around 4.37 Ga to 4.20 Ga (Fig. 8, right panel; Hopkins and Mojzsis, 2015). The asteroid impacts cratered on the surface of Moon, particularly nearside. The largest one is called Procellarium ca. 3200 km across and characterized by KREEP-like basalt flows after the bombardment (Fig. 8 left panel). Distribution of craters is also related to the eruption of mare basalts ca. 100 Myr after the bombardment, because of mantle rebound convection. The rest of the surface of the Moon is white-colored, and traditionally called as continents representing anorthositic crust. The estimated crustal thickness is 20–60 km for the upper crust in total with 60–100 km thick continental crust on the farside Moon whereas the upper crustal thickness is only 10–30 km with a total 50–60 km thick continental crust on the nearside (Shearer et al., 2006). This contrasting crustal thickness is apparently related to selective bombardment specifically large (3200 km across) to middle size (1000–600 km across) of asteroids on nearside, and related frictional heating and volatile input into mantle on nearside. Resultant mantle heating and lowering of viscosity promoted mantle upwelling to thin the continental crust on nearside under extension, whereas compression occurred on the farside.

Due to the extensive bombardment during 43.7–42.0 Ga, the timing of solidification of magma ocean had been obliterated in the mineral ages or isochron ages. Based on a comparison of several model ages and isochron ages, Borg et al. (2015) considered the age of primordial continents to be older than 4.40 Ga, and the timing of eruption of the final residue of magma ocean (KREEP) as 4389 ± 45 Ma.

3.5. Formation of terrestrial anorthosite

A petrologic model of primordial continents can be formulated based on a speculation of fractional crystallization (Fig. 9). At pressures less than 7 kbar, plagioclase and olivine sinks down with gabbroic layer at the bottom. At pressures higher than 7 kbar, plagioclase floats above to generate anorthositic cap and olivine sinks down to form ultramafic cumulate layer. The density cross over point migrates upward through time when the residual melt becomes more Fe rich. Thus, during the early evolutionary history of our planet, just after the consolidation of magma ocean (at ca. 4.53 Ga), a concentric structure can be envisaged with primordial anorthositic crust on the top which is anchored by KREEP gabbro underneath, followed by dunite layer. Bottom of the upper mantle is speculated to be majoritic garnet cumulate. In the subsequent stage (at ca. 4.50 Ga), mantle convection was initiated with rising plumes and komatiitic oceanic crust.

In contrast, if we consider density crossover, the scenario would be different. Curves of density versus pressure for peridotitic melts


a
Pl
limit
Pressure (GPa)
3
2
1
0
Depth (km)
500
200
0
0
20
40
60
80
100
b
1640°C
1615°C
1291°C
1234°C
Crystallized minerals (wt%)
100
80
60
40
20
0
Ol
Opx
Pl
Cpx
0
20
40
60
80
100
c
Composition of melt (wt%)
60
50
40
30
20
10
0
SiO2
Al2O3
FeO
MgO
CaO
Na2O
0
20
40
60
80
100
d
Mg# of minerals
1.0
0.9
0.8
0.7
0.6
0.5
0.4
Opx: 0.83
Ol 0.81
Ol
Opx
melt
0
20
40
60
80
100
Solidification of LMO (%)

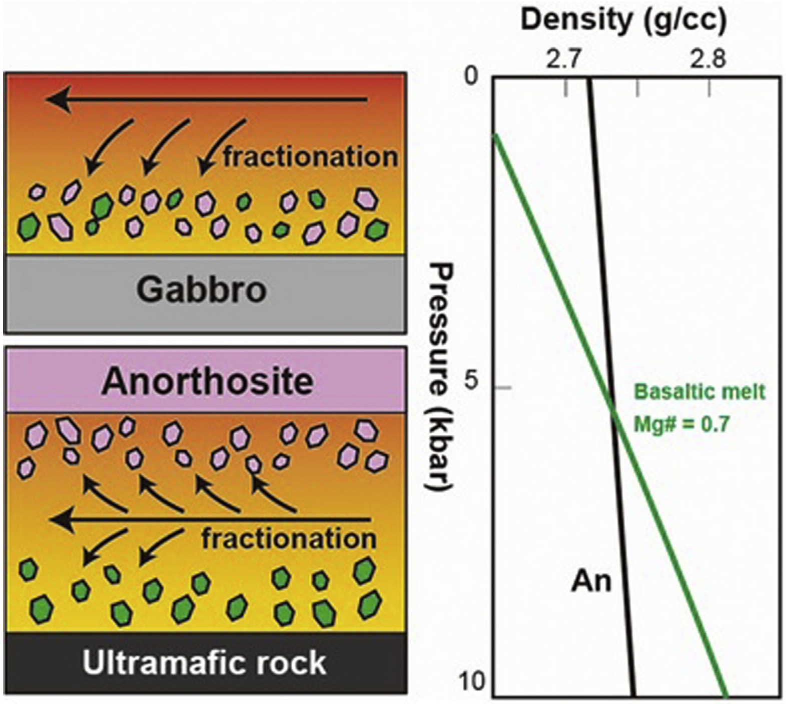


Figure 7. Schematic illustration of the formation of anorthosite when density-inversion exists between melt and anorthite. The density of melt was calculated by using formula of Lange and Carmichael (1987). For the shallow part of the magma where density of plagioclase is larger than that of the melt, plagioclase along with olivine sinks down and gabbroic cumulate will form. For the deep part of the magma where density of plagioclase is smaller than that of the melt, only plagioclase will float and pure anorthite will form. The petrogenetic proposals are after Kushiro and Fujii (1977).

(Mg$^{\#}$ value 0.8) and olivine (Fo 90) intersect at about 8 to 9 GPa (Fig. 9; after Ohtani, 1985). At depths below the density cross-over point, olivine which is the first liquidus phase, floats upward whereas above the density cross over point olivine will sink down to form monomineralic rock (dunite) layer, possibly as thick as 100 km before the next liquidus phase appears. Although the role and strength of mantle convection must also be considered, this anomalously thick dunite layer would separate melt zone in the upper mantle into two domains. This separation is a critical constraint to retain the concentration of Ca and Al as high as or more than 4 or 5 wt.% to form thick anorthositic layer. Without the dunite barrier, fractional crystallization would result as proposed in the model of Herzberg (1983). The formation of such anorthositic layer through density cross over might be similar to the process on the Moon. The anorthositic layer can be formed at about 14 km depth corresponding to about 7 kbar (density cross over point by Fujii and Kushiro, 1977), below which the rock will break down to an assemblage of grossular, kyanite and quartz although the bulk composition is the same. At shallower domains above the density cross over point, plagioclase floats and olivine sinks. Thus, dominantly monominerallic rocks of anorthosite can be formed at pressures less than 7 kbar, and at deeper levels, olivine and plagioclase sink down to make gabbro. The thickness of primordial anorthositic crust resulting from the fractionation of magma ocean would be significantly different between Moon and Earth, because of the difference in the plagioclase stability arising from the six times higher gravity on Earth as compared to that of the Moon.

3.6. Formation of layered structure of mantle after the solidification of magma ocean

Here we evaluate the differences between Moon and Earth during the stage of magma ocean.

Considering the fractional crystallization of upper mantle, yet another density cross over can be envisaged (Fig. 10), in addition to the plagioclase density cross over mentioned above. The density crossover of olivine occurs at around 200 km depth (Ohtani, 1985). Olivine was a liquidus phase when magma ocean started to consolidate. As olivine dominates the mantle component (ca. 60%), extensive amounts of olivine crystallized to sink down to depths of 200 km, below which olivine floats to thicken and form olivine layer centered at ca. 200 km as discussed in a previous section. Through time, the olivine floor attains a thickness of the order of ca. 100 km and functions as a barrier between the upper and lower half of upper mantle, inhibiting extensive convection within the magma ocean. Enrichment of the remaining CaO and Al_2O_3 becomes nearly double in the residual liquid. Following this, in the upper half (<200 km depth) of the upper mantle, olivine and plagioclase

Figure 6. Results of calculations as a function of solidification vol.% of the lunar magma ocean by using MELTS/pMELTS. We used thermodynamic equilibrium software, pMELTS for $P > 1$ GPa (Ghiorso et al., 2002; #1139) and rhyolite-MELTS for $P < 1$ GPa (Gualda et al., 2012; #1140). (a) Schematic illustration of input pressure condition for each step. One step represents 1/10th of the residual magma ocean volume. (b) Composition of melt. (c) Modal abundance of crystallized minerals for each step. (d) Mg$^{\#}$ of crystallized olivine and pyroxene along with Mg$^{\#}$ of coexisting melt. Results for >87% of solidification of LMO are blank because the MELTS program outputs error for the highly evolved magma. In (a), the magma starts to consolidate from depth to the surface through time. Plagioclase stability line is also shown. In (b), the proportion of crystallized minerals with respect to Ol, Opx, Plg and Cpx is and temperature limits are also shown. In (c), the most notable change is the MgO content, from 38 wt.% to about 8 wt.% when the melt is solidified. Aluminum gradually increases from ca. 4 to 15 wt.%, when plagioclase starts to crystallize, followed by a late decrease. In (d) the variation in Mg$^{\#}$ in major minerals with progressive solidification of magma ocean is apparent. LMO—lunar magma ocean.

