

Magma Oceans in the Inner Solar System

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Annu. Rev. Earth Planet. Sci. 2012. 40:113–39

First published online as a Review in Advance on
January 3, 2012

The *Annual Review of Earth and Planetary Sciences* is
online at earth.annualreviews.org

This article's doi:
10.1146/annurev-earth-042711-105503

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0084-6597/12/0530-0113\$20.00

Keywords

accretion, planetesimal, differentiation, volatile, crust formation,
atmospheric degassing

Abstract

Theory and observations point to the occurrence of magma ponds or oceans in the early evolution of terrestrial planets and in many early-accreting planetesimals. The apparent ubiquity of melting during giant accretionary impacts suggests that silicate and metallic material may be processed through multiple magma oceans before reaching solidity in a planet. The processes of magma ocean formation and solidification, therefore, strongly influence the earliest compositional differentiation and volatile content of the terrestrial planets, and they form the starting point for cooling to clement, habitable conditions and for the onset of thermally driven mantle convection and plate tectonics. This review focuses on evidence for magma oceans on planetesimals and planets and on research concerning the processes of compositional differentiation in the silicate magma ocean, distribution and degassing of volatiles, and cooling.

1. INTRODUCTION

The silicate and metal material making up the terrestrial planets was likely melted one or more times during accretionary processes before reaching its final solid state in the young terrestrial planets as evidenced by early melting in planetesimals (e.g., Urey 1955, Fish et al. 1960, Lee et al. 1976) and by the energetics of accretionary impacts (e.g., Safronov 1969, Wetherill 1980, Tonks & Melosh 1993). This processing through magma oceans created compositional differentiation and affected the final volatile contents of these bodies.

Planetesimals (whose radii range from tens to hundreds of kilometers) that accreted within the first ~ 1.5 million years (Ma) following the first solids in the solar system would likely have experienced significant and perhaps complete melting owing to radiogenic heating from short-lived radioisotopes (e.g., Urey 1955, Fish et al. 1960, Lee et al. 1976, LaTourrette & Wasserburg 1998). After approximately the first 3 Ma after formation of the first solids, ^{26}Al , the most active short-lived heat source for melting rocky material, was extinct.

Similar extents of melting are subsequently expected in the large terrestrial planets owing to the conversion of kinetic energy to heat during energetic accretionary impacts with planetary embryos (e.g., Safronov 1969, Wetherill 1980, Tonks & Melosh 1993, Halliday et al. 2001). Giant accretionary impacts likely occurred over the next 100 Ma (and still could, although the probability is very low). Thus, it is widely assumed that magma oceans are significant events in the earliest stages of planetary evolution and are thus highly influential in setting the initial compositional differentiation and structure of planets.

Long before short-lived radioisotopes were discovered and the energetics of giant impacts investigated, philosophers and scientists suggested that the planets were originally molten. As early as 1694, Gottfried Wilhelm Leibniz postulated in his book *Protogaea* that Earth began as a uniform liquid and differentiated compositionally as it cooled. In 1748, Georges-Louis Leclerc, Comte de Buffon, began publishing his 36-volume *Histoire Naturelle, Générale et Particulière*, in which he hypothesized that Earth and other planets began in the molten state after being fragmented from the Sun. He further calculated that cooling Earth to its current temperature would require 75,000 years (Svante Arrhenius wrote in 1909, commenting on Buffon's work, that Earth's interior was likely still incandescent with its original heat). Although the hypothesized sources of the heat and the spans of cooling time have changed over the centuries, the hypothesis of an originally molten body that cools and differentiates persists.

Taylor & Norman (1992) suggest that a magma ocean can be defined by two criteria. First, the magma behaves rheologically as a liquid, having a sufficiently small crystal fraction that the crystals are suspended within the liquid and not fused into a network. Second, the magma encompasses a substantial fraction of the body, perhaps more than 10% (equivalent to a 60-km-deep surface magma ocean on the Moon, or 230-km deep on Earth). Here I specifically address not only surface magma oceans but also interior magma oceans produced on planetesimals by radiogenic heat from short-lived isotopes, and possibly at depth in larger planets through processes discussed below.

The processes of solidification of the magma ocean determine initial compositional differentiation of the silicate portion of the planet. The subsequent stability of this differentiation has significance for magmatic source regions, convective instability, and magnetic field generation. This review summarizes the existing observational evidence for magma oceans on planetesimals and planets and the processes believed responsible for their formation; discusses outstanding problems in magma ocean fluid dynamics [Solomatov (2000) gives an excellent summary of the state of knowledge]; and covers the compositional differentiation in planets produced by magma ocean processes. These processes include core formation; major-, trace-, and volatile-element differentiation within the silicate mantle and crust; and atmospheric formation.

2. CHEMICAL EVIDENCE FOR MAGMA OCEANS

Evidence for magma oceans in the inner solar system falls broadly into four categories. First, some compositions and mineralogies of samples indicate early widespread silicate melting and subsequent fractional crystallization; this is the case for the Moon, Mars, and Vesta. Second, the crystallization ages of samples may verge on the age range of primary planetary formation, ~ 4.4 billion years ago (Gya) and before, indicating a compressed timescale for accretion and therefore a greater chance that accretionary and radiogenic heat would produce melting rather than be radiated to space. Samples from the Moon, Vesta, and Earth [in the form of ancient zircons; see, e.g., Valley et al. (2002) and Harrison (2009)] may indicate age of accretion. Third, ^{182}W and ^{142}Nd contents in crustal and mantle rocks can suggest early core formation, also requiring a short accretionary timescale; this may be particularly true for Mars.

Fourth, the siderophile element content of a planet's mantle is thought to be determined by a combination of conditions of core formation and accretion following core formation. Core formation in a magma ocean environment provides an opportunity for equilibrium element partitioning between metal and silicate. Ringwood (1966) suggested for Earth that the siderophile elements V, Cr, and Mn were incorporated into the core in reduced form at the bottom of a magma ocean, thus setting their mantle abundances. This model requires that metal-silicate equilibration is reached at a small range of pressures at the bottom of a liquid magma ocean overlying solid silicates and possibly preexisting core material. Because partitioning behavior is dependent on pressure and temperature, the depth of the magma ocean would determine the elements' mantle abundances (Righter et al. 1997, Righter & Drake 1999, Walker 2009, Righter & Chabot 2011). This model has been applied in particular to Earth and Mars.

2.1. Moon

Samples returned from the Apollo missions were the first to provoke the hypothesis of a magma ocean in its modern form. The discovery of anorthosites, rocks consisting almost entirely of the relatively low-density aluminosilicate mineral anorthite (a variety of plagioclase feldspar), inspired Smith et al. (1970) and Wood et al. (1970) to suggest that they formed through flotation to the top of a magma sea on the early Moon; no other mechanism for the formation of these rocks has attained general acceptance (but see Boyet & Carlson 2007 and Longhi 2003). The lunar highlands consist of more than three-quarters anorthite by volume.

Other major suites of rocks and their implied source regions in the Moon, including the picritic glasses and mare basalts and the KREEP (K for potassium, REE for rare-earth elements, and P for phosphorus) component of enriched incompatible elements, are consistent with fractional solidification of a magma ocean on the Moon that was originally hundreds of kilometers deep. A key piece of evidence is the abundance of the trace element europium. The anorthitic crust appears to have scavenged europium from the magma ocean, leaving matching europium deficits in the mare basalt source region; the complementary compositions require solidification from a common magmatic progenitor (Philpotts & Schnetzler 1970, Wakita & Schmitt 1970).

Furthermore, lunar crustal rocks have ages that are determined by radioactive isotopic systems to be extremely old: Many of these rocks formed within the first 200 Ma after the formation of the first solids in the solar system (Papanastassiou & Wasserburg 1971 and many subsequent publications; see summary in Borg et al. 1999). Thus, evidence that there was a lunar magma ocean and that it solidified fractionally rather than in bulk comes from the anorthosite flotation crust, ages of crustal rocks, and trace elements in several suites of rocks, including the KREEP component that is thought to have been produced by high degrees of fractional solidification.

Further reviews of observations and hypotheses concerning the lunar magma ocean can be found in Warren (1985) and Shearer et al. (2006). Thus, although the magma ocean hypothesis remains strong for the Moon, questions remain about depths and extents of magma oceans on other rocky bodies and whether their existence can be detected in observations today.

2.2. Mars

Mars likely formed before Earth and the Moon were in their current states (Kleine et al. 2004). Early core formation would result in accretionary heat being concentrated over a brief time period, thus giving greater melting potential. Core formation itself would be enhanced by the presence of a magma ocean. Measurements of ^{182}W and ^{142}Nd in Martian meteorites yield model ages suggesting that the core and mantle fractionated within 13 ± 2 Ma of the origin of the solar system (Lee & Halliday 1995, Blichert-Toft et al. 1999, Kleine et al. 2002, Foley et al. 2005). Shih et al. (1999) concluded that Mars differentiation had occurred within 10 Ma following accretion.

Foley et al. (2005) further proposed that the Martian mantle differentiated into source reservoirs for the shergottite and nakhlite meteorite compositions within ~ 60 Ma (and perhaps as little as 20 Ma) of the first solar system solids. Marty & Marti (2002) found that open-system degassing of a magma ocean within 30 Ma of the first solar system solids was consistent with noble gas isotopic data. In contrast to these short estimates for magma ocean duration, both Blichert-Toft et al. (1999) and Debaille et al. (2007) argue for silicate differentiation (possibly in a Martian magma ocean) that lasted ~ 100 Ma after core formation.

