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Lunar Magma Ocean, Size

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Introduction

The Earth's Moon today is chemically and physically stratified as a result of planetary differentiation that occurred within the first several 100 Myrs of the Solar System (Fig. 1a). The aseismic lunar crust varies in thickness from 60 to 100 km, and its surface is visually characterized by the mare basalts, although these comprise <1 vol.% of the lunar crust (Taylor 1979, Fig. 1b). The lunar mantle extends from beneath the crust to depths of ~1,100 km with a prominent seismic discontinuity at *c*. 500 km and detectable seismic events, "moonquakes," which cluster at ~700–1,200 km depth (Lammlein et al. 1974; Nakamura et al. 1982). The nature of the lunar core is poorly understood but likely comprises <2 vol.% of the Moon and is constrained to a radius of ~170–360 km (Fig. 1, Taylor 1979).

The presence of magma oceans, ponds, or magmaspheres during the early stages of planetary and planetesimal evolution has been documented, discussed, and widely accepted in the planetary scientific community for decades (Warren 1985; Elkins-Tanton 2012). Since sample return from the Apollo missions in the 1960s and 1970s, the lunar magma ocean (LMO) concept and the early formation of the Moon, *c*. 70–110 Myrs after the onset of nebular condensation (*c*. 4.50–4.45 Ga), have been central to lunar science (Fig. 2, Wood et al. 1970; Smith et al. 1970; Taylor and Jakes 1974; Taylor 1975; Wood 1975; Shirley 1983; Warren 1990; Heiken et al. 1991; Snyder et al. 1992; Papike et al. 1998; Longhi 2006; Nemchin et al. 2009; Elardo et al. 2011; Gaffney and Borg 2014; McLeod et al. 2014; Boyce et al. 2015). The return of anorthositic samples from the Apollo 11 mission spurred the notion that segregation of buoyant plagioclase during primordial cooling could give rise to a dominantly feldspathic crust on an early Moon (Fig. 2, Wood et al. 1970). From Green et al. (1971) and the study of returned Apollo 12 mare basalts, it was later demonstrated that plagioclase was not a residual phase during later mare basalt petrogenesis (*c*. 3.9–3.0 Ga), nor was it a crystallizing phase during low-pressure differentiation. This implied that the source reservoir to these mare basalts had experienced plagioclase depletion during an earlier event in the Moon's history: the extraction of plagioclase-rich feldspathic crust (Fig. 2, Philpotts et al. 1972). The clear demonstration of these distinct, yet mineralogically and geochemically complementary reservoirs on the Moon supported an early, lunar-wide differentiation event which today is known as the LMO hypothesis (Fig. 2). Since then, numerous efforts have been made to calculate the depth of a primordial magma ocean on a young Moon (Fig. 2) while accounting for observed

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mineralogy, heat transfer processes, and resulting crustal thicknesses. To date, estimates for the depth of the primordial LMO range from <20 km to the entire Moon (Binder 1976, 1982; Kopal 1977; Runcorn 1977; Warren 1985; Longhi 2006; Elkins-Tanton et al. 2011).

The degree and timescales of melting on an early-formed Moon depend on the source of primordial heat (Wetherill 1975, 1981; Alven and Arrhenius 1976; Sonett and Reynolds 1979). Potential sources include impact-induced accretionary heating (Safronov 1978; Wetherill 1975; Ransford 1982) and electromagnetic induction derived from early intense solar activity (Herbert et al. 1977), core formation, increased solar luminosity, short-lived radioisotopes, and tidal dissipation (Binder 1978). If accretion-derived heat was the main driver of melting within an early-formed Moon, understanding the conditions under which this occurred is fundamental to characterizing the nature of the primordial LMO. If accretion was slow, $>10^8$ years, this would have led to inefficient heat storage within the lunar mantle and an early-formed Moon would not have undergone differentiation and crystallization through magma ocean processes (Shearer and Papike 1999). If accretion occurred over short timescales, 10^6–10^8 years, and the radii of projectiles was >40 km (Warren 1985), then significant degrees of melting would have occurred. Moon formation through rapid accretion is therefore needed in order to establish whole Moon melting (e.g., Wood 1972, 1975). In this scenario, portions of the material from which the Moon would have accreted from in the post-impact terrestrial disk would have been above solidus temperatures. A whole-mantle magma ocean on a young Moon is therefore physically plausible (Elkins-Tanton et al. 2011 and references therein). Slower accretion on timescales of $>10^8$ years associated with smaller projectiles could have had the potential to establish a "magmafier" (a zone of partial melt: Shirley 1983; Warren 1985).