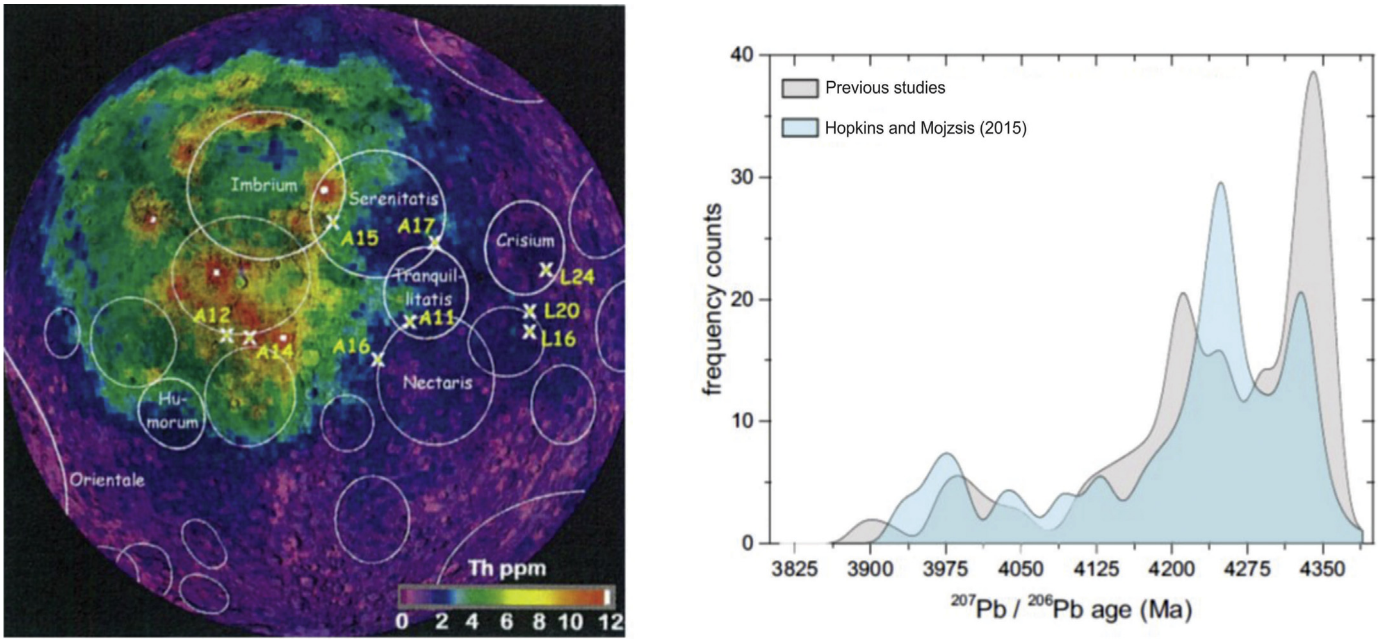


Figure 8. Geology of the Moon, nearside and farside (left panel; after Bradley et al., 2000). Cross section shows the primordial anorthositic crust thick on the far side (<100 km) thinner on nearside (<50 km) (right panel; after Hopkins and Mojzsis, 2015).

fractionation takes place. Thus, yet another density cross over plays a crucial role to form primordial anorthositic continents in addition to generating the residue of KREEP basalts or gabbros. The final extremely iron rich mafic melts must be substantially enriched in LILE including U, Th and K, as well as P and Cl. These elements must have formed ores within the primordial continents and are also important components in terms of the prebiotic chemical evolution of first life on this planet.

3.7. Double-layered mantle convection in magma ocean

Solidification of the magma ocean starts from bottom of upper mantle to shallower levels. This is because mantle adiabat intersects with the liquidus curve at depth, and moves to shallow levels during cooling. The model of a ca. 2000 km thick magma ocean assumes different densities for the upper and lower mantle, by up to 2 to 3 orders of magnitude. This is true in the case of a solid mantle as on the present day Earth as confirmed from geophysical studies. However, if the mantle was in liquid state, the viscosity is different between upper and lower mantle domains which would mean double-layered convection as against the whole mantle convection. Thus, as a reasonable assumption, both Ca and Al would also be present in the upper mantle.

In this case, there will be no magma mixing between the upper and lower mantles, both convecting as independent cells, which could have been the plausible scenario in the Hadean Earth. In this case, CaO and Al_2O_3 are largely equally divided between the upper and lower mantle. Therefore, the fractionation of upper mantle magma ocean is the key to dictate the formation of primordial continents.

During fractional crystallization of the magma ocean, the density cross over point moves upward as the residual melt becomes iron rich, generating a ‘floating’ cumulate layer of nearly 100–200 km thick dunite. The top layer is composed of fractionated basalt and thin komatiite. In the final stage, the anorthositic layer is capped by thin komatiite and fractionated basalt.

In summary, a thick (up to 35 km) anorthositic crust is petrologically possible capping the Hadean primordial continent, below which would be a mafic crust composed largely of KREEP basalts of variable thickness and cumulate layers of olivine, and majorite beneath. The formation of primordial continents on both Earth and Moon might have been linked to the magma ocean with the essential difference that on Earth, with nearly six times the gravity of Moon, extensive fractional crystallization might have been facilitated. On the Moon, density cross over between magma and plagioclase played a key role after the fractionation of the liquidus phase of olivine. After olivine crystallized, the cotectic surface between olivine and plagioclase (Ohtani, 1985) explains the appearance of plagioclase after olivine. Olivine is dense enough to move down to the bottom of magma ocean and plagioclase as the next phase to crystallize follows the density cross over point. A pressure of 7 kbar corresponds to approximately 100 km depth on the Moon, whereas on the Earth this is only 21 km. In this region, plagioclase can form nearly monomineralic rock as thick as 20 to 40 km. Fractional crystallization and residual Fe enrichment in the residual liquid would lead to increase in density. Thus, the density cross over point migrates to shallow level, and finally to the surface to form primordial continents. Fractionated iron rich magma underplates to form the lower crust of primordial continents, or intrude as dykes in the overlying buoyant anorthositic continental crust when during the consolidation of the magma ocean. This mechanism explains the situation in Moon.

The critical problem is the fate of CaO and Al_2O_3 which are major components of felsic crust both in the Moon and Earth. If magma ocean on the Earth fractionated and developed not only upper mantle but also lower mantle, then consolidation must have started at the middle domain of the lower mantle because of liquidus intersecting first with the mantle adiabat. At mid mantle depths, the first liquids phase to crystallize would be perovskite (both Ca and Mg types) (Kato et al., 1988). The Ca perovskite contains both Ca and subordinate amounts of Al_2O_3. If Ca perovskite appears first in the lower mantle, then no Al and Ca will remain in the upper mantle. This could be the major reason why the Earth had no primordial continent. This would also mean that the primordial continent is unrelated to the presence of water.

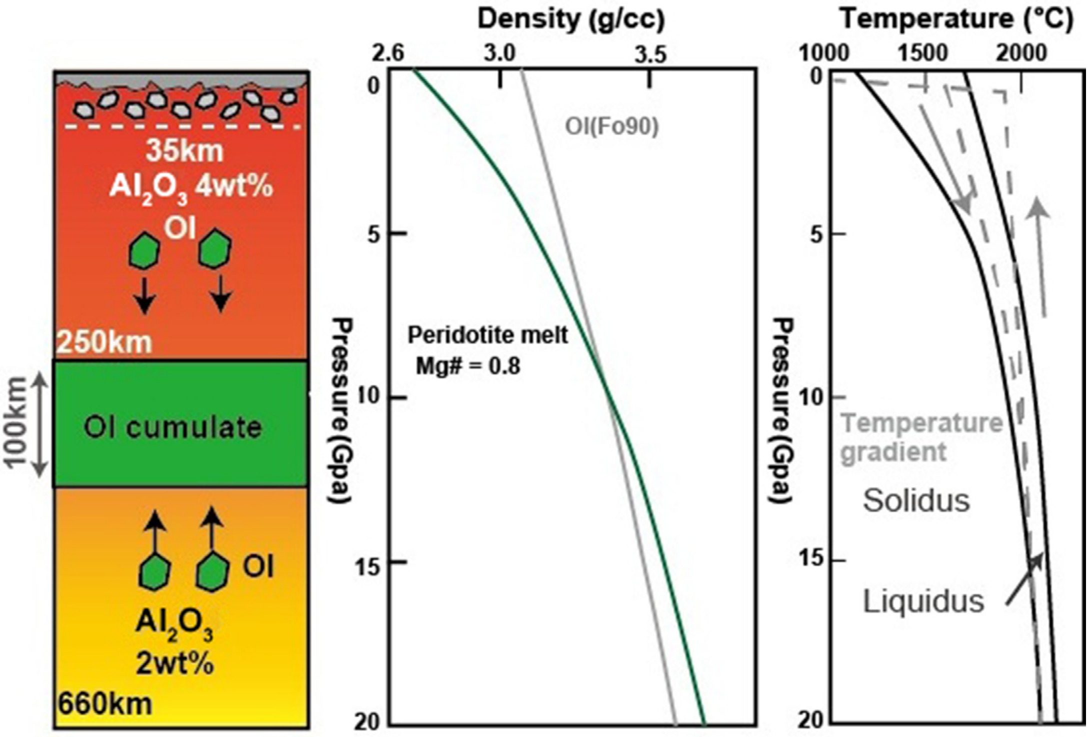


Figure 9. Schematic illustration of the terrestrial magma ocean. From left to right, in the first panel, the depth of magma ocean is assumed to be 660 km, corresponding to the depth of the upper mantle. The second panel shows the relationship between the adiabat of the terrestrial magma ocean and liquidus/solidus line for ultramafic melt. Since adiabat crosses the liquidus for the deep part of the magma ocean, majorite starts to crystallize first. The third panel shows the relationship between the adiabat of the terrestrial magma ocean and liquidus/solidus line for ultramafic melt. Since the adiabat lies between the liquidus and solidus for the pressure ranging from 5 to 15 GPa, olivine may crystallize. The fourth and fifth panels show the density of peridotite melt and olivine, implying that the density of melt becomes larger than olivine for $P >$ ca. 10 GPa. The density of melt was calculated by using formula of Lange and Carmichael (1987). Therefore, the crystallized olivine will cumulate at the depth where the density of melt overcomes that of olivine.