Siderophile-element abundances and silicate partitioning data have been used to model a range of possible Martian magma ocean depths: 700–800 km (Richter et al. 1998), $\sim 1,100 \pm 300$ km (Richter & Chabot 2011), and ~ 400 km (Agee & Draper 2004). This model may be too simple: Physical and chemical arguments for metal-silicate equilibrium in droplets raining through the magma ocean appear convincing (Li & Agee 2001, Rubie et al. 2003), and equilibration may therefore occur over a wide range of pressures. Indeed, disequilibrium is also likely if there exist larger droplets or unemulsified blocks, which would sink faster. Furthermore, multiple magma oceans of varying depths must have occurred during the process of accretion, so mantle abundances are the cumulative result of complex processes (Rubie et al. 2011).

Thus, Mars appears to be a very early-formed planet (or surviving planetary embryo) that experienced its last magma ocean only a few tens of millions of years into solar system history. The preservation of whole-rock Rb-Sr isochrons and ^{182}W and ^{142}Nd anomalies over the age of the solar system is also evidence that our exceedingly small inventory of Martian samples experienced little or no crustal recycling since magma ocean solidification, implying insufficient mantle convection to homogenize the mantle.

2.3. Vesta and Other Small Bodies

Many chondrules and all CAIs (calcium-aluminum inclusions, the oldest solids in the solar system and components of many chondritic meteorites) have ages that fall within the first few million years after solids began to form in the solar system, and planetesimals appear to have been forming at the same time (see list of references in Elkins-Tanton et al. 2011b). The class of meteorites termed achondrites is characterized by members whose textures and compositions indicate that they were produced from melting in their parent body, thought in most cases to be an early-forming planetesimal (Greenwood et al. 2005). A subset of achondrites, the howardites, eucrites, and diogenites (HEDs), is inferred to have originated on the asteroid 4 Vesta (Binzel & Xu 1993). If that inference is true, then isotopic evidence from HED meteorites indicates that igneous rocks

existed on Vesta within 10 Ma of formation of the first solids in the solar system (Papanastassiou & Wasserburg 1969).

The ages of iron meteorites may be the most compelling evidence for widespread magma oceans in planetesimals. Iron meteorites are thought to be the cores of destroyed planetesimals, so they imply planetesimal formation, melting, and differentiation before breakup. Compositionally, they result from fractional solidification of a liquid iron (Scott 1972), presumably the core of a planetesimal. Their isotopic composition indicates that a majority of them solidified less than 1 Ma after the first solids in the solar system formed Scherstén et al. (2006), indicating that scores of bodies accreted quickly, melted internally, and differentiated.

2.4. Earth

Billions of years of plate tectonics and mantle convection have mixed Earth's mantle to a high degree and have recycled the vast majority of whatever earliest crust existed. The hallmarks of the magma ocean on the Moon—a plagioclase flotation crust, a high-titanium component, and KREEP—are not preserved on Earth, and the extremity of isotopic heterogeneity found in the Martian mantle is not preserved, either. Theory suggests that the silicate Earth underwent partial if not total melting (Abe & Matsui 1985, Stevenson 1987, Ahrens 1990, Sasaki 1990, Wetherill 1990, Tonks & Melosh 1993, Canup 2008), but geochemical evidence requiring such a scenario has not been identified.

Perhaps the most extensive effort in understanding a terrestrial magma ocean(s) to date has been made by groups proposing that the siderophile-element composition in Earth's mantle is set during core segregation at high pressure and temperature at the bottom of a deep magma ocean, as discussed above in Section 2.2. Both Li & Agee (1996) and Righter et al. (1997) recognized that the mantle siderophile-element budget is consistent with core segregation at approximately 27 GPa and $\sim 2,000^\circ\text{C}$. Righter & Drake (1999) found that these data are also consistent with the magma ocean containing water rather than being anhydrous. Work in this field is ongoing; a summary can be found in Walter et al. (2000).

Furthermore, Earth's $^{142}\text{Nd}/^{144}\text{Nd}$ composition as measured in crustal igneous rocks differs from that of chondritic meteorites (Boyet & Carlson 2005, Caro et al. 2006). Thus, either Earth's mantle has a nonchondritic composition or there exists an untapped reservoir of mantle material with a $^{142}\text{Nd}/^{144}\text{Nd}$ composition that mass-balances with the upper mantle compositions measured by igneous rocks to produce a net chondritic composition (Carlson & Boyet 2008). Carlson & Boyet (2008) and Caro et al. (2005, 2006) argue for silicate differentiation in a terrestrial magma ocean to produce such material through fractionation.

3. HEAT SOURCES FOR MAGMA OCEANS

3.1. Melting Planetesimals with Heat from Radioactive Decay

As early as 1955, Urey recognized the importance of ^{26}Al radiogenic heating early in the solar system because of its short half-life and relatively high abundance. The discovery of evidence for this short-lived isotope in ancient solar system materials came from the existence of its daughter product, ^{26}Mg (Gray & Compston 1974, Lee et al. 1976), in crystal sites overwhelmingly preferred by aluminum. Less important extinct radioactive isotopes from the point of view of heat generation include ^{60}Fe , ^{146}Sm , ^{244}Pu , and ^{182}Hf ; in addition, the most important producers of radiogenic heat today— ^{238}U , ^{235}U , ^{40}K , and ^{232}Th —were more abundant early in the solar system. The isotope ^{26}Al , however, largely dominates the heat budget for silicate materials and thus determines via its

abundance and half-life that the opportunity for melting planetesimals is limited to bodies that accreted within ~ 2 Ma after formation of the first solids in the solar system.

The effects of heating from decay of ^{26}Al on planetesimals accreted within ~ 2 Ma of the formation of the first solids in the solar system have been investigated by numerous researchers, including Miyamoto et al. (1981), Ghosh & McSween (1998), LaTourrette & Wasserburg (1998), Young et al. (1999, 2003), Merk et al. (2002), Ghosh et al. (2003), Sahijpal et al. (2007), and Elkins-Tanton et al. (2011b). Melting begins in the planetesimal's interior and is tempered by conduction to its cold radiative surface boundary, where heat is released to space according to a black-body law: $F = \sigma T^4$, where σ is the Stephan-Boltzmann constant, T is temperature in Kelvin, and F is heat flux. The resulting interior magma ocean, in many cases, transports heat by convection, and the rate-limiting step for cooling is conduction through the unmelted lid. Models yield contrasting results about the thickness of such conductive lids; some of this disagreement stems from numerical technique and the ways in which convective heat transport in the internal magma ocean is handled.

The parameters for calculating heat from decay of ^{26}Al itself also differ among studies. A thorough analysis of parameters and a discussion about calculation are presented in Castillo-Rogez et al. (2009). Although it is likely that some heating planetesimals did not melt, models indicate that bodies larger than ~ 20 km in radius that accreted within ~ 2 Ma of the first solids in the solar system would have melted to some extent (Hevey & Sanders 2006). Planetesimals of this size are expected to accrete on a timescale of 10^5 years (Goldreich & Ward 1973; see also recent reviews in Chambers 2004 and Canup 2008).

3.2. Melting Embryos and Planets from Energy of Accretion and Metal-Silicate Differentiation

Although a minor heat source in small bodies such as planetesimals, the transformation of kinetic energy to heat during accretionary impacts, combined with the heat from loss of potential energy during core formation (metal-silicate differentiation), is the major heat source for melting in embryos and planets. Separation of metallic from silicate components and sinking of the metallic component to form a core may happen initially in planetesimals during internal melting from radiogenic heating and almost certainly occurred in embryos from kinetic energy of impact. In subsequent impacts of previously differentiated planetesimals or embryos, as long as they are roughly equal in size, the cores of the two bodies are likely to merge (see Canup 2008 and references therein).

The total heat of accretion can be calculated as gravitational binding energy, the energy required to pull all material apart to infinity:

$$E = \int_0^R -\frac{GM(r)dm}{r},$$

where E is accretionary gravitational potential energy (in J); G is the gravitational constant; R is the total radius; $M(r)$ is the mass of the planet as a function of radius r , simply expressed as $\frac{4}{3}\pi r^3 \rho$ (in kg), where ρ is density; and dm can be expressed as $4\pi r^2 \rho dr$. The result of this integration is

$$E = -\frac{16}{15}\pi^2 \rho^2 GR^5.$$

For Earth, total accretionary gravitational potential energy E is $\sim 2 \times 10^{32}$ J; for Mars, $\sim 5 \times 10^{30}$ J; and for Mercury, $\sim 2 \times 10^{30}$ J.

Energy can be converted to temperature in a simplistic way using the expression $\Delta T = \frac{E}{MC_P}$, where C_P is heat capacity, $\sim 800 \text{ J kg}^{-1} \text{ K}^{-1}$ for silicates (Fabrichnaya 1999) and only slightly higher for metallic core material (Bartels & Grove 1991). Each of these planets would be heated to 10^3 – 10^4 K if all accretionary energy were instantaneously applied to the whole planet. This temperature is more than sufficient to melt or even vaporize the entire planet.