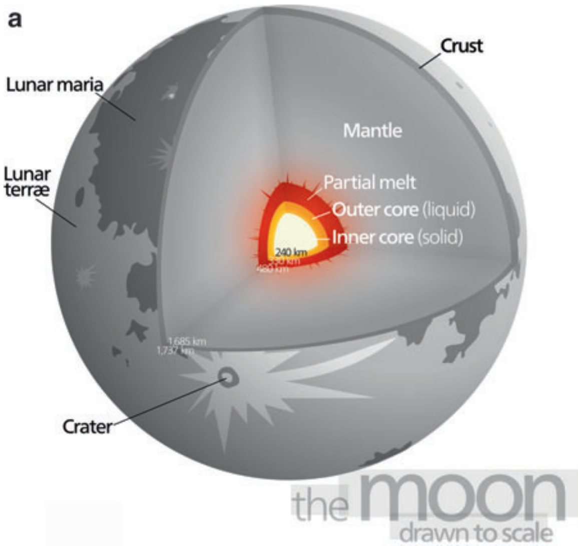
It should perhaps be noted that the LMO hypothesis has not been without challenges with several authors continuing to debate the validity of a global magma reservoir on an early-formed Moon (O'Hara 1970, 2000, 2001; Wetherill 1975; Walker 1983; Longhi and Ashwal 1985; O'Hara and Niu 2015, 2016).

Geochemical Constraints on the Depth of the LMO

In a simple LMO model, the generally accepted crystallization sequence is as follows: olivine → orthopyroxene ± olivine → olivine → clinopyroxene ± plagioclase → clinopyroxene + plagioclase → clinopyroxene + plagioclase + ilmenite (Shearer and Papike 1999). Several studies have aimed to further quantify this sequence with respect to (1) the initial size of the LMO using depths from 400 km (Snyder et al. 1992) to whole Moon melting (Elardo et al. 2011) and (2) bulk Moon compositions using the Taylor Whole Moon (TWM, Taylor 1982), Lunar Primitive Upper Mantle (LPUM, Longhi 2003), and other proposed bulk compositions (see compilation in Elkins-Tanton et al. 2011). The results of these studies are summarized in Fig. 3.

The composition of ferroan anorthosites (FANs), which make up a fundamental component of the feldspathic crust (Figs. 2 and 3), can be used in mass balance models to assess the degree of melting that would be needed in order to produce the plagioclase-rich lunar crust (Gast 1972; Walker et al. 1975; Warren 1985). If the FAN suite constitutes 50 % of the primordial upper lunar crust, and the bulk Moon contains 2.8 wt.% Al_2O_3 (assuming 100 % extraction of Al_2O_3 to form the lunar crust), 36 % of the Moon had to have melted. These constraints correspond to a minimum primordial LMO depth of ~250 km (Warren 1985).

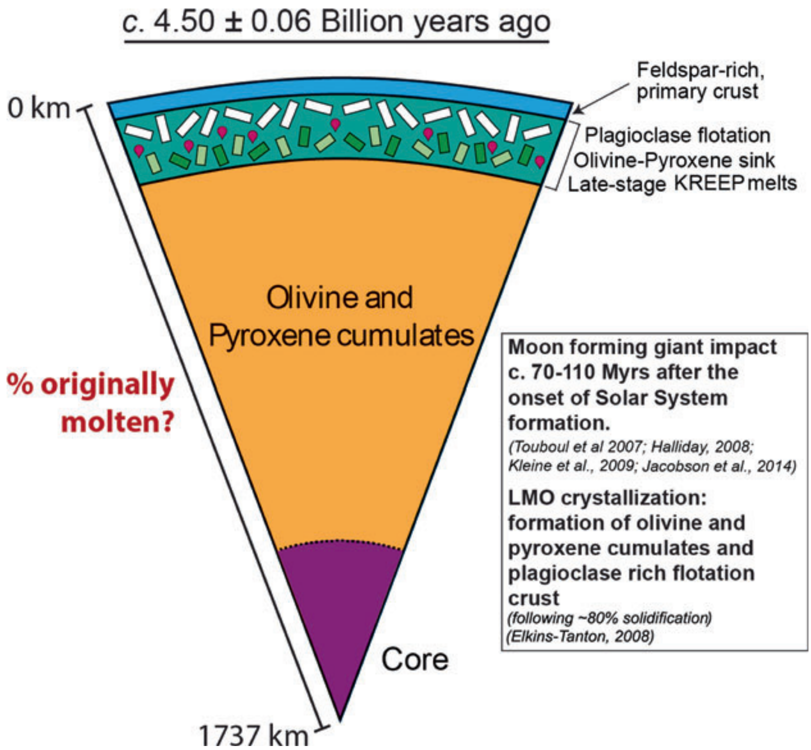
A greater primordial LMO depth of ~500 km was derived from mass balance models calculated from incompatible element enrichments in the lunar crust by Taylor (1978). From crustal potassium (K) abundances (~500 ppm; Metzger et al. 1977) and bulk Moon K concentrations (1,200 ppm), lunar crust 60 km thick would contain 46 % of the lunar K. In a simple LMO model where all the K is extracted from the lunar mantle and placed into the plagioclase flotation crust