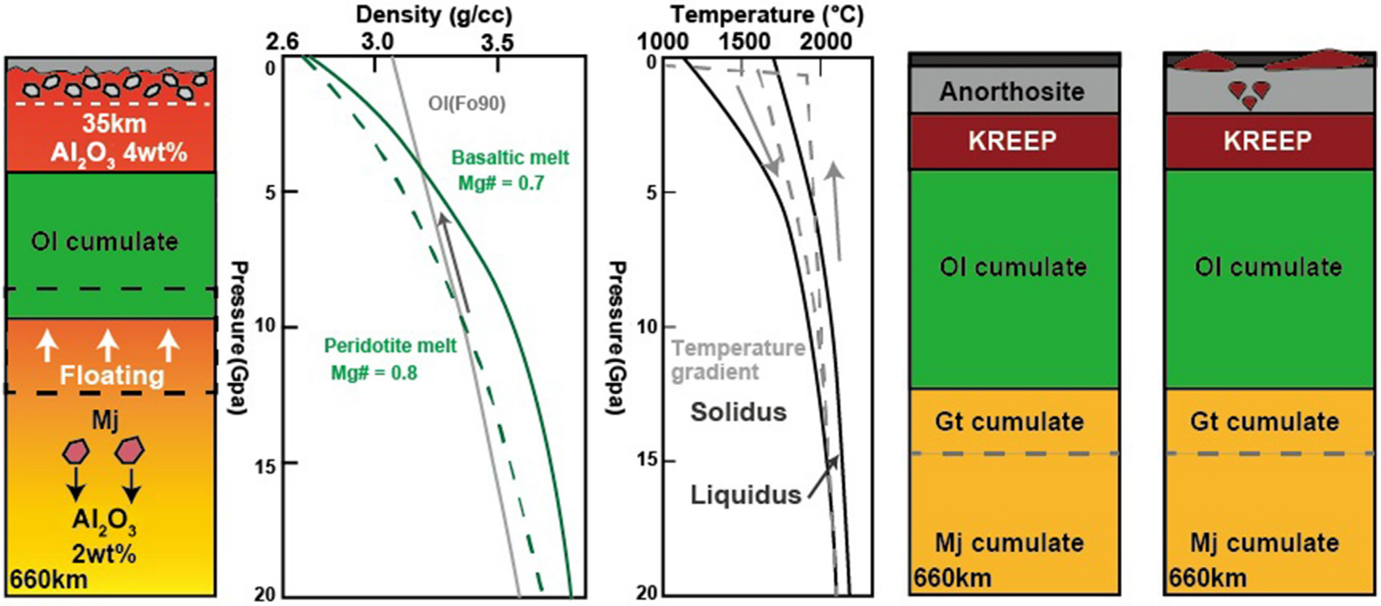


Figure 10. Schematic illustration of the magma ocean. The depth of magma ocean is assumed to be 660 km, corresponding to the depth of the upper mantle. The three panels on the left side indicate that as the residual melt becomes basaltic and $Mg^{\#}$ decreases, the depth where the density of melt overcomes that of olivine becomes small. Therefore the cumulate of olivine may go up as the density of melt increases. The density of melt was calculated by using formula of Lange and Carmichael (1987). The right side two panels show schematic representation of layers formed after the terrestrial magma ocean solidified. The residual melts are enriched in Fe, K, P and REE and form KREEP basalt/gabbro. The top layer is composed of fractionated basalt and thin komatiite. The cross sections of the final stages in the right side panels how that anorthositic layer is capped by thin komatiite and fractionated basalt. The lower crust is KREEP, followed downwards by olivine cumulate layer (dunite), garnet cumulate layer and majorite layer at the bottom. Mj—majorite; Ol—olivine; Gt—garnet. Owing to the density crossover between olivine (Fo = 90) and residual melt, a thick olivine cumulate layer 200 km thick separates the magma ocean into two zones, keeping high Ca and Al contents in both zones. Because of this, anorthosite component remains in the upper melt zone which produces thick anorthositic crust on the top. Note the presence of thick KREEP lower crust in the right side panels, which may not be gravitationally stable.

4. Difficulty for survival of felsic crust on the hot mantle in the Hadean

Subduction erosion or tectonic erosion is an efficient process to destroy continental crust in convergent margins (e.g., Von Huene and Scholl, 1991; Yamamoto et al., 2009; Santosh, 2010; Stern, 2011). Intra-oceanic arcs with continental crust are dragged into the deep mantle, captured by descending oceanic lithosphere, and the net effective buoyancy of these overrides the buoyancy of the island arc. This is true even in the case of relatively big microcontinents on the oceanic lithosphere, such as those floating on the Indian ocean plate which in future will be totally subducted together with the Indian oceanic lithosphere. Similar fate also awaits Falkland plateaus and other small continental remnants in the Atlantic ocean.

Tectonic erosion thus plays a critical role in decreasing the volume of continental crust on the Earth. Subduction erosion is another process to tectonically erode the hanging wall continental crust into deep mantle as seen along the subduction margin in the Pacific realm (Yamamoto et al., 2009). Presumably similar process also operates along other subduction zones on the globe from Mediterranean through Middle East to Himalaya and the Indonesian margin down to New Zealand. Along this plate boundary, a number of continents were delivered to the northern margin of Paleo Tethyan ocean from west to the east. The African collision to form Alpine chain and Saudi Arabia colliding to form the Middle East Orogenic belts, the Indian collision to form Himalaya, several microcontinents in the southeast Asian margin and the megablocks of Indonesian region, together with the ongoing continent subduction against intraoceanic arc of Banda Arc and Campbell plateau of New Zealand all involve the destruction of continental crust.

Prior to the advent of plate tectonics (before 4.4 Ga), with plume dominated upwelling and downwelling, tectonic erosion might have been powered by mantle convection. Mantle convection was much more extensive and vigorous in the Hadean Earth. Mantle potential temperature was around 1600 °C, almost 300 °C higher than that of the modern Earth. The surface temperature of Earth during the magma ocean stage was around 1600 °C (e.g., Trail et al., 2011), but this isotherm migrated to the depth of 660 km at the boundary between upper and lower mantles (Ito and Takahashi, 1987; Maruyama et al., 2007). The mantle potential temperature at 3.8 Ga was ca. 350 °C lower (Ohta et al., 1996; Komiya et al., 1999). These values demonstrate ca. 250 °C decrease over the last 3.8 Ga. However, during the Hadean, the temperature decrease of 250 °C within a very short period of 4.56 to 4.0 Ga resulted in extensive mantle convection in the upper mantle as compared to that in the subsequent Earth history. Under such extensive mantle convection, the primordial continents on the surface of the Earth which were penetrated by numerous mantle plumes from below could have been disrupted and destroyed by convection, particularly after the initiation of plate tectonics at 4.4 Ga. Extensive mantle convection on the Moon indicates presence of primordial continents even in the Moon's deep mantle, as recently documented by seismic tomography (Zhao et al., 2012). The P and S wave velocity perturbation at given depth fluctuates ± 2% which is equivalent to the velocity perturbation within the Earth. In the case of the Earth, such perturbation can be explained by horizontal temperature difference which corresponds to 1000 to 2000 K variation. Although temperature variation is the dominant factor, such features can also result from compositional variation. Since the Moon is smaller than the Earth, P and S wave perturbations cannot be explained by temperature variation alone, and the dominant factor would be compositional variation. The lesson from the Moon suggests that the primordial continents may have been destroyed and subducted even back to the time of their formation before plate tectonics started on Earth at ca. 4.4 Ga.

5. Discussion

In this paper, we have speculated the presence of primordial continents on the Earth equivalent to lunar anorthositic crust. This is different from the previous ideas which refuted the presence of anorthositic crust and instead, envisioned a komatiitic crust with a thickness of 25 km or more. We will now discuss why we disagree with the komatiitic crust model, and then move on to evaluate the presence of diversified surface environment if primordial continents were present.

5.1. Previous models

Most geologists and petrologists have long believed that the Earth was covered by komatiite and fractionated rocks right after the solidification of magma ocean at 4.53 Ga. Primordial continents similar to those on the Moon were absent. However, Kawai et al. (2009) was the first who proposed the idea of anorthositic primordial continents, based on simple petrologic modeling. High-pressure stability limit of Ca-plagioclase is ca. 10 kbar at magmatic temperature, implying much thinner continental crust, compared to the Moon. Assuming the density crossover of anorthosite with respect to coexisting basaltic magma is 7 kbar (Fujii and Kushiro, 1977), monomineralic anorthosite can be formed above this depth, 21 km. Therefore, Kawai et al. (2009) speculated that primordial continents were present with 21 km thick upper crust with lower KREEP crust, although the exact total thickness was unclear. They have assumed also the topmost magma ocean contained anorthite component if the magma composition was similar to bulk silicate Earth.

Although Herzberg (1983) pointed out the majorite crystallizes earlier at greater depth from magma ocean, the topmost residual magma cannot contain anorthite component. This prompted Snyder et al. (1992) and Elkins-Tanton (2008) to propose the model of komatiites and no anorthosite on the Hadean Earth. However, the density crossover of anorthite over coexisting melt at 7 kbar, and olivine density vs. coexisting melt at 12 GPa, the majorite problem can be solved, with the possibility that primordial continents formed on the Hadean Earth.

We further extended the model of primordial continental crust, using MELTS program, together with subsolidus phase relations of CaO-Al_2O_3-SiO_2 system. This yields the three layers model of primordial continental crust. The upper felsic crust is composed of top anorthosite with gabbro and komatiite, and lower grossular garnet + kyanite + quartz assemblage. The lower mafic crust is KREEP basalt in composition. Total thickness could be over 100 km. Another possibility is that the FeO-enriched lower KREEP crust may have not been present. Instead, it could have been removed down to deep mantle before the final solidification of magma ocean.

5.2. Primordial dry continents followed by "ABEL bombardment": a turning point toward habitable planet

The Earth was formed at 1.0 AU and $T > 1300$ K where Na and K were nearly evaporated (Imaeda and Ebisuzkai, 2017), and thus the planet was born naked without ocean and atmosphere (Maruyama et al., 2013). The stages of evolution of the early Earth from (a) after the consolidation of magma ocean, (b) through the breakup of primordial continents, and (c) finally to the Late Heavy Bombardment stage, with subduction and annihilation of the primordial continents which were dragged down to the deep mantle are schematically illustrated in Fig. 11.