Such a reductionist approach, however, gives only a maximum of heating because the heat of accretion is not delivered to the planet at one time, or homogeneously, or completely. In contrast to the short timescale for radiogenic heating in planetesimals, the giant accretionary impacts that produce terrestrial planets are thought to occur over tens to hundreds of millions of years (Wetherill 1980, Chambers 2004, Huestoft & Kohlstedt 2006). This timescale for accretion means that heat of accretion is added to a growing planet at discrete intervals over its accretionary timescale, as planetesimals accrete into planetary embryos and as embryos and planetesimals accrete into planets. A portion of this accretionary heat is radiated to space between each impact.

Furthermore, each giant accretionary impact partitions its energy into the target and the impactor and plume in ways that are not well understood. Canup (2008) maintains that the Moon-forming impact may have melted Earth to a depth of as much as 2,000 km. Tonks & Melosh (1993) argue that a Moon-forming impact would melt 30–65% of the target planet and that complete melting requires that the projectile be at least 40% the mass of the planet.

Heat from the two processes, heat from accretionary impact and heat from metal-silicate differentiation, also may be provided to different regions of the planet. The energy of accretionary impacts primarily heats the exterior, whereas the potential energy of differentiation may be concentrated in the planetary interior, where it is carried by the sinking metal. Thus, an interesting question is whether a large accretionary impact may melt more of the mantle than suggested by these studies because it might melt from both above and below simultaneously (Monteux et al. 2009).

Such a basal magma ocean, overlain by solids, can be formed in one of four ways: First, of course, it can be formed in a planetesimal melted from its interior. Second, in larger planets, it can be formed by a density inversion between solids and liquids where liquids sink. A version of such a magma ocean at the bottom of a partially solid mantle was investigated by Labrosse et al. (2007); they argue that fractional solidification played a role in producing deep, dense material that still might exist today. Third, basal magma oceans may be created by the potential energy of core formation, as discussed in the above paragraph; fourth, they may be created by heating of a radiogenically enriched layer following a magma ocean's fractional solidification and overturn (Elkins-Tanton et al. 2005b).

Whereas impacts of a Moon-forming size are likely to produce a magma ocean that covers the surface of the planet, hemispheric or local magma oceans or ponds are also likely. Tonks & Melosh (1993) demonstrated the likelihood of hemispheric magma ponds and suggested that if isostatic rebound time is less than the freezing time of the magma, then the return of the planet's center of mass to the center of figure would result in extrusion of melt over the planet's surface; see also Reese & Solomatov (2006). The extrusion of magma onto a planetary surface by isostatic rebound may also occur in tidally locked exoplanets, with long-lasting magma ponds on their star-facing surfaces (Gelman et al. 2011).

The measurements of samples from the Moon and Vesta indicating fractionation from a large magma body and the calculations of melting from radiogenic heat and from the physics of accretion, taken together, suggest that at least some early-accreting planetesimals melted from the interior owing to radiogenic heating and that the silicate mantles of the terrestrial planets were processed through serial magma oceans (or ponds), one for each giant accretionary impact (**Figure 1**).

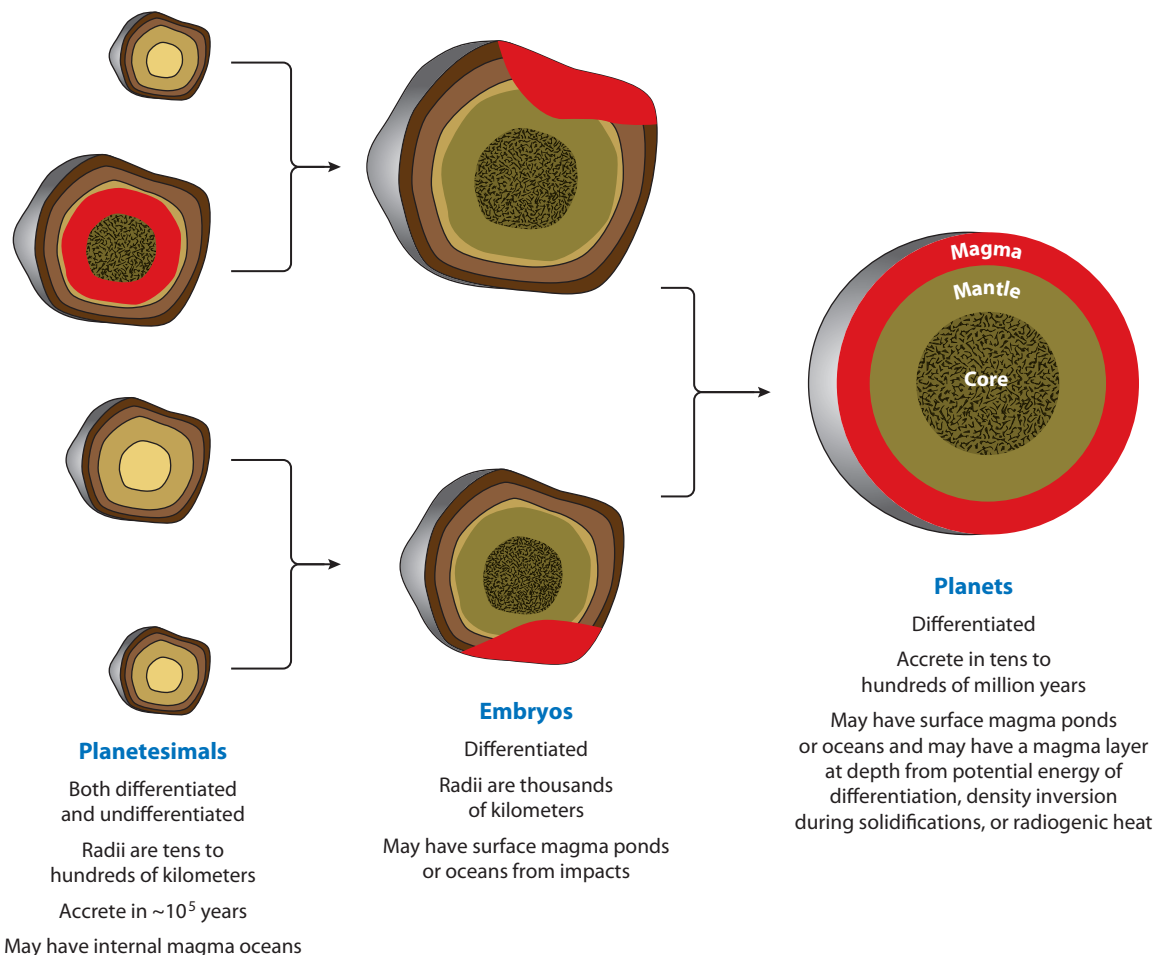


Figure 1

Schematic of differentiation and accretion, showing stages of accretion from planetesimal to embryo to planet (Safronov 1969, Wetherill 1980). Core differentiation likely occurs in several stages, and magma oceans (*red*) (which may be internal or external) likely occur multiple times and to varying extents. The same package of silicate material may melt more than once during these processes—for example, through heat of accretion during several large impacts during growth to planetary size.

4. PHYSICS AND CHEMISTRY OF SOLIDIFICATION

4.1. Heat Transport

Walker et al. (1975) suggested that solidification of the lunar magma ocean would begin at the bottom because the steep slope of the adiabat with respect to the solidus in pressure-temperature space causes them to intersect first at depth (**Figure 2**; see also Solomatov 2000). The greater the pressure range of the planet, the stronger the case for this argument. In small bodies with small pressure ranges, the cooling magma ocean produces a well-mixed liquid adiabat that lies between the liquidus and solidus at all pressures, producing a magma ocean with, to first approximation, equal crystal fraction over the magma ocean depth (**Figure 2**). The view of a magma ocean divided into layers of distinct rheology is simplistic, however, because the magma ocean likely does not