Lunar Magma Ocean, Size, Fig. 1 Present day Moon consists, simply, of a crust, mantle and core (**a**), all of which are geochemically and geophysically distinct. The lunar crust today measures 55–75 km in thickness (Toksöz 1979, average: 34–43 km; Wieczorek et al. 2013). The oldest portions are anorthositic with 75–95% plagioclase and likely formed as a result of buoyant plagioclase ($\rho = 2.7$ g cm^{-3}) crystallizing from denser mafic magma and rising to form an early plagioclase flotation crust after ~80 vol. % solidification (Warren 1985; Elkins-Tanton and Bercovici 2014). The lunar mantle extends for c. 1200 km and is characterized by a seismic discontinuity at ~500 km (not shown here). The Moon's core is small (radius <350 km) and is thought to be at least partially molten (Williams et al. 2006). Estimates of the depth over which a primary Lunar Magma Ocean crystallized on an early formed Moon range from <20 km (Kopal 1977) to ≥1100 km (Longhi 2006; Elkins-Tanton et al. 2011) (Image from Wikipedia Commons: By Kelvinsong - CC BY 3.0, https://commons.wikimedia.org/w/index.php?curid=23350443. In (**b**) The nearside of the Moon is shown here with the crater Tycho and its ejection blanket prominent in the SW field of view, and the dark lunar mare basalts characterizing the northern hemisphere. Full Moon photograph taken 10-22-2010 and owned by Gregory H. Revera. Available through wikipedia commons at https://en.wikipedia.org/wiki/File:FullMoonZolo.jpg

(Figs. 2 and 3), a minimum initial depth of the LMO is constrained to ~320 km (Kaula 1977; Taylor 1978). From an arguably more realistic approach to modeling LMO crystallization by accounting for the K concentrations of the lunar mare basalts source regions (~500 ppm without the consideration of high-K and high-Al varieties), an initial lunar melting volume is constrained to 60 % and a corresponding LMO depth of 450 km. Allowing for the presence of interstitial, residual melts within the mare basalt source region (the olivine-pyroxene cumulate pile Figs. 2 and 3), this depth estimate can be revised to 500 km; however, these estimates strongly depend on lunar crustal thickness and K concentrations, and if the concentration in the lunar crust were two thirds of that discussed here, a melt volume is calculated at 56 % complemented by an initial LMO depth of 300 km (Taylor 1978). As noted in Taylor (1978), an LMO depth of 300 km raises a few issues. High bulk Al_2O_3 contents would be required in crustal source regions in order to establish the lunar crust as observed. This likely requires an increase in the concentrations of other elements, U and Th, for example. From this, constraints on associated heat flow are potentially exceeded (Taylor 1978).