Oxygen isotope ratio of the Earth clearly indicates that the origin of Earth-Moon system was derived from Enstatite chondrite which is free of volatiles, consistent with the naked planet earth model.

The Tandem model considers that rocky planets in general are always dry when formed. Advent of bio-elements ("ABEL" model, Maruyama and Ebisuzaki, 2016) hence seem to be accidental, particularly, the birth of ocean and atmosphere became possible through the gravitational resonance by Jupiter with Saturn and another Gas Giant. Large amounts of icy asteroids next to the Saturn at 4–5 AU were destroyed and transported to the inner rocky planets including Earth. Through this event, the Earth's radius increased by about 10 km. These events were deduced from the records on the Moon (Maruyama and Ebisuzaki, 2016).

To summarize the events on the Hadean, the earth was born as dry planet first at 4.53 Ga followed by extensive bombardments by icy asteroids to transport volatile components approximately 190 Myr after the solidification of magma ocean. It was the turning point for the Earth becomes life-sustaining planet (Fig. 12).

5.3. *Lost continents*

The primordial continents on the Earth are believed to have been destroyed and subducted into the deep mantle (Komabayashi et al., 2009). The anorthositic primordial crust on the Moon was also partly removed into the mantle, although most part still remains on the surface due to the smaller size of the planet as compared to that of Earth. Since the Earth is large enough to be characterized by extensive mantle convention, the destruction of the primordial crust was near total. Recently, anorthosite has also been discovered on the surface on Mars (Carter and Poulet, 2013), a planet with only one tenth volume as compared to Earth. Thus, Mars is small enough to preserve the primordial continents when magma ocean was consolidated.

The D″ layer at the core-mantle boundary (CMB) of Earth has been correlated to the presence of subducted slab graveyards in previous studies (e.g., Maruyama et al., 2007). This layer possibly involves the destroyed and deep subducted components of the primordial continents on globe when the magma ocean was consolidated, and if this is true, it calls for new models on the thermal structure of the Earth. The prevailing concept that accretion and amalgamation of asteroids during the Giant Impact stored the thermal budget of planet Earth with a gradual release of the thermal energy through time implies that radiogenic elements are more or less homogenously distributed in the mantle. However deep subduction of the primordial crust to CMB depth and its

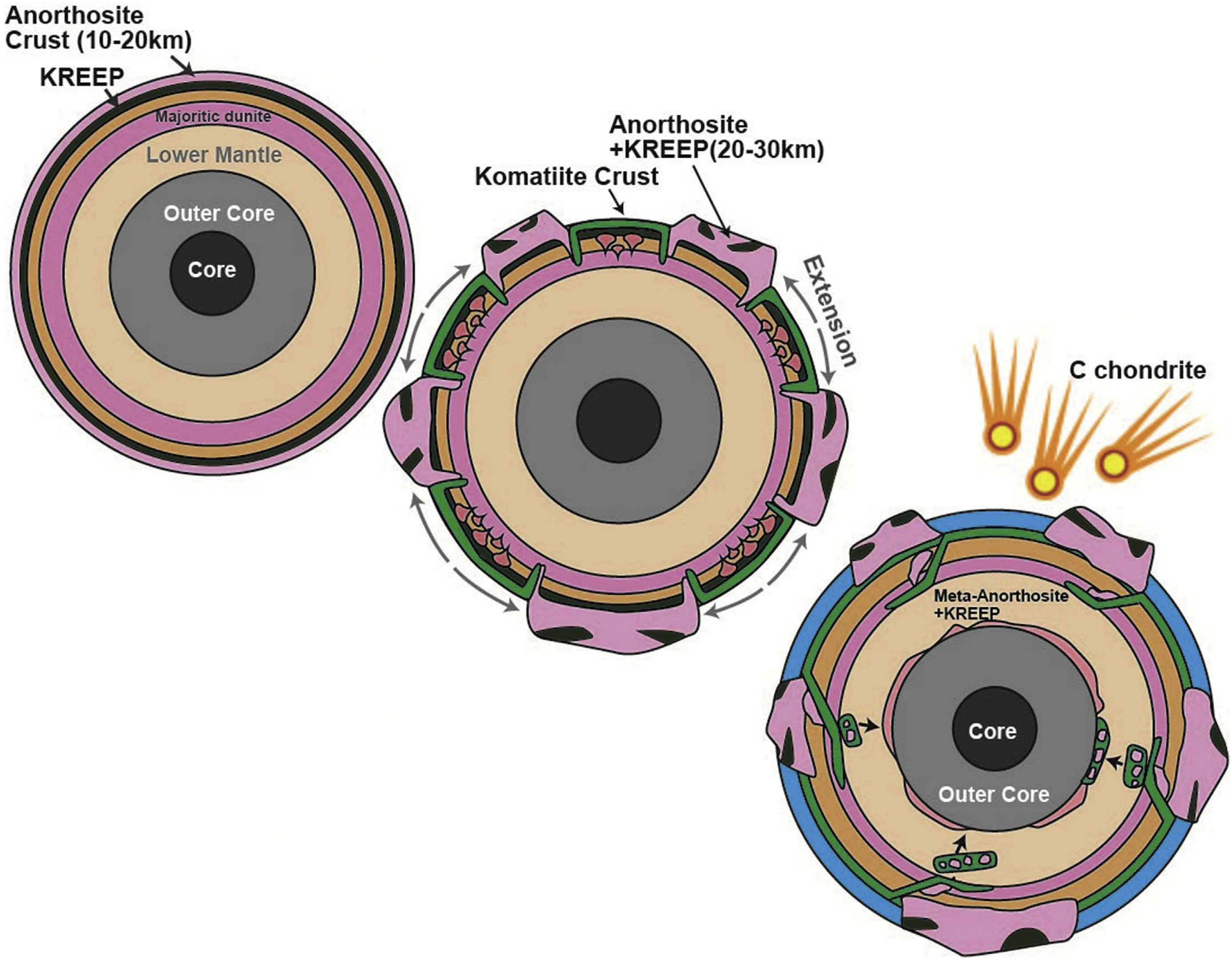


Figure 11. Stages of evolution of the early Earth from the consolidation of magma ocean through the breakup of primordial continents, and finally to the Late Heavy Bombardment stage, with subduction and annihilation of the primordial continents which were dragged down to the deep mantle. The transient period from magma ocean to plate tectonics might have been marked by bombardment tectonics which extensively destroyed the surface and also deposited a thick veneer of chondritic debris.

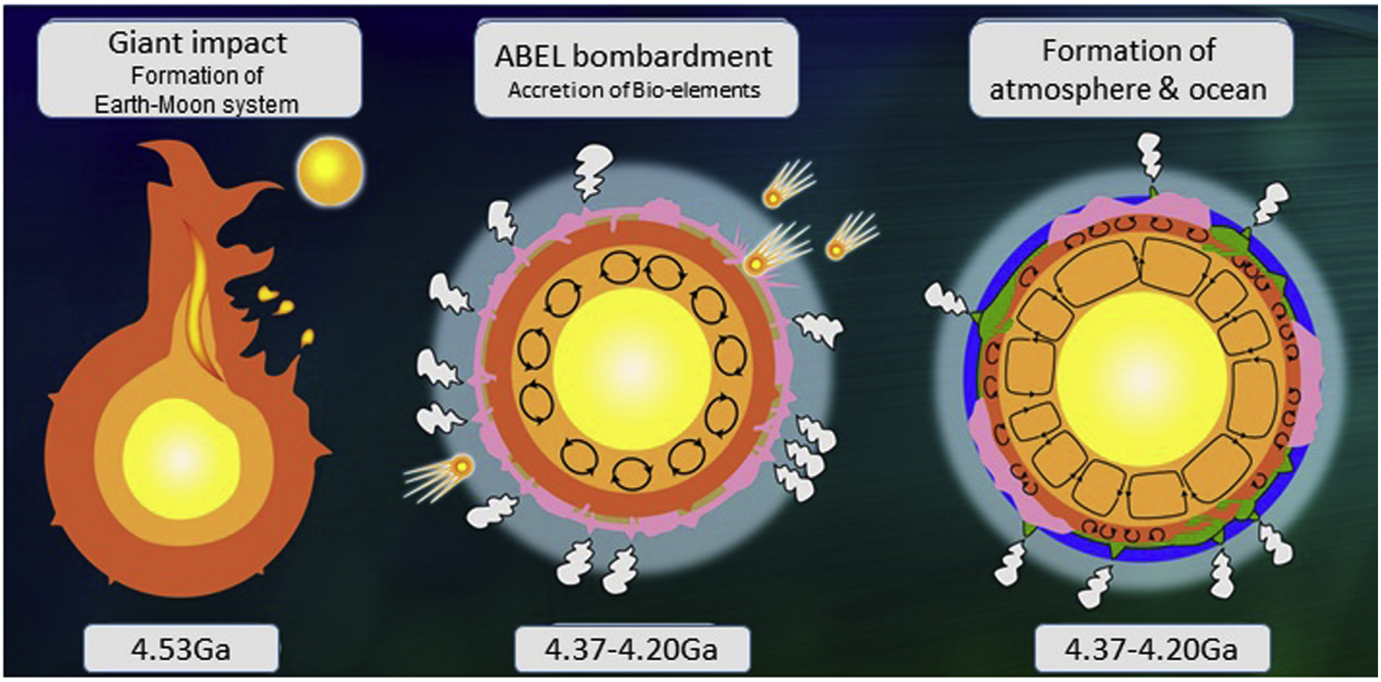


Figure 12. Two-step model of the Earth, first as a dry planet at 4.53 Ga, followed by icy asteroid bombardment at 4.37–4.20 Ga to become life-sustaining planet (Maruyama et al., 2013; Maruyama and Ebisuzaki, 2016).

accumulation as a thick (ca. 250 km) layer in the D″ layer would mean high heat input from radiogenic elements in the debris including U, Th and K bound within accessory minerals. The debris of primordial continents accumulated at the D″ layer as disrupted but gravitationally stable 'third continent' at the bottom of the mantle (the 'second continents' being in the mantle transition zone between 410 and 660 km, see Kawai et al., 2009, 2013), would contain several magnitudes higher concentration of heat producing elements distributed heterogeneously which would serve as the source for superplume and speculatively, as the major agent of core dynamics and possibly geomagnetism.