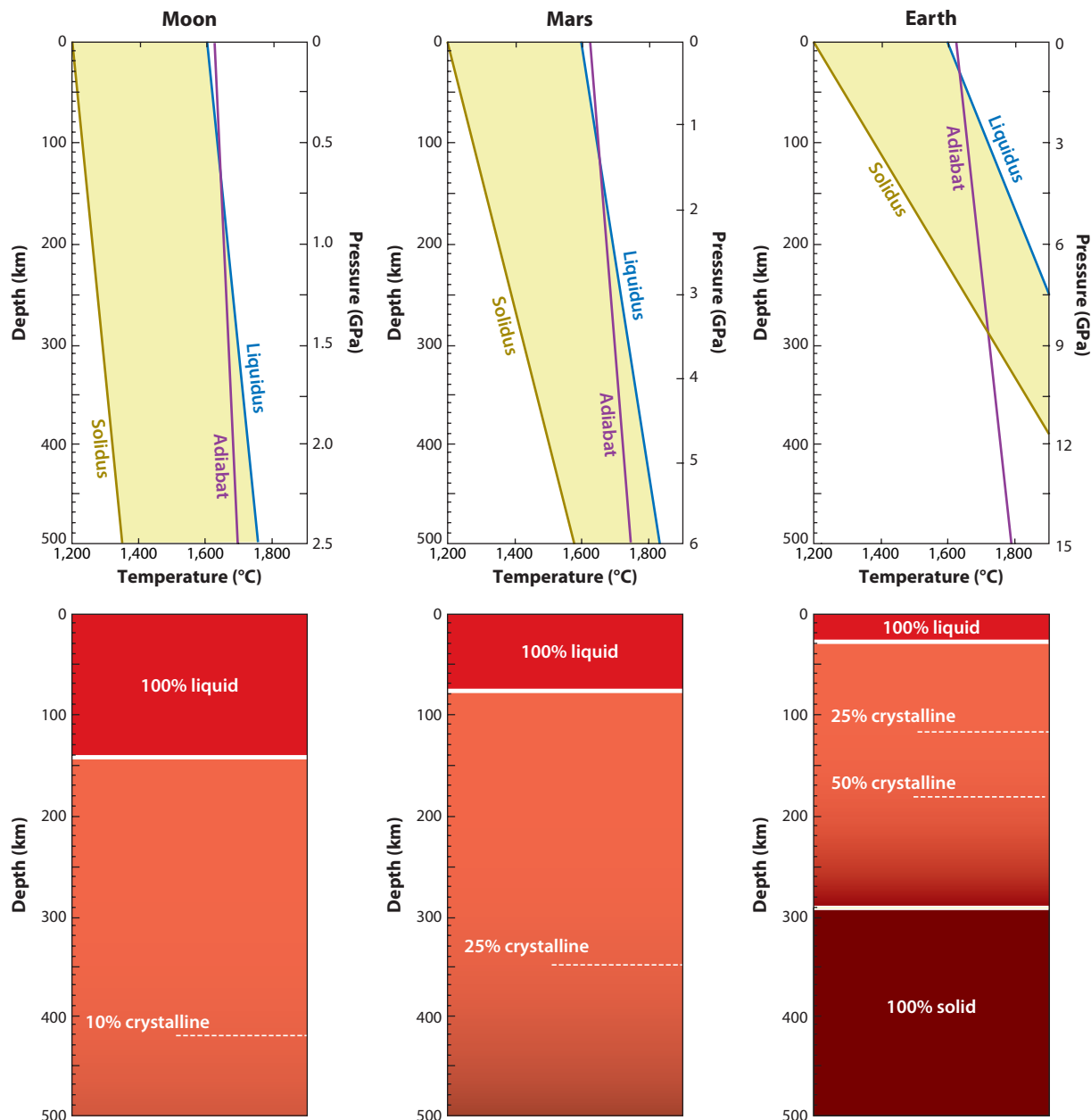


Figure 2

(Upper panels) Solidi, liquidi, and adiabats for planets of different sizes, shown to 500 km depth in each panel. This simplistic view suggests that the planet's pressure range would determine the crystallinity and therefore the viscosity of the magma ocean. (Lower panels) The solid fraction in the magma ocean forms a stress-supporting network at some critical crystal fraction (between 10% and 60%, depending on crystal shape; Campbell & Forgacs 1990, Saar et al. 2001). The viscosity of this "networked" zone becomes close to that of a solid, deforming by thermally activated creep. Above the stress-supported network lies a region of liquid with suspended crystals, above which lies any remaining completely liquid magma ocean (Figure 3 shows a more complex fluid dynamics interpretation).

have a potential temperature that is homogeneous azimuthally or radially; patterns of thermal convection produce packages of magma ocean fluid with different potential temperatures and therefore crystallinities.

Magma oceans are sufficiently extreme in their conditions that they are thought to obtain Rayleigh numbers in the range of 10^{20} – 10^{30} . The Rayleigh number is the nondimensional ratio of thermal buoyancy forces to viscous and diffusive dissipative effects:

$$Ra = \frac{\rho g \alpha \Delta T L^3}{\eta \kappa},$$

where ρ is fluid density, g is gravitational acceleration, α is thermal expansivity, ΔT is the temperature range across the fluid that drives the convection, L is a characteristic length scale for the fluid body, η is dynamic viscosity, and κ is thermal diffusivity. The Rayleigh number of Earth's present-day solid convecting upper mantle, in contrast, is on the order of 10^7 . Magma ocean fluid dynamics therefore differ significantly from solid mantle dynamics and are closely modeled by the dynamics of atmospheres as shown by Priestley (1957, 1959) (see also Golay 1982, Solomatov 2000).

Heat is transported from the interior to the surface in response to cold thermal plumes descending from the upper boundary; these cold plumes are accompanied by hotter, more diffuse return flow from the interior (**Figure 3**). Turbulent shear zones develop with increasing width along the margins of the downgoing plumes, whereas plume interiors may remain relatively unmixed. Heat radiation from the planetary surface to space is inhibited by any solid lid and by

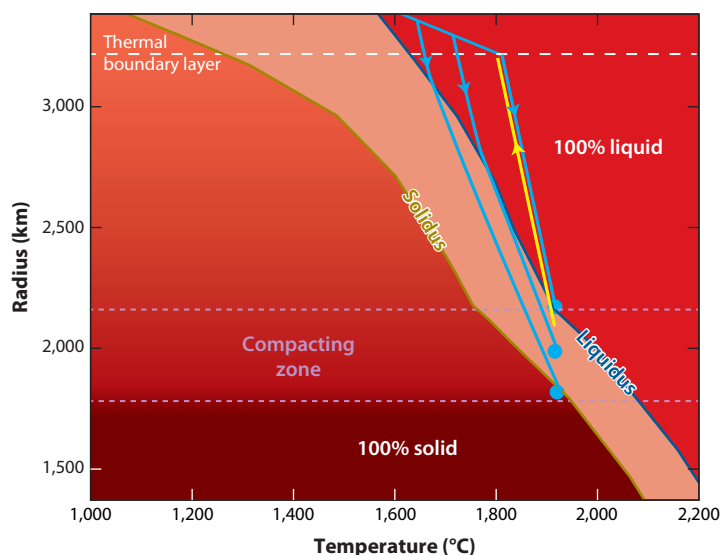


Figure 3

The magma ocean cools from its top boundary, which obtains a conductive temperature profile and periodically sinks in plumes (*blue lines*). Thickness of the thermal boundary layer is highly exaggerated. The plume center follows the blue line representing the coldest temperature; the plume exterior follows the blue lines that represent warmer temperatures. The plume interior therefore gains crystallinity faster. Solids are delivered to the bottom of the magma ocean by the sinking plume, and compaction forces out interstitial liquids. Adiabatic ascent of return flow is shown with the yellow line. Turbulent mixing occurs increasingly at the edges of the plume during descent, widening the plume, and more diffuse return flow continues mixing. Depth scale appropriate for a deep Martian magma ocean. Adapted from figure 1 in Parmentier et al. (2007).

the emissivity of any atmosphere (Abe & Matsui 1985, 1986), with its own conductive/convective profile. Processes of magma ocean solidification are discussed in Abe (1993, 1997), Solomatov & Stevenson (1993a,b,c), Solomatov (2000, 2007), Elkins-Tanton et al. (2003), Elkins-Tanton (2008), and Monteux et al. (2009).

Heat transfer through the free surface of a convecting fluid is rapid, as high as 10^6 W m^{-2} before a thick atmosphere develops, and heat transfer through a significant atmosphere is 10^2 W m^{-2} or less (Abe & Matsui 1986, Zahnle et al. 1988, Elkins-Tanton 2008). Conductive heat transfer, however, is far slower than convective heat flow. Solidification of a magma ocean that has a free surface, therefore, would be rapid, possibly as rapid as 10^5 to 10^6 years for a deep terrestrial magma ocean (Elkins-Tanton 2008). If the magma ocean had a solid conductive lid, then heat flux would be reduced by many orders of magnitude, and thus cooling and solidification times would be lengthened proportionately.

Conductive lids on magma oceans can be formed in one of three ways. First, in the case of an internal magma ocean on a planetesimal, the outer portions of the planetesimal may remain unmelted and thus conductive. Second, buoyant phases may form in the cooling magma ocean and float to the surface, as is hypothesized to have happened with anorthosite plagioclase on the Moon. Third, the mafic silicate magma of a surface magma ocean may quench to a solid crust against the cold of space.

During solidification, the quench formation of such a stagnant conductive lid on the magma ocean of a planet-sized body is unlikely. First, even a modest insulating atmosphere maintains the magma ocean surface above its liquidus initially and above its solidus for much of solidification of the volume of the magma ocean. The importance of an early water atmosphere has been recognized and studied by Abe & Matsui (1985), Matsui & Abe (1986), Zahnle et al. (1988), Abe (1993, 1997), and Elkins-Tanton (2008). These studies found that even a nominal steam atmosphere retains sufficient heat to keep the planetary surface above its solidus. Degassing from the magma ocean thus maintains the magma ocean's liquid surface. This critical conclusion means that a magma ocean beginning with as little as 100 parts per million (ppm) water retains a free liquid magma ocean surface (the likelihood of late degassing is discussed below, but even a thin initial atmosphere residual to impact would keep the surface free from quench material).

Second, any quench material that forms would sink. In the absence of such atmospheric insulation, the surface of the magma ocean quenches to solid against the cold of near-space or a thin atmosphere. This solid is denser than the magma ocean liquids beneath and therefore is prone to foundering (Walker et al. 1980, Spera 1992). On terrestrial volcanic magma pools, solid crusts may persist through support at the edges of the pool, but on a planetary magma ocean, gravitational forces dominate and any solid lid would founder. This lid is vulnerable to convective stresses on its bottom boundary and to disruption by impacts. Such a lid may persist only if it is relatively whole over the entire planet and thick enough to resist breakup from convective stresses and impacts. A lid tens of kilometers thick is required for this degree of strength on a planet-sized body, but evolving from no lid to a complete, thick lid would appear to be unlikely in the active environment of a magma ocean.