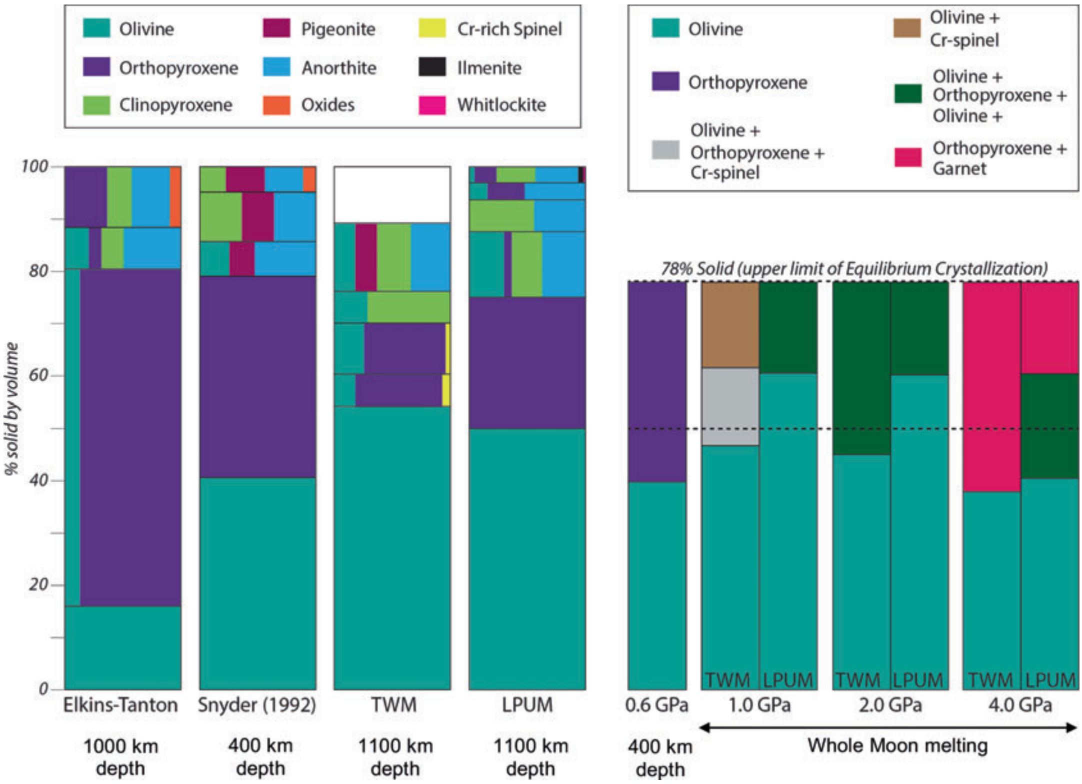
The transfer of heat from the lunar interior to space plays a fundamental role in the solidification of the primordial LMO, and the production of anorthositic lunar crust has the potential to prolong these solidification timescales, acting as a conductive lid (Figs. 2 and 3; Elkins-Tanton 2008). From Elkins-Tanton et al. (2011) a LMO extending to 1,000 km depth could account for



Lunar Magma Ocean, Size, Fig. 2 Crystallization of the Lunar Magma Ocean (LMO) on an early-formed Moon as result of the Moon-forming giant impact at c. 70–100 Myrs. Buoyant anorthitic plagioclase formed a primary feldspathic lunar flotation crust following ~80% solidification whilst earlier-formed, denser olivine and pyroxene formed cumulates at depth. Melts rich in incompatible elements; K (Potassium), REE (Rare Earth Elements) and P (Phosphorus) formed during the last stages of LMO crystallization. Image modified from McLeod et al. (2016). Estimates for the depth of a primordial LMO, and thus the degree of lunar melting on an early-formed Moon, range from <20 km to a completely molten Moon.

observed lunar crustal thicknesses, the ages of the anorthositic crust and crystallization timescales prior to the onset of plagioclase fractionation. Applying a bulk silicate composition for the Moon similar to that of Buck and Toksöz (1980), LMO solidification proceeded with olivine and pyroxene crystallization up to ~80 % solid before anorthite crystallized and formed the plagioclase-rich floatation crust (Figs. 2 and 3). This model produced a 40–50 km thick anorthitic lunar crust, broadly consistent with observations (Toksöz 1979; Wieczorek et al. 2013). In addition, this plagioclase-rich crust could have acted as an insulatory lid, thus prolonging solidification timescales of a primordial LMO with 80 % solidifying in ~1,000 years. The remaining 20 % would then crystallize over a ~10 Myr period, without the influence of tidal heating from Earth. In the presence of tidal heating, however, the current range in anorthositic crustal ages could be accounted for and is thus one potential mechanism though which reworking of primordial lunar crustal lithologies could have occurred (Meyer et al. 2010).