According to the pressure-dependence of density calculation of anorthosite, MORB, lherzolite, harzburgite, and granite, the primordial anorthosite must be on the bottom of mantle at 2900 km depth (Kawai et al., 2009). However, the density of all of these rocks are nearly same (4.5) at the depth range of 660–1200 km which corresponds to the topmost domain of the lower mantle. This suggests that gravitationally collapsed primordial continents would have been removed to the bottom of upper mantle and accumulated at 660 km depth. But direct removal down to CMB is difficult, and therefore it has been speculated that primordial continents were once stagnant at 660 km depth until 4.0 Ga, and were then removed by the cumulative growth of TTG (tonalite-trondhjemite-granodiorite) material from the surface brought down to depths of 660 km by plate subduction on the Hadean Earth.

Here we propose an alternative scenario that FeO-enriched KREEP lower crust performed a critical role for the descending primordial continents from the surface, straight down to the CMB, because FeO-enriched KREEP are significantly denser than the other rocks listed above. Moreover, they contain approximately 400 times more radiogenic elements such as U, Th and K as compared to the surrounding mantle (Maruyama et al., 2013). Thus, mantle dynamics must have been extensively controlled by the site and volume of KREEP rocks at CMB through time.

One of the analogs is modern Pacific superplume which was born ca. 600 Ma in the present central southern Pacific and episodically active, and still producing numerous volcanic chains (Maruyama et al., 2007; Utsunomiya et al., 2009) (Fig. 13). The site of superplume seems to be fixed over long time of over 600 Ma. The driving force of superplume has been once considered to be enriched MORB component, because the site marks the largest slab graveyards formed during the amalgamation of the supercontinent Rodinia (Maruyama et al., 2007). Because MORB is partially melted earlier than the surrounding slab peridotites, Fe-rich melt is released. At the same time, MORB-restite moves upward to form superplume enriched in Ca-perovskite (Maruyama et al., 2007).

Ishii and Tromp (1999) showed density tomography for the whole Earth scale, and suggested that the densest material, and hence gravitationally stable, exists underneath the Pacific superplume. This would KREEP-dominated mass. The Hadean primordial continents dominated by FeO enriched KREEP are enriched U, Th and K, and hence it must have worked as heat generator to produce plumes and superplumes through time. The site of the lost primordial continents may have controlled the thermal history of Earth and mantle dynamics through time.

5.4. *Stratified mantle layering and role of lost primordial continents*

5.4.1. *Olivine cumulates, very thick dunite layer with or without juvenile mantle*

As discussed earlier, mantle peridotite that formed after the solidification of magma ocean at 4.53 Ga, might have carried an unusually thick olivine cumulate layer (ca. 2000 km or even thicker) that comprises the whole lower mantle. It is transformed into perovskite + wüstite in the lower mantle condition, and partly in the mantle transition zone and thus constitutes two layers in the upper mantle. These cannot be partially melted, and act as unreactive material, because solidus of dunite (Fo = 90) is very high ca. 2000 °C at 1 bar. Solidus and liquidus of lherzolitic mantle peridotite is at 1300 °C and 1600 °C at 1 atm, and increases up to 1900 °C and to 2000 °C at 20 GPa, respectively. In the lower mantle condition, peridotite and MORB solidus is close, only 100 K

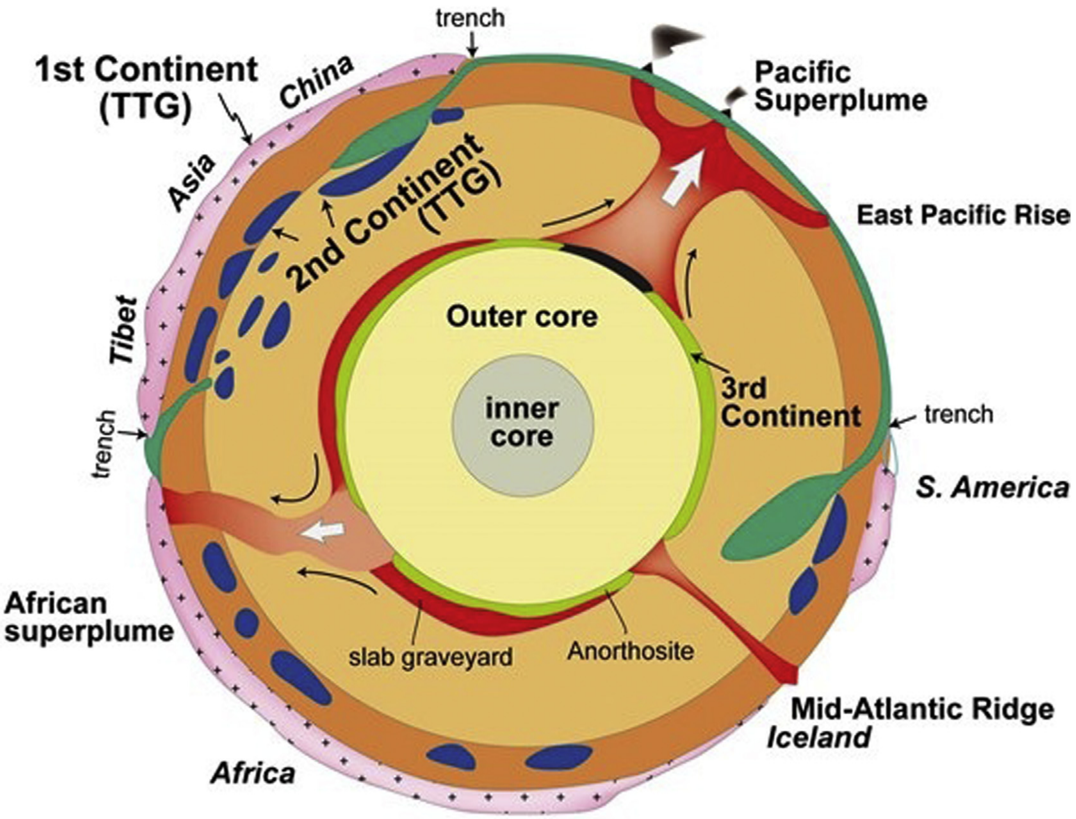


Figure 13. Schematic cross section of the Earth showing the two major superplumes (Pacific and African) and the 'second continent' dominated by TTG material in the mantle transition zone formed through subduction (after Maruyama et al., 2007; Kawai et al., 2013).

difference, reaching to 4000 K at 2900 km in depth (Hirose et al., 2006). The monomineralic dunite or perovskite + wüstite layer in the lower mantle behaves as inactive materials suggesting that the lower mantle is inactive.

If the juvenile mantle is present on the lowermost domain of the mantle, it must have reacted extensively because of chemical variants incorporated in a variety of minerals. But if the magma ocean covered the whole mantle scale, the situation is simple, and the lower mantle never reacted until plate tectonics was initiated. When plate tectonics started through the ABEL bombardment at 4.37 Ga (Maruyama and Ebisuzaki, 2016), both primordial continents including KREEP lower crust, and the powdered sediments derived from bombardments of icy asteroids with CI chondrite composition (up to 10^{22-23} kg with and ca. 20 wt.% volatiles; Maruyama and Ebisuzaki, 2016), transformed the mantle into active from the topmost domain into deeper mantle through subduction.

The upper mantle must have been active since the bombardment began, along the consuming plate boundaries. Through time, volatiles started to infiltrate into deeper mantle along the Benioff thrusts. Volatile infiltration, however, does not seem to be efficient until Neoproterozoic, ca. 800 Ma, because of the high-temperature of Earth's mantle (Maruyama and Liou, 2005; Maruyama et al., 2007).

However, the mantle overturn at 2.6–2.5 Ga (Spohn and Schubert, 1982; Tsuchiya et al., 2013) drastically changed the mantle composition for both upper and lower mantle. Since then, variable amounts of upper mantle was removed into lower mantle to be replaced by around 50% upper mantle, whereas the upper mantle was replaced completely by the inactive lower mantle in which perovskite + wüstite reacted to form olivine through decreasing pressure. If the whole upper mantle was replaced by olivine, then there is no reaction leading to a shutdown of magmatism. However, geological records do not support the idea, although magmatism decreased extensively through time. Mid-Proterozoic seems to have been least active in terms of magmatism, plate movement and metamorphism. This might reflect such mantle replacement at 2.6 Ga. Infiltration of volatiles into deep mantle started at ca. 800 Ma along the Benioff plane, although did not reach the 660 km depth boundary.

Thus the concept of original stratification of mantle layers since the solidification of magma ocean provides new insights into the mantle dynamics and history of the Earth.

5.4.2. Lost primordial continents (with KREEP rocks) and mantle dynamics through time

The primordial continents with upper felsic and lower mafic KREEP rocks have been lost from the Earth's surface by 4.0 Ga and presumably collapsed by subduction erosion and sunk into deep mantle (Fig.14). These are speculated to rest on the CMB (Core-Mantle Boundary) immediately after plate tectonics started at ca. 4.37–4.20 Ga. The upper felsic crust may have been segregated and separated from KREEP lower crust at 660 km depth. The FeO-rich lower crust may have dropped down to 660 km depth, because of its higher density than the surrounding mantle, and further down onto the CMB. If these primordial continents dominated by KREEP rocks rest above CMB, they would serve as heat engines to warm up the mantle above and cause partial melting of the underlying solid core. When the Earth was formed at 4.56 Ga, the central core may have been in solid-state. If this is true, the solid core was partially melted to initiate the dynamo only after the primordial continents were dropped down to the CMB.

Most geophysicists believe that radiogenic elements such as U, Th and K are distributed homogeneously in the whole mantle. However, we consider that the distribution of these elements are highly heterogeneous in the mantle, and nearly 400 times more concentrated within the collapsed KREEP-bearing primordial continents at CMB and since the Hadean time. If this speculation is valid, mantle dynamics and thermal history of the Earth must be reevaluated.