The only likely way to form a relatively complete conductive lid on a planetary-scale magma ocean is by flotation of buoyant phases. This is likely to occur only on small, dry planets such as the Moon; on larger planets, plagioclase does not become stable until the magma ocean has solidified to a high degree close to the planetary surface, where high crystal fractions prevent flotation. Additionally, water suppresses the crystallization of plagioclase, further delaying its appearance in the magma ocean (e.g., Goldsmith 1982). Planetary-scale magma oceans therefore are unlikely to develop conductive lids, and they have high heat fluxes and rapid solidification, tempered only by any degassed atmosphere.

4.2. Crystal Settling Versus Entrainment

Whether growing mineral grains remain suspended or settle out is fundamental to understanding the chemical evolution of a magma ocean. The compositional process of solidification can be viewed simply as consisting of two end members: fractional solidification and batch solidification. For fractional solidification to occur, mineral grains must settle from flow and be effectively removed from communication with the remaining magma ocean liquids. The residual liquid composition thus evolves with the removal of each package of solidified material. In batch solidification, liquid contact and equilibrium are maintained throughout solidification. Fractional solidification of a magma ocean thus leads to predictions entirely different from those of equilibrium solidification for mineral assemblage, cumulate bulk composition, and cumulate trace-element compositions.

The major-element composition of the magma ocean determines the minerals that will crystallize, their densities, and therefore their gravitational drive for settling. The bulk compositions of the terrestrial planet's mantles (and therefore of their magma oceans) are not known with great precision, but their similarities are greater than their differences. The oxides SiO_2 , Al_2O_3 , MgO , FeO , and CaO make up 97 weight percent (wt%) or more of the bulk silicate fraction of terrestrial planets: SiO_2 makes up $\sim 41\text{--}46$ wt%; MgO , $\sim 30\text{--}40$ wt%; FeO , $\sim 8\text{--}18$ wt% (Mars's mantle is far more iron-rich than Earth's mantle); and CaO and Al_2O_3 , $\sim 3\text{--}6$ wt% each. All other elements are present in quantities less than ~ 1.5 wt% (Priestley 1957, Ringwood & Kesson 1976, Morgan et al. 1978, Wänke & Dreibus 1984).

Crystal settling (and the possibility of fractionation) is dependent on the densities of minerals and their difference from the density of the coexisting liquid. Mineral density depends on mineral composition, pressure, and temperature. Typical mineral assemblages expected for bulk silicate compositions such as these are given in **Figure 4**.

Mafic phases (those containing primarily silica, magnesia, and iron oxides) are generally denser than their coexisting liquids, although there are some notable density crossover points at relevant pressures at which mafic phases are buoyant in their coexisting liquid. The prospect of neutrally buoyant minerals has led to the suggestion that a magma ocean convects in layers separated by mineral septa (Franck 1992, Morse 1993). Olivine and pyroxene sink with respect to their coexisting dry mafic silicate liquids at shallow depths, but at pressures greater than approximately 7.5 GPa to 10 GPa, depending on bulk composition, both minerals become positively buoyant and thus float at this point of neutral buoyancy (Stolper et al. 1981). At still higher pressures, majorite may have a density crossover with coexisting liquids (Ohtani & Maeda 2001, Stixrude & Karki 2005, Mosenfelder et al. 2007), leading to a basal magma ocean (Labrosse et al. 2007).

If a septum forms, however, convective motions may not allow it to produce compositional segregation. The depth of neutral buoyancy for both olivine and pyroxene is ~ 250 km for a bulk mantle magma ocean composition for Earth; this depth is near the center of the radius range of the remaining convecting magma ocean, where vertical convective velocities are the highest and therefore where magma is most likely to penetrate the septum and mix liquids from the two sides of the septum. Rapidly moving, cold downwelling plumes are therefore likely to disrupt any mineral septa, mixing liquids back to homogeneity.

Solidification beneath the pressure of neutral buoyancy is therefore likely to be a combination of fractional and batch solidification. In addition, in the range of pressure over which olivine and pyroxene are positively buoyant, garnet is cocrystallizing and is denser than the coexisting liquid. Thus, garnet may settle while olivine and pyroxene remain entrained, creating additional compositional differentiation (Elkins-Tanton et al. 2003).

Other phases, most notably plagioclase feldspar, are buoyant with respect to their coexisting liquids. Plagioclase requires low pressures for stability (less than ~ 1 GPa) and enrichment in

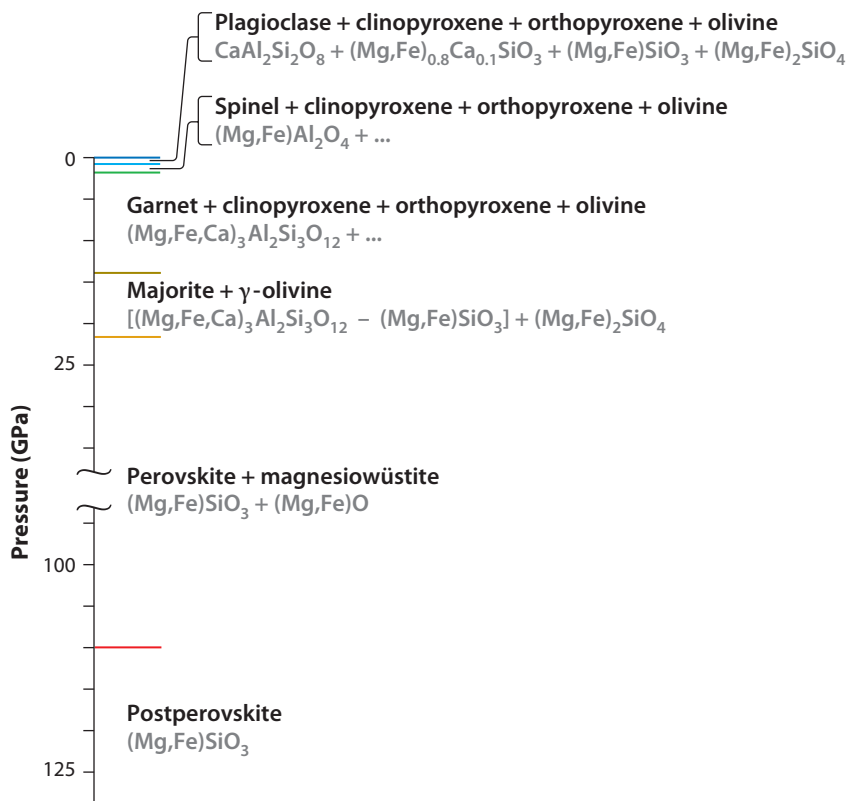


Figure 4

Expected mineral assemblages determined by pressure for an approximately Earth-like bulk mantle composition. Trace elements and volatiles are not shown but would be incorporated into each phase in varying quantities. Perovskite and postperovskite contain calcic and aluminous phases in addition to the mafic phase shown. The exact stability boundaries, compositions, and phase fractions need to be determined uniquely for each bulk composition and planetary size. Major mantle phases at pressures higher than the postperovskite transition have not been determined with assurance. Adapted from figure 2 in Elkins-Tanton (2011).

alumina; its stability and flotation are likely to be possible only on small bodies (see, however, Harrison 2009). In a Martian magma ocean, for example, plagioclase can solidify only in the last tens of kilometers near the surface, at the end of magma ocean solidification when crystal density is likely to be high and flotation difficult. Liquid dense enough to produce flotation also requires the coexisting liquid to contain more than a few weight percent iron oxide (see Figure 5).

The problem of settling versus entrainment of grains in vigorous fluid flow, however, is not thoroughly understood. Numerous central questions must be answered more satisfactorily before the process of solidification of magma oceans can be addressed in detail and with confidence. These include the rheology of crystal-rich liquids, the density of mafic liquids under great pressure, crystal size distributions, and the mode and timescales of liquid expulsion from cumulate piles. This statement made, progress has been achieved to varying degrees in all of these fields.

Mass conservation requires that the average vertical velocity of convective motions across any horizontal plane in the magma ocean vanishes, so both upward and downward fluid motions must occur. The settling of dense crystals results in persistent sinking relative to the convecting liquid,

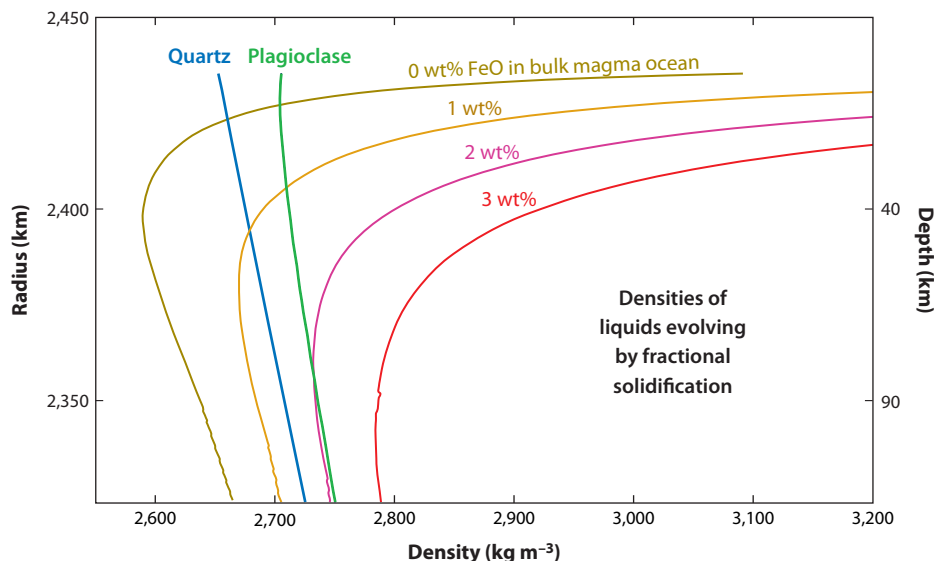


Figure 5

Flotation of a buoyant mineral phase such as plagioclase depends on the density of the coexisting liquid. Models for varying iron oxide content in the liquid are shown. Fractionation increases iron oxide content and therefore liquid density. Fractionated mafic phases are assumed to sink, so evolving liquid moves to shallower depths. Adapted from figure 4 in Brown & Elkins-Tanton (2009). Abbreviation: wt%, weight percent.

so crystals must settle out even if the settling velocities are small compared with the velocities of convective motions (Marsh 1988, Martin & Nokes 1989). Nonetheless, Tonks & Melosh (1990) argue that mineral grains are unable to settle on the available timescale in a turbulently convecting magma ocean; in contrast, Martin & Nokes (1989) argue that settling in magma oceans is likely. Further work demonstrates that settling may be possible and in some cases probable.