With the inference that all of the Al_2O_3 in the Moon is accommodated in the plagioclase crust, a LMO <400 km depth cannot generate enough Al_2O_3 to produce the observed lunar crustal thickness where as a LMO >800 km deep would generate crust in excess of what is observed



Lunar Magma Ocean, Size, Fig. 3 Lunar Magma Ocean mineral stratigraphy. The mineralogical characteristics of different LMO models are shown here with the width of the mineral boxes representing the fraction present of that particular mineral. During LMO crystallization after ~80% solid, plagioclase feldspar becomes a liquidus phase and forms the anorthitic crust. This phase is less dense than the earlier formed olivine and pyroxene cumulates floats to the top of the solidifying LMO. This figure does not illustrate therefore the depth at which the different LMO-derived minerals are present. This figure illustrates the depth at which different mineral phases form, assuming different LMO starting depths as noted below each model (Data sources: Snyder (1992, 400 km); Elardo et al. (2011, Whole Moon Melting) Elkins-Tanton et al. (2011, 1000 km); Rapp and Draper (2016); TWM – Taylor Whole Moon and LPUM – Lunar Primitive Upper Mantle, both modelled at 1100 km.

(Elkins-Tanton et al. 2011; Wieczorek et al. 2013). In a model where the LMO is 1,000 km deep, the excess alumina component can be modeled as being accommodated in Mg-Fe spinel pleonaste and interstitial melts. This scenario allows an LMO to be modeled with similar Al_2O_3 contents as previous studies but allows for a deeper magma ocean while still stabilizing an anorthositic crustal thickness consistent with geophysical observations (Fig. 3, Elkins-Tanton et al. 2011).

An LMO of 1,000 km depth in the Elkins-Tanton et al. (2011) study is deeper than previous estimates for LMO depths (Warren 1985; Kirk and Stevenson 1989; Hess and Permentier 1995) but consistent with the depths at which the green glasses of the mare basalts were likely generated (Longhi 2006). The chemical compositions of the mare basalts are inconsistent with melting of primitive lunar mantle reservoirs; they do not exhibit chondritic proportions of the refractory elements. From fractional melting models, Longhi (2006) assessed the generation of the mare basalts and the Apollo green glasses based on a range of previously published terrestrial and lunar mantle sources (Melson et al. 1976; Basaltic Volcanism Study Project 1981; Taylor 1982; Kinzler and

Grove 1992; Longhi 2003). The compositions of the mare green glasses are indicative of low Al_2O_3 mantle source regions (Longhi 1992). This is consistent with earlier extraction of Al_2O_3 (Figs. 2 and 3). From the polybaric fractional melting models of Longhi (2006), the source reservoirs to these glasses was constrained to between 700 and 1,000 km, with 1.2–2.4 wt.% Al_2O_3 for the olivine and pyroxene-rich reservoirs respectively. These calculations imply that LMO differentiation was fluid enough to facilitate the movement of differentiated material, from which a significant Al_2O_3 component had been extracted, to depths of 700–1,000 km. This could have been achieved by LMO crystallization over moderate depths of 400–500 km after which dense pyroxene ± ilmenite plumes sank through the lunar interior and transported this material to depth. However, the associated pressure-temperature time paths of the mare basalts are arguably inconsistent with this (Longhi 2006). It is therefore unlikely that significant proportions of undifferentiated lunar material were present above *c.* 1,000 km when the mare green glasses were generated. From this, the depth of primordial LMO differentiation is constrained to 700–1,000 km depth (Longhi 2006).

One of the fundamental aspects of the LMO model is the formation of the plagioclase-rich lunar crust (Fig. 2), and its occurrence has been applied to models of mass balance involving lunar crust formation with respect to wt. % Al_2O_3 contents (see earlier). With this approach and considering the following bulk Moon compositions, (1) Taylor Whole Moon (TWM) and (2) Lunar Primitive Upper Mantle (LPUM) from Taylor (1982) and Longhi (2003, 2006), respectively, and (3) a chondritic composition, lunar crustal composition, and thickness can be evaluated. This approach was also taken by Longhi (2006) and constrained the primordial LMO depth to 500 km. From the chondritic model, the wt. % Al_2O_3 composition of the lunar mantle source reservoir to the mare green glasses could not be reproduced, unless the Moon was initially completely molten, near perfect extraction of Al_2O_3 occurred during primary crust formation, and the average composition of the lunar mantle was lower in wt. % Al_2O_3 than the green glass source region. Given the petrogenetic history of the mare green glasses (as discussed