5.5. Role of primordial continents as a cradle of life

The rocky planets seem to be barren of life in general, because of too high temperatures (>1300 K). The accidental advent of bio-elements through impact from outer space and delivery of volatile-bearing components (Maruyama and Ebisuzaki, 2016) have facilitated the Earth to be become a life-sustaining planet. However, several additional conditions are required to give rise to the birth of life on the Earth (Maruyama et al., 2016).

One of the most important conditions is the diversified surface environments, and is possible only if the large continents are present. To realize this, the volume of ocean is critical. For example, one third of the modern Earth is covered by continents, and the remaining two third by ocean. But if the sea-level was only 9 km higher than now, no continents will be exposed above sea-level. In this case there will be no nutrient supply to evolve and sustain life. If the Earth was bombarded to add icy asteroids that would raise the sea-level by 9 km, then the Earth's life would be extinct. The first plants on landmass would also be severely damaged due to lack of photosynthesis leading to drop of oxygen level in the atmosphere. Metazoans would also perish because of the absence of free oxygen in atmosphere. There will also be no nutrient supply from landmass.

When the dry magma ocean was solidified without ocean and atmosphere, the surface environment was monotonous, similar to that in the present Moon. Facing the Sun, temperature was as high as 100 °C in daytime, and −100 °C at night, if we consider the Sun luminosity to be same as today. In fact, the young Sun during early Earth had low luminosity almost less than 70% of present day, which means that the temperatures could have been several tens of degrees lower. The surface of primordial continents was geologically highly diversified, because the final residue of magma ocean must be remarkably diversified in mineralogy as discussed in an earlier section.

Fig. 15 shows the speculated scenario of surface environment of Hadean earth immediately after "ABEL bombardment" at 4.2 Ga. The Earth was covered by 4 km thick ocean. Plate tectonics had already started. Continental rifting was followed by numerous hotspot volcanoes in the primordial continents. Surface geology was remarkably variable and included komatiitic lava flows, fractionated high-Mg andesite lava flows, weathering, erosion, transportation of sediments by river network to ocean, mountain building by continent-continent collision, Pacific-type subduction orogeny, arc subduction, active rifts, orogenic belts to expose deep crust of 4.53 Ga primordial continents such as anorthosite, komatiitic gabbros, dikes, KREEP gabbros, and extremely fractionated ore bodies such as Fe_3P, REE minerals, together with U-ores but in the absence of hydrous minerals might have prevailed. The Hadean Earth must also have been covered by river network, with several lakes, desert, glaciers on both poles, wet lands at equatorial latitudes, and evaporate formation. The dry Earth covered by the primordial continents offered the most critical nutrients such as P, K, REE and other nearly 32 elements during the solidification of magma ocean. If the surface was covered by komatiitic rocks with the same composition as primitive mantle, there will be no concentration of incompatible elements required to emerge life on the Earth. Even if "ABEL bombardment" delivered bio-elements enriched C, H, O, and N, to bear life on the planet, the fundamental factor is the diversified physical, chemical and geological features of the surface environment. The pre-biotic chemical reactions occurred under both oxidized and reduced conditions. A combination of reactions involving CO_2, H_2O, and N_2 together with several other elements supplied from the hinterland acted in conjunction to generate amino-acids, peptides, proteins, enzymes, ribozymes, and RNA, to DNA, an lead to step evolution of membranes, and coding evolution of self-replication, to finally bear the first life termed LUCA(s) or commonote(s) under the diversified environment (Kuruma, 2015).

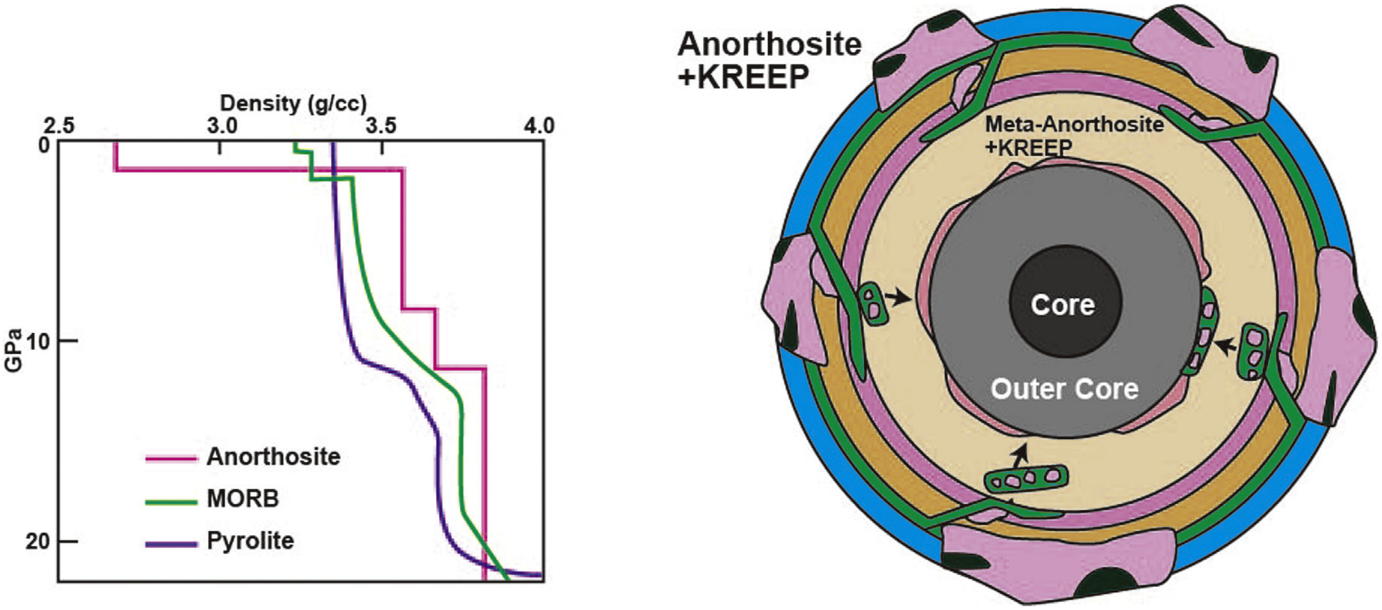


Figure 14. The process of destruction of the primordial anorthositic crust on Earth after the birth of ocean and initiation of plate tectonics. Left hand side figure shows density change versus pressure curves for anorthosite, MORB and pyrolite (after Kawai et al., 2009) suggesting that anorthositic crust becomes highly dense below 200–300 km depth and therefore sinks to the bottom of the upper mantle. Right hand side figure is a schematic illustration showing the destruction of primordial anorthositic crust by subduction-induced tectonic erosion and final removal to the core mantle boundary.

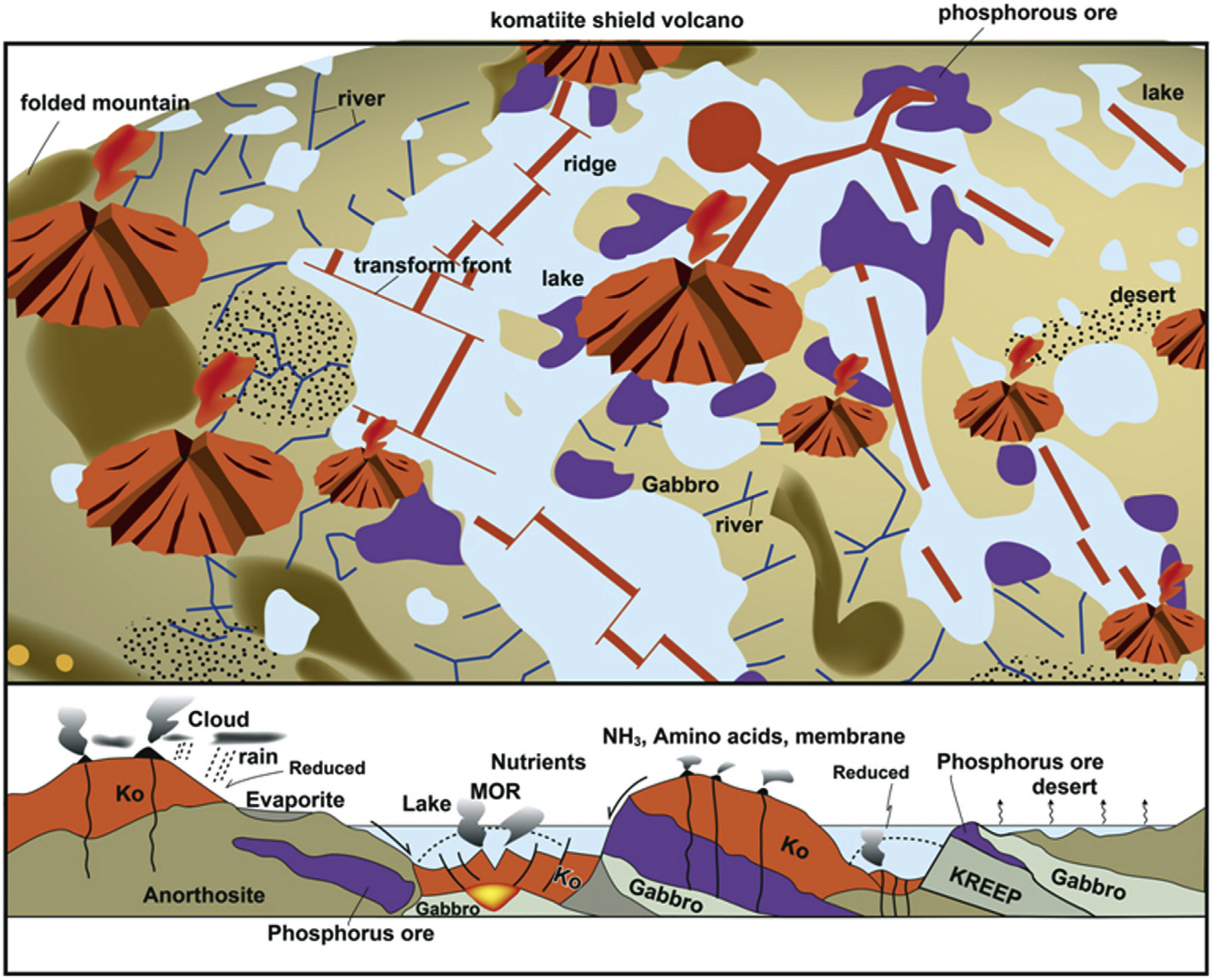


Figure 15. Speculative illustration of the surface environment of Hadean earth immediately after "ABEL bombardment" at 4.2 Ga (plan view at the top and cross section at the bottom). See text for discussion.