The ability of crystals to accumulate at the bottom of a magma ocean depends on a competition between settling and entrainment (Martin & Nokes 1989, Tonks & Melosh 1990, Solomatov & Stevenson 1993c, Solomatov et al. 1993). Adjacent to a horizontal boundary, vertical velocities are small. However, convective motions can move particles horizontally along the boundary toward regions of convergence beneath buoyant upwellings, where viscous stresses can lift the particles. Laboratory experiments on a fluid layer heated from below suggest that particles smaller than diameter D can be entrained in this way (equation 5 in Solomatov et al. 1993):

$$D = \frac{1}{\Delta\rho g} \left(\frac{0.1\eta\alpha g F}{C_p} \right)^{0.5},$$

where $\Delta\rho$ is the density difference between the particles and the liquid (~ 100 – 300 kg m $^{-3}$), g is the acceleration of gravity, η is the fluid viscosity (0.1 Pa s; Reese & Solomatov 2006), α is the coefficient of thermal expansion (3×10^{-5} K $^{-1}$), C_p is the specific heat (~ 800 J kg $^{-1}$ K $^{-1}$ for silicates; Fabrichnaya 1999), and F is the heat flux from the magma ocean (between 10^{-4} and 10^6 W m $^{-2}$, depending on the existence of an atmosphere; Abe & Matsui 1986, Zahnle et al. 1988, Elkins-Tanton 2008). Thus, higher viscosity and higher heat flux result in larger particles that can be entrained (**Figure 6**).

Solomatov & Stevenson (1993a) argue that solids in a magma ocean are likely to be between 0.001 cm and 1 cm in diameter, overlapping the boundary between fractional and nonfractional

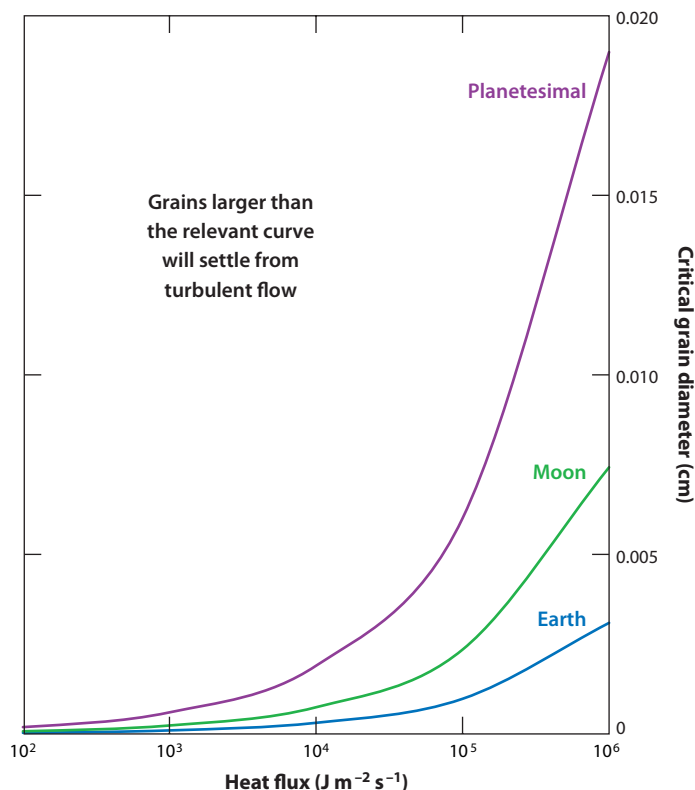


Figure 6

Crystal settling. At a given heat flux, mineral grains larger than the critical crystal diameter will settle, according to Solomatov et al. (1993), whereas smaller grains will remain entrained in flow. Heat flux from surface magma oceans is at the high end of the scale early in solidification, and, with an insulating atmosphere or conductive lid, it may fall to smaller values than the ones shown. This analysis shows settling to be likely but neglects the effects of high crystallinity in the liquid.

solidification. In this analysis, both fractional and equilibrium crystallization might be expected over the course of magma ocean solidification. However, the experiments that motivated this solution are, of course, an inexact parallel to a magma ocean because, at the least, the magma ocean is cooled from the surface rather than heated from below.

Early in solidification of any body, the heat flux is maximized before a thick atmosphere or a conductive lid slows energy transfer into space. Under a high heat flux, grains must be larger to settle from turbulent flow. Although larger grains are required for settling, settling may be facilitated at this time by relatively low crystal fraction, at least in larger bodies. Later in the solidification process, the heat flux is lower, and smaller grains may be able to settle. However, if the crystal fraction is high, the higher effective viscosity of the fluid will work against settling, as will frequent collisions among grains. A further consideration is how fast crystals grow once nucleated. Smaller entrained crystals could continue to grow until reaching a size at which they will settle out; lower heat flux and thus slower solidification may allow both larger grains and larger grain-size distributions to develop, both of which facilitate settling.

Thus, on the basis of arguments of grain size required for settling, likely grain-size distributions, and compositional observations (Walter & Trønnes 2004, Walter et al. 2004, Liebske et al.

2005), both Earth and Mars likely experienced some fractional solidification and some equilibrium solidification of their magma oceans. Smaller bodies, such as the Moon or planetesimals with conductive lids, may have more efficient and longer-lived settling (and therefore fractional solidification) because of their slow cooling rates.

4.3. Compositional Differentiation

Magma ocean processes exert significant control over compositional differentiation in planetesimals and in planets. First, metallic core segregation may require at least some degree of silicate melting. Second, the solidification of molten silicates produces both major- and trace-element differentiation, including degassing of volatiles and enrichment of incompatible elements in fractionating liquids. Processes of solidification, therefore, strongly influence or control the initial compositional heterogeneity of planets.

4.3.1. Metallic core formation. The first step in compositional differentiation is segregation of metals into a core. The segregation of metallic iron and nickel into a planetary core is widely thought to require a magma ocean (Ringwood 1966, Davies 1985) because the wetting angle of these metals on silicate mineral grains is too high to allow percolative flow of metals in solid silicates (see Stevenson 1990 and references in McCoy et al. 2006). (Some researchers have proposed mechanical processes that might allow core segregation; see Hustoft & Kohlstedt 2006.)

The first metallic melts from chondritic materials at low pressures occur at approximately 950°C and produce metals with approximately 40 mol% sulfur (Kullerud 1963), an amount insufficient to form interconnected networks in olivine. Terasaki et al. (2005) showed that in iron-rich Martian mantle, high dihedral angles prevent percolation of iron-rich core material at pressures above 3 GPa. Core formation through a solid mantle, they conclude, is impossible. Formation of a core requires a magma ocean.

Both differentiated and undifferentiated planetesimals later accreted into planetary embryos and subsequently into planets. Thus, chemical equilibrium or disequilibrium processes between silicate and core metals likely occurred at several pressures and at several times during the growth of a terrestrial planet (**Figure 1**) (Rubie et al. 2011). This scenario has implications for determining the differentiation age of rocky bodies through the use of the ^{182}Hf - ^{182}W system (e.g., Lee & Halliday 1995, Touboul et al. 2007, Yin et al. 2007).

4.3.2. Volatile behavior in magma oceans. Major concerns of the planetary formation community are the delivery mechanisms to, and behavior of, water and other volatiles in growing planets. Following magma ocean solidification, the abundance and distribution of hydroxyl and carbon, in particular, determine mantle viscosity and melting temperature and therefore convective vigor, volcanic activity, atmospheric composition, mass, and eventual clement surface conditions.

How much water and other volatiles might be present in a planetary magma ocean? Melting in planetesimals would have a strong influence on the quantity and distribution of volatiles in the planetesimal, and thus on the delivery of volatiles to growing embryos and planets. Eucrite meteorites from Vesta, for example, are lower in volatiles than are chondrites (meteorites that never underwent melting) (Jarosewich 1990). The volatiles are thought to have been released to space upon eruption of the original eucritic lavas (Mittlefehldt 1987).

Not all early-accreting bodies, however, are expected to melt entirely. In some regions, their unmelted crusts retain near-original volatile contents, and in other regions, they obtain higher water contents through fluid fluxes from the interior. As the planetesimal's interior heats radio-genically, volatile-rich minerals begin to break down and release fluids that percolate toward the

exterior (Young 2001). Some of the fluid may metasomatize the unmelted crust, some may freeze near the surface, and some may be released to space. As heating in the interior proceeds to the silicate solidus, the material being melted is likely to be nearly dry through the dehydration effects of heating to that point. Planetesimals accreting into embryos likely are heterogeneous in volatile content.

If this scenario of volatile transport and composition in planetesimals holds true, the addition of volatiles to an eventual planetary magma ocean produced by giant impacts appears to be inevitable. The impactor also may be partially devolatilized upon impact (O'Keefe & Ahrens 1977, Canup 2004, Ni & Ahrens 2005, Pahlevan & Stevenson 2007); if this is the case, then retention of the atmosphere and its interaction with the magma ocean become the pertinent questions. Both Vickery & Melosh (1990) and Ni & Ahrens (2005) argue that a Mars-sized impactor would eject a portion of the atmosphere above the plane tangent to an Earth-sized planet at the point of impact. Genda & Abe (2003) studied a solid-surface planet and found that after being struck by a Mars-sized impactor, the Earth retained 70% of its atmosphere; if the impactor had only 10% the mass of Earth, the Earth retained 80% to 90% of its atmosphere. The planet, therefore, has a more massive atmosphere following the impact than it had before. Planets with liquid surfaces lose far more of their atmospheres during large impacts because of enhanced surface motion. During the magma ocean stage, large impacts therefore would have removed large fractions of, but not the entire, atmosphere (Genda & Abe 2005). This atmosphere may then reequilibrate with the magma ocean, dissolving volatiles up to the saturation limits of the magma.

Examining higher-pressure processes, Rubie et al. (2004) demonstrated that at high temperatures and pressures in magma oceans, iron moves into a metallic phase rather than the oxidized phase, reducing the possibility of significant metallic iron oxidation and water loss. In deep magma oceans, therefore, metallic iron may remain stable, leaving water in the magma ocean. Righter & Drake (1999) further argue that water in a magma ocean is necessary to explain trace-element characteristics of Earth.

Core formation therefore may have depended on a preference in iron to stay metallic at high pressures, on an overall low oxygen fugacity, or simply on the ability of iron to sink quickly through any potentially oxidizing magma slurry and to aggregate into a core before oxidation could occur. This suggestion is supported by the lead, xenon, and strontium data and modeling of terrestrial accretion by Halliday (2002). Through the use of a Stokes law for sinking iron and a simple diffusion law for oxygen, it can be shown that even on small planetesimals, iron spheres larger than a centimeter form a core without reacting with water in the magma they sink through. In these ways, terrestrial planets may accrete from differentiated planetesimals that contain water and carbon dioxide and still retain a high metallic iron ratio. This scenario may also be required to explain the nonzero water content of many achondrites.

Almost all trace elements and volatiles are incompatible in almost all common mantle minerals. Nominally anhydrous minerals can contain a dynamically and petrologically significant amount of OH, however, as much as 1,000–1,500 ppm for olivine and orthopyroxene (e.g., Hauri et al. 2006). Carbon dioxide partitions into mantle minerals in even smaller quantities (Shcheka et al. 2006). Even complete fractional solidification seldom drives water contents of the evolving magma ocean liquids high enough to create water-saturated minerals (water-saturated magma ocean liquids are discussed below); initial bulk magma ocean compositions above 1–3 wt% water are required in Earth-sized planets to create water-saturated, shallow, late-stage cumulates (Elkins-Tanton 2011). Whether or not saturation is reached, later cumulates may form under higher oxygen fugacities than did early cumulates in a low-water magma ocean.

The distribution during magma ocean solidification of any incompatible elements in common mantle minerals is largely controlled by fractional crystallization. As solids are removed, remaining

liquids are progressively enriched in incompatible elements such as radiogenic potassium and uranium and in volatiles such as hydroxyl, nitrogen, and carbon. Thus, interstitial liquids dominate the cumulate mantle bulk composition for incompatible elements. The incompatible-element content of both early- and late-solidifying cumulates is dominated by retained interstitial liquids. Late-solidifying minerals obtain slightly higher trace- and volatile-element compositions simply from partitioning out of an enriched fluid, but even late-solidifying minerals contain only tiny quantities of volatiles in comparison with the enriched interstitial fluids from the evolved magma ocean. Tonks & Melosh (1990) argue that crystallization may be rapid in comparison with liquid percolation, so high fractions of retained liquid may be left in the cumulate mantle. Therefore, the earliest solid mantle of the planet would have highly heterogeneous incompatible-element composition. Consequently, interstitial melts and surface films may control the internal trace-element budget of the terrestrial planets, including hydroxyl, rare-earth elements, and noble gases.

4.3.3. Degassing the early atmosphere. Volatile loss from a magma ocean into a growing atmosphere is neither instantaneous nor inevitable. The work to date assumes that all volatiles in excess of the saturation limits of the silicate liquid would be degassed into the growing atmosphere (Abe & Matsui 1985, Zahnle et al. 1988, Elkins-Tanton 2008). Two critical barriers must be overcome, however, for volatiles to leave the magma ocean and enter the atmosphere: The magma ocean liquids must attain supersaturation in the volatiles in question, and bubbles of these volatiles must breach the upper boundary layer and burst at the surface.

The posited bulk water content of magma oceans on terrestrial planets is likely to be similar to or less than that of achondritic meteorites, in the assumption that planets are built from previously differentiated bodies. Achondritic meteorites contain 3 wt% water or less; most achondrites have a water content far less than 1 wt% (Jarosewich 1990). The saturation of silicate magma with water depends strongly on pressure (see quantification of this in Papale 1997). At these low abundances and particularly under any significant atmospheric pressure, bulk magma ocean liquids are not supersaturated in water. Because water (or hydroxyl) is incompatible in the pertinent silicate minerals, fractional solidification is the most reasonable way to enrich the magma ocean liquids in water to the point that the magma is supersaturated and thus inclined to nucleate bubbles.

In an Earth-scale magma ocean, a whole-mantle bulk magma ocean with 10 ppm water does not reach saturation until after 95 volume percent (vol%) solidification via pure fractional crystallization [assuming that equilibrium traces of hydroxyl are sequestered in solidifying minerals and that the techniques and databases of Elkins-Tanton (2008) are used]. The same magma ocean that begins with 100 ppm water saturates at ~90 vol%; one that begins with 500 ppm saturates at ~80 vol%. For carbon dioxide, ~95, ~80, and ~40 vol% solidifications are required for saturation for magma oceans that begin with 10, 100, and 500 ppm CO₂, respectively. Degassing into an atmosphere can begin only when the first gas reaches sufficient supersaturation to form bubbles; other species then diffuse into the newly formed bubbles. Recall that Earth's hydrosphere is equivalent to 200–300 ppm water in Earth's mantle, so a 100-ppm magma ocean with later additions of volatiles is not unreasonable under a scenario of planet-building with achondritic material with later hydrous additions from chondrites or comets.

For degassing to occur, bubbles must be large enough to decouple from fluid flow and rise through the boundary layer to burst at the surface. Alternatively, degassing requires a level of supersaturation that allows dense bubble nucleation and the rise of the bubbly magma ocean fluid as a continuum; this fluid forms a foamy diapir that reaches the planetary surface.

The bulk volatile composition of the magma ocean exerts control over degassing quantities and rates. Kadik & Lukanin (1986) note that if the magma ocean originated from material containing graphite, its reducing nature would change the speciation of outgassing. Hirschmann & Withers (2008) extended this work to show that, rather than degassing, a magma ocean in equilibrium with graphite retains significant portions of the planet's carbon budget in its solidifying interior. In contrast, if the magma ocean is not sufficiently reducing to stabilize graphite, carbon species dominate the degassing budget and the emissivity of the atmosphere (Elkins-Tanton 2008).

Following solidification, the young planet is blanketed by a dense, possibly steam-dominated atmosphere that cools toward the water critical point. Once sufficient cooling is complete, the atmosphere collapses into a water ocean. This cooling process requires millions to tens of millions of years (Abe & Matsui 1986; Matsui & Abe 1986; Kasting 1988; Zahnle et al. 1988; Elkins-Tanton 2008, 2011).

4.3.4. Development of density gradients in the solidifying cumulate mantle. The cumulate mantle that results from solidification of a magma ocean is likely to be gravitationally unstable, that is, likely to have denser material overlying less dense material (Solomatov 2000). Three processes may produce an unstable density gradient: The first is enrichment of the magma ocean liquids in heavy elements as fractional solidification progresses; the second is thermal expansivity because the solids form at solidus temperatures, colder near the planetary surface; and the third is Fe-Mg exchange in mafic minerals.

Fractional solidification of a lunar magma ocean may produce an extreme enrichment in titanium in late-stage liquids, allowing the solidification of high-titanium oxides such as ilmenite and ulvospinel (Schnetzler & Philpotts 1971). An oxide-rich layer, as predicted in these solidification models, is denser than underlying, earlier-solidifying cumulates and thus prone to flow in the solid state through convective instabilities to gravitational stability (Spera 1992, Hess & Parmentier 1995). Hess & Parmentier (1995) examined the process of creating these late cumulates and their fate. When they sank into the lunar interior, they carried the radiogenic elements with them and thus heated over time. This hot, titanium-rich layer deep in the Moon may have been a progenitor for high-titanium melts that later erupted onto the lunar surface.