5.6. Origin of metabolism

Life involves a triplet of metabolism, membrane and self-replication in general. Among these, metabolism is considered to be most fundamental. Metabolic reactions are composed of chain reactions which continued over 4.0 b y. since the first life appeared on the Earth. The element P is centered in these reactions. Dry primordial continents must have been enriched in highly reduced P-bearing ores such as schreibersite (Fe_3P) without OH (Pasek, 2017). When C,H,O, and N were delivered, oxygenized agents such as H_2O, CO_2, not CO, H_2, and CH_4 started to react extensively because those two were extreme end-members in terms of $P(O_2)$.

6. Concluding remarks

If the model of CaO and Al_2O_3 rich continental crust is acceptable, the Hadean Earth might have been covered by nearly 35 km thick anorthositic continental crust. The extensive and vigorous mantle convection might have broken this apart, analogous to the disruption of supercontinents by plume and superplume activity in the younger Earth, dragging them down to the deep mantle, with accelerated destruction since the initiation of plate tectonics at ca. 4.4 Ga, and the primordial continents settled down on top of the D″ layer of the Core-Mantle Boundary.

Thus, the major stages of evolution of the early Earth included (a) consolidation of magma ocean, (b) breakup of primordial continents, (c) Late Heavy Bombardment ("ABEL bombardment"), and (d) subduction and annihilation of the primordial continents which were dragged down into the deep mantle. The transient period from magma ocean to plate tectonics might have been marked by bombardment tectonics which extensively destroyed the surface and also deposited a thick veneer of chondritic debris, identical to thick sedimentation in modern orogenic belts. The record of the impact of large asteroidal and cometary materials on the Moon's surface is preserved in the form of regoliths with thickness varying from 2 m to 20 m carrying pulverized material which is a mixture of rock fragments from the primitive lunar crust together with glassy particles from the impact. Due to the small size of Moon, the volatiles delivered by the chondritic material escaped whereas in the case of Earth, the volatiles and ultra-high temperatures might have caused extensive mineral transformations in the underlying anorthositic crust to make it denser and to be dragged down into the deep mantle.

The dry Earth covered by the primordial continents offered the critical nutrients such as P, K, REE and several other elements during the solidification of magma ocean. If the surface was covered by komatiitic rocks with the same composition as primitive mantle, there will be no concentration of incompatible elements required to emerge life on the Earth. Possible phosphorous ores were formed by fractionation of KREEP basalts. Although the early ocean was highly toxic, clean water became available in isolated inland lakes and possible transform fault-like features are also speculated. With nutrient supply from the landmass, the cradle of prebiotic life in Earth might have been created. "ABEL bombardment" delivered bio-elements enriched C, H, O, and N. Together with CO_2, H_2O, N_2 and several other elements supplied from the hinterland, complex reactions proceeded to generate amino-acids, peptides, proteins, enzymes, ribozymes, and RNA to DNA, marking the dawn of early life on Earth.

Acknowledgments

We thank Editor Prof. Sanghoon Kwon and two anonymous referees for helpful comments. Santosh is supported by Foreign Expert Funding from China University of Geosciences Beijing and Professorial position at the University of Adelaide, Australia.

References

Albarede, F., 2009. Volatile accretion history of the terrestrial planets and dynamic implications. Nature 461, 1227–1233.

Borg, L.E., Gaffney, A.M., Shearer, C.K., 2015. A review of lunar chronology revealing a preponderance of 4.34–4.37 Ga ages. Meteoritics and Planetary Science 50, 715–732.

Bradley, L.J., Gills, J.J., Haskin, L.A., Korotev, R.L., Wieczorek, M.A., 2000. Major lunar crustal terranes: surface expressions and crust-mantle origins. Journal Geophysical Research 105, 4197–4216.

Canup, R.M., Asphaug, E., 2001. Origin of the Moon in a giant impact near the end of the Earth's formation. Nature 412, 708–712.

Carter, J., Poulet, F., 2013. Ancient plutonic processes on Mars inferred from the detection of possible anorthositic terrains. Nature Geosciences 6, 1008–1012.

Condie, K.C., Kröner, A., 2008. When did plate tectonics begin? Evidence from the geologic record. Geological Society of America Special Papers 440, 281–294.

Dohm, J.M., Maruyama, S., 2015. Habitable trinity. Geoscience Frontiers 6, 95–101.

Dollfus, A., 2000. Langrenus: transient illuminations on the Moon. Icarus 146, 430–443.

Elkins-Tanton, L.T., 2008. Linked magma ocean solidification and atmospheric growth for Earth and Mars. Earth and Planetary Science Letters 271, 181–191.

Elkins-Tanton, L.T., Burgess, S., Yin, Q.Z., 2011. The lunar magma ocean: reconciling the solidification process with lunar petrology and geochronology. Earth and Planetary Science Letters 304, 326–336.

Floran, R.J., Dence, M.R., 1976. Morphology of the Manicouagan ring-structure, Quebec, and some comparisons with lunar basins and craters. In: Proceedings 7th Lunar Science Conference, Lunar and Planetary Science Institute, NASA, pp. 2845–2865.

Fujii, T., Kushiro, I., 1977. Melting relations and viscosity of abyssal tholeiite. Carnegie Institute Washington Yearbook 76, 461–465.

Ghiorso, M.S., Hirschmann, M.M., Reiners, P.W., Kress, V.C., 2002. The pMELTS: a revision of MELTS for improved calculation of phase relations and major element partitioning related to partial melting of the mantle to 3 GPa. Geochemistry, Geophysics, Geosystems 3, 1–35.

Gualda, G.A.R., Ghiorso, M.S., Lemons, R.V., Carley, T.L., 2012. Rhyolite-MELTS: a modified calibration of MELTS optimized for silica-rich, fluid-bearing magmatic systems. Journal of Petrology 53, 875–890.

Herzberg, C.T., 1983. The solidus and liquidus temperatures and mineralogy for anhydrous garnet lherzolite to 15 GPa. Physics of the Earth and Planetary Interior 32, 193–202.

Herzberg, C., 1992. Depth and degree of melting of komatiites. Journal of Geophysical Research: Solid Earth 97, 4521–4540.

Hirose, K., Sinmyo, R., Sata, N., 2006. Determination of post-perovskite phase transition boundary in $MgSiO_3$ using Au and MgO pressure standards. Geophysical Research Letters 33, L01310. http://dx.doi.org/10.1029/2005GL024468.

Hopkins, M.D., Mojzsis, S.J., 2015. A protracted timeline for lunar bombardment from mineral chemistry, Ti thermometry and U-Pb geochronology of Apollo 14 melt breccia zones. Contributions to Mineralogy and Petrology 169, 30–36. http://dx.doi.org/10.1007/s00410-015-1123-x.

Hui, H., Peslier, A.H., Zhang, Y., Neal, C.R., 2013. Water in lunar anorthosites and evidence for a wet early Moon. Nature Geoscience 6, 177–180.

Imaeda, Y., Ebisuzaki, T., 2017. Tandem planet formation for solar system-like planetary systems. Geoscience Frontiers 8, 223–231.

Ishii, M., Tromp, J., 1999. Normal-mode and free-air gravity constraints on lateral variations in velocity and density of Earth's mantle. Science 285, 1231–1236.

Ito, E., Takahashi, E., 1987. Ultrahigh-pressure phase transformations and the constitution of the deep mantle. In: Manghnani, M.H., Syono, Y. (Eds.), High-pressure Research in Mineral Physics: a Volume in Honor of Syun-iti Akimoto. American Geophysical Union, Washington, D. C.. http://dx.doi.org/10.1029/GM039p0221

Javoy, M., 1995. The integral enstatite chondrite model of the Earth. Geophysical Research Letters 22, 2219–2222.

Kato, T., Ringwood, A.E., Irifune, T., 1988. Experimental determination of element partitioning between silicate perovskites, garnets and liquids: constraints on early differentiation of the mantle. Earth and Planetary Science Letters 89, 123–145.

Kawai, K., Tsuchiya, T., Tsuchiya, J., 2009. Lost primordial continents. Gondwana Research 16, 581–586.

Kawai, K., Yamamoto, S., Tsuchiya, T., Maruyama, S., 2013. The second continent: existence of granitic continental materials around the bottom of the mantle transition zone. Geoscience Frontiers 4, 1–6.

Kemp, A.I.S., Wilde, S.A., Hawkesworth, C.J., Coath, C.D., Nemchin, A., Pidgeon, R.T., Vervoot, J.D., DuFrance, S.A., 2010. Hadean crustal evolution revisited: new constraints from Pb–Hf isotope systematics of the Jack Hills zircons. Earth and Planetary Science Letters 296, 45–56.

Komabayashi, T., Maruyama, S., Rino, S., 2009. A speculation on the structure of the D″ layer: the growth of anti-crust at the core–mantle boundary through the subduction history of the Earth. Gondwana Research 15, 342–353.

Komiya, T., Maruyama, S., Masuda, T., Nohda, S., Hayashi, M., Okamato, K., 1999. Plate tectonics at 3.8–3.7 Ga: field evidence from the Isua accretionary complex, southern West Greenland. Journal of Geology 107, 515–554.

Kranendonk, V.M.J., Smithies, R., Hickman, A.H., Champion, D.C., 2007. Secular tectonic evolution of Archean continental crust: interplay between horizontal and vertical processes in the formation of the Pilbara Craton, Australia. Terra Nova 19, 1–38.

Kuruma, Y., 2015. Creation of simple biochemical systems to study early cellular life. Origins of Life and Evolution of Biospheres 45, 359–360.

Kushiro, I., Fujii, T., 1977. Floatation of plagioclase in magma at high pressures and its bearing on the origin of anorthosite. Proceedings of the Japan Academy, Series B 53, 262–266.

Lange, R.A., Carmichael, I.S.E., 1987. Densities of Na_2O-K_2O-MgO-MgO-FeO-Fe_2O_3-Al_2O_3-TiO_2-SiO_2 liquids: new measurements and derived partial molar properties. Geochimica et Cosmochimica Acta 51, 2931–2946.

Lucey, P.G., Taylor, G.J., Malaret, E., 1995. Abundance and distribution of iron on the Moon. Science 268, 1150–1153.

Maruyama, S., Ebisuzaki, T., 2016. Origin of the Earth: a proposal of new model called ABEL. Geoscience Frontiers (under revision).

Maruyama, S., Liou, J.G., 2005. From snowball to Phaneorozic Earth. International Geology Review 47, 775–791.

Maruyama, S., Santosh, M., Zhao, D., 2007. Superplume, supercontinent, and post-perovskite: mantle dynamics and anti-plate tectonics on the Core–Mantle Boundary. Gondwana Research 11, 7–37.

Maruyama, S., Ikoma, M., Genda, H., Hirose, K., Yokoyama, T., Santosh, M., 2013. The naked planet Earth: most essential pre-requisite for the origin and evolution of life. Geoscience Frontiers 4, 141–165.

Maruyama, S., Santosh, M., Sawaki, Y., Suda, K., 2016. Birth place of early life on Earth: not mid-oceanic ridge, but geyser driven by natural nuclear reactor. Geoscience Frontiers (submitted for publication).

Melosh, 1989. Impact Cratering: a Geologic Process. Calredon Press, Oxford, 245pp.

Nance, R.D., Murphy, J.B., Santosh, M., 2014. The supercontinent cycle: a retrospective essay. Gondwana Research 25, 4–29.

Ohta, H., Maruyama, S., Takahashi, E., Watanabe, Y., 1996. Field occurrence, geochemistry and petrogenesis of the Archean mid-oceanic ridge basalts (AMORBs) of the Cleaverville area, Pilbara craton, Western Australia. Lithos 37, 199–221.

Ohtake, M., Matsunaga, T., Haruyama, J., et al., 2009. The global distribution of pure anorthosite on the Moon. Nature 461, 236–240.

Ohtani, E., 1985. The primordial terrestrial magma ocean and its implication for stratification of the mantle. Physics of the Earth and Planetary Interior 38, 70–80.

Pasek, M.A., 2017. Schreibersite on the early earth: scenarios for prebiotic phosphorylation. Geoscience Frontiers 8, 329–335.

Pillinger, C.T., Greenwood, R.C., Johnson, D., Gibbson, E.K., Miller, M.F., Franchi, L.A., 2014. Three isotopes of oxygen in lunar samples – the same as Earth or different?. In: Lunar and Planetary Science Conference, NASA, March 2014 (abstract).

Pirajno, F., 2015. Intracontinental anorogenic alkaline magmatism and carbonatites, associated mineral systems and the mantle plume connection. Gondwana Research 27, 1181–1216.

Ram Mohan, M., Satyanarayanan, M., Santosh, M., Sylvester, P.J., Tubrett, M., Lam, R., 2013. Neoarchean suprasubduction zone arc magmatism in southern India: Geochemistry, zircon U-Pb geochronology and Hf isotopes of the Sittampundi Anorthosite Complex. Gondwana Research 23, 539–557.

Ringwood, A.E., 1990. Earliest History of the Earth-moon System. Origin of the Earth, Harvard University.

Roedder, E., Weiblen, P.W., 1970. Lunar petrology of silicate melt inclusions, Apollo 11 rocks. In: Levinson, A.A. (Ed.), Geochimica et Cosmochimica Acta Supplement, Proceedings of the Apollo 11 Lunar Science Conference, Houston, TX. Volume 1: Mineraolgy and Petrology. Pergammon Press, New York, p. 801.

Saal, A.E., Hauri, E.H., Van Orman, J.A., Rutherford, M.J., 2013. Hydrogen isotopes in lunar volcanic glasses and melt inclusions reveal a carbonaceous chondrite heritage. Science 340, 1317–1320.

Santosh, M., 2010. A synopsis of recent conceptual models on supercontinent tectonics in relation to mantle dynamics, life evolution and surface environment. Journal of Geodynamics 50, 116–133.

Santosh, M., Maruyama, S., Yamamoto, S., 2009. The making and breaking of supercontinents: some speculations based on superplumes, super downwelling and the role of tectosphere. Gondwana Research 15, 324–341.

Santosh, M., Yang, Q.Y., Shaji, E., Tsunogae, T., Ram Mohan, M., Satyanarayanan, M., 2015a. An exotic Mesoarchean microcontinent: the Coorg Block, southern India. Gondwana Research 27, 165–195.

Santosh, M., Teng, X.M., He, X.F., Tang, L., Yang, Q.Y., 2015b. Discovery of Neoarchean suprasubduction zone ophiolite suite from Yishui Complex in the North China Craton. Gondwana Research. http://dx.doi.org/10.1016/j.gr.2015.10.017.

Shearer, C.K., Papike, J.J., 2005. Crustal building processes on the Moon: models for the petrogenesis of the magnesian suite. Geochimica et Cosmochimica Acta 69, 3445–3461.

Shearer, C.K., Hess, P.C., Weisckorek, M.A., Pritchard, M.E., Parmentier, E.M., Borg, L.E., Longhi, J., Elkins-Tanton, L., Neal, C.R., Antonenko, I., Canup, R.I., Halliday, A.N., Grove, T.L., Hager, B.H., Lee, D.-C., Wiechert, U., 2006. Magmatic and thermal history of the Moon. In: Wieczorek, J.B., Sheraer, C.K., Neal, C. (Eds.), New Views of the Moon, Reviews in Mineralogy, vol. 60. Mineralogical Society of America, Washington D.C., pp. 365–518

Sleep, N.H., 2010. The Hadean-Archaean environment. Cold Spring Harbor Perspectives in Biology. http://dx.doi.org/10.1101/cshperspect.a002527.

Smithies, R.H., Van Kranendonk, M.J., Champion, D.C., 2007. The Mesoarchean emergence of modern-style subduction. Gondwana Research 11, 50–68.

Spohn, T., Schubert, G., 1982. Modes of mantle convection and the removal of heat from the Earth interior. Journal Geophysical Research 87, 4682–4696.

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

Stern, C.R., 2011. Subduction erosion: rates, mechanisms, and its role in arc magmatism and the evolution of the continental crust and mantle. Gondwana Research 20, 284–308.

Synder, G.A., Taylor, L.A., Neal, C.R., 1992. A chemical model for generating the sources of mare basalts: combined equilibrium and fractional crystallization of the lunar magmasphere. Geochimica et Cosmochimica Acta 56, 3809–3823.

Teng, X.M., Santosh, M., 2015. A long-lived magma chamber in the Paleoproterozoic North China Craton: evidence from the Damiao gabbro-anorthosite suite. Precambrian Research 256, 79–101.

Trail, D., Watson, E.B., Tailby, N.D., 2011. The oxidation state of Hadean magmas and implications for early Earth's atmosphere. Nature 480, 79–82.

Tsuchiya, T., Kawai, K., Maruyama, S., 2013. Expanding-contracting earth. Geoscience Frontiers 3, 341–347.

Utsunomiya, A., Suzuki, N., Ota, T., 2009. Preserved paleo-oceanic plateaus in accretionary complexes: implications for the contributions of the Pacific superplume to global environmental change. Gondwana Research 14, 115–125.

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

Wanke, H., Dreibus, G., 1982. Chemical and Isotopic Evidence for the Early History of the Earth-moon System. Tidal Friction and the Earth's Rotation II. Springer-Verlag, Berlin.

Warren, P.H., Wasson, J.T., 1979. The origin of KREEP. Reviews of Geophysics 17, 73–88.

Wilde, S.A., Valley, J.W., Peck, W.H., Graham, C.M., 2001. Evidence from detrital zircons for the existence of continental crust and oceans on the Earth 4.4 Gyr ago. Nature 409, 175–178.

Windley, B.F., 1984. The Evolving Continents. John Wiley and Sons, New York, 416 pp.

Wood, J.A., Dickey, J.S., Marvin, U.B., Powell, B.N., 1970. Lunar anorthosites and geophysical model of the Moon. In: Proceedings, Apollo 11 Lunar Science Conference, pp. 965–988.

Yamamoto, S., Senshu, H., Rino, S., Omori, S., Maruyama, S., 2009. Granite subduction: arc subduction, tectonic erosion and sediment subduction. Gondwana Research 15, 443–453.

Yang, Q.Y., Santosh, M., Collins, A.S., Teng, X.M., 2016. Microblock amalgamation in the North China Craton: evidence from Neoarchaean magmatic suite in the western margin of the Jiaoliao block. Gondwana Research 31, 96–123.

Young, E.D., Kohl, I.E., Warren, P.H., Rubie, D.C., Jacobsen, S.A., Morbidelli, A., 2016. Oxygen isotopic evidence for vigorous mixing during the Moon-forming giant impact. Science 351, 491–496.

Zahnle, K., Arndt, N., Cockell, C., Halliday, A., Nisbet, E., Selsis, F., Sleep, N.H., 2007. Emergence of a habitable planet. Space Science Reviews 129, 35–78.

Zhai, M.G., Santosh, M., 2011. The early Precambrian odyssey of the North China Craton: a synoptic overview. Gondwana Research 20, 6–25.

Zhao, D., Arai, T., Liu, L., Ohtani, E., 2012. Seismic tomography and geochemical evidence for lunar mantle heterogeneity: comparing with Earth. Global and Planetary Change 90–91, 29–36.