The development of gravitational instabilities and solid-state overturning of the cumulate mantle have been discussed in a variety of papers, including Hess & Parmentier (1995), Solomatov (2000), Elkins-Tanton et al. (2003), and Elkins-Tanton (2008). The overturning process can deliver dense, cold, incompatible-element-rich material to the core-mantle boundary, creating a core flux sufficient to drive a magnetic dynamo and producing a dense layer at the core-mantle boundary that may persist over the age of the solar system but that does not participate in thermally driven convection (Elkins-Tanton 2008). Simultaneously, hot upwelling material may melt adiabatically and produce the planet's first basaltic crust (Elkins-Tanton et al. 2005a). Once overturn is complete, the compositionally stable mantle is resistant to the onset of thermal convection and may remain quiescent for a period of time (Zaranek & Parmentier 2004). A violently convecting early planetary interior may therefore be delayed by magma ocean solidification.

Therefore, a body can obtain its initial crust in one of several ways. Planetesimals may have an unmelted primitive lid over an internal magma ocean. Bodies of embryo size, such as the Moon, have a sufficiently small internal pressure range that allows them to produce plagioclase that floats to form an initial crust. Both of these crusts may be breached or intruded upon by contemporaneous volcanism. Larger bodies can obtain their earliest basaltic to tonalitic crusts through decompression melting during overturn of gravitationally unstable magma ocean cumulates or through later thermal convection (**Figure 7**). All of these processes for crustal production are tied to magma oceans, either directly or indirectly, through remelting of magma ocean cumulates.

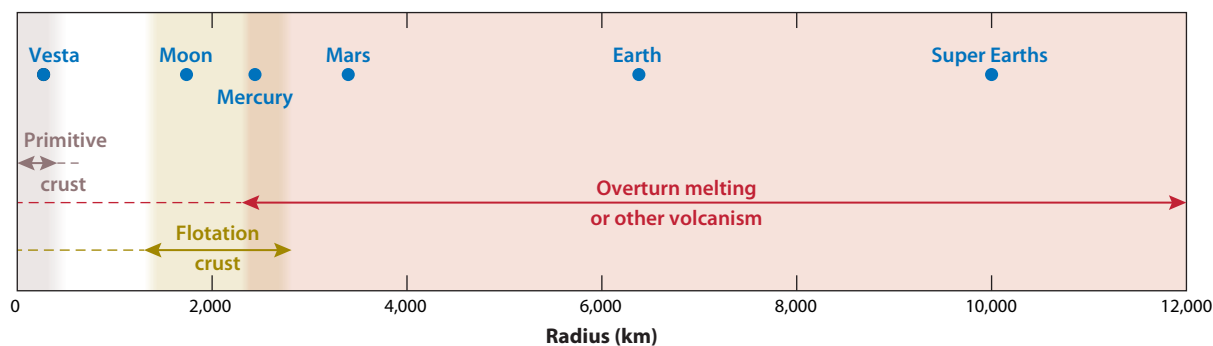


Figure 7

The mechanism for producing a body's first crust depends largely on its size. Planetesimals may retain a primitive undifferentiated crust if melted from within. Small planets with low pressure ranges can fractionate plagioclase, which may float to the surface given sufficient time. Larger planets do not retain primitive crusts or float mineral but develop basaltic crusts from melting during cumulate mantle overturn or from later volcanism. Extrusive volcanism may occur on any body given sufficient driving force but occurs most likely on larger bodies. Adapted from Brown & Elkins-Tanton (2008).

5. SUMMARY

Theory and some observations point to the occurrence of magma ponds or oceans in the early evolution of all terrestrial planets and in many early-accreting planetesimals. The complexity of the fluid dynamics and heat transport in a magma ocean stimulates active research. Heat transport to space is facilitated by convection and limited by conduction. A planet-sized magma ocean is unlikely to have a solid conductive lid and thus cools faster than a smaller body such as the Moon, which can produce a flotation lid of low-pressure plagioclase. Planetesimals with thick, porous, primitive lids may solidify the most slowly of all.

A particularly critical research area is crystal settling in turbulent flow. The ability of grains to settle determines whether the magma ocean can fractionally crystallize. Fractional crystallization, in turn, controls many processes of compositional differentiation, including (*a*) the enrichment of incompatible elements in evolving fluids sufficient to degas an early atmosphere efficiently and enrich a mantle layer in radiogenic elements; and (*b*) the development of a sufficiently iron-rich, late-stage liquid to float buoyant phases.

The origin and fate of volatiles in the magma ocean are critical to understanding planetary accretion and the development of a habitable planet. Increasing research indicates that magma oceans may have contained a level of volatiles sufficient to produce thick atmospheres yet able to coexist with metallic core formation.

Further understanding of magma ocean processes and rates will allow a timeline of the early solar system to be pieced together more completely than would be possible with an understanding of radiogenic ages alone (**Figure 8**). The ages of core formation in iron meteorites, the formation of chondrules, and the genesis of angrites and eucrites indicate internal melting in early-forming planetesimals. Formation of the Moon is contemporaneous with the time of onset of the lunar magma ocean, and magma ocean solidification past ~75 vol% is marked by plagioclase flotation to form the anorthitic crust. The time of the Moon-forming giant impact is not well constrained: It could be moved to any time between 4.54 Gya and almost 4.41 Gya (**Figure 8**) without being inconsistent with Hf-W and crustal age data.

The ages of the oldest terrestrial zircons (Harrison 2009) form a concrete marker on the young end: Earth had to have had water oceans with ocean sediments remelting to granite in

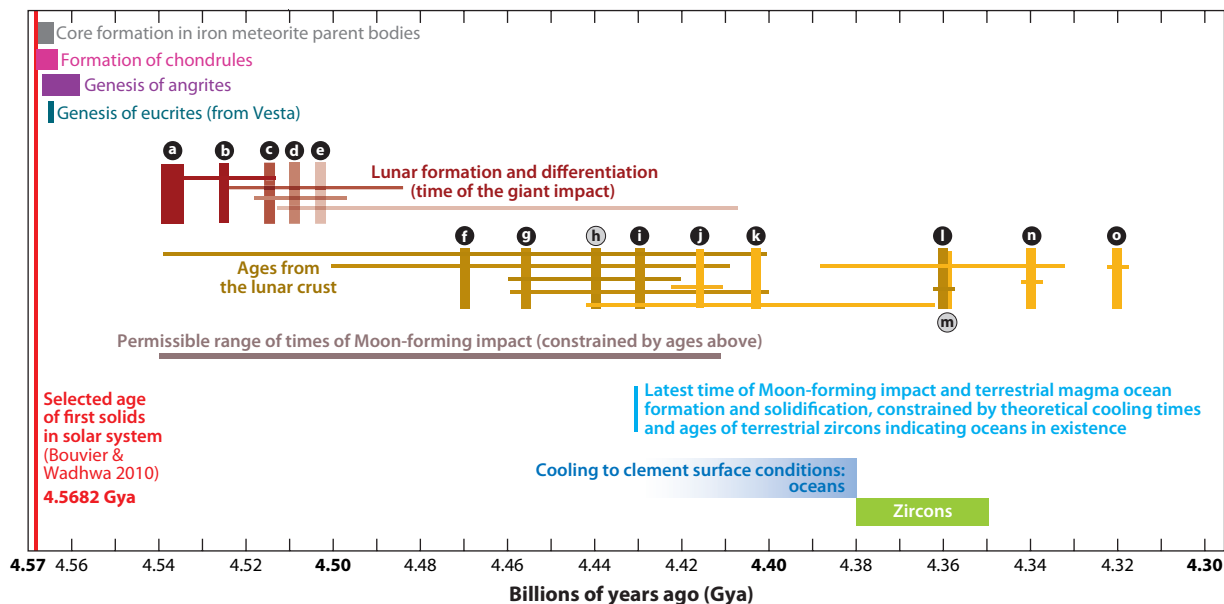


Figure 8

Accretion timeline. Timeline of early solar system processes related to magma oceans, constraining the age of the Moon-forming impact using models and data from the Moon and Earth. References for ages (*far left*) of core formation in iron meteorites, formation of chondrules, and genesis of angrites and eucrites from Elkins-Tanton et al. (2011b) and this review. Lettered circles correspond to references associated with the lunar formation entries (*red*) and lunar crust entries (*dark yellow*) (the complete list of references is given in the Supplemental Material associated with this review; to access it, follow the Supplemental Materials link from the Annual Reviews home page at <http://www.annualreviews.org>). Lettered circles *b* and *m* refer to the same sample. Zircon ages from Harrison (2009). Adapted from figure 1 in Elkins-Tanton et al. (2011a).

time to solidify these zircons by 4.38 Gya. Assuming ~ 50 Ma is required for Earth to cool to clement surface conditions following the giant Moon-forming impact (Matsui & Abe 1986, Elkins-Tanton 2008), the Moon-forming impact would be predicted at no later than ~ 4.43 Gya, as shown in **Figure 8**. Zahnle et al. (1988) predict a shorter cooling time, approximately 5 Ma, which would allow the impact to occur as late as 4.41 Gya, a time still constrained by the lunar crustal ages. Thus, the existing data indicate that while the magma ocean on the Moon was still solidifying, perhaps as a result of tidal heating from Earth (Meyer et al. 2010), Earth had already attained clement surface conditions with water oceans and formation of granite.

DISCLOSURE STATEMENT

The author is not aware of any affiliations, memberships, funding, or financial holdings that might be perceived as affecting the objectivity of this review.

ACKNOWLEDGMENTS

This work was funded by grant #0747154 through the NSF Astronomy program and through the NASA Lunar Science Institute at Brown University and MIT.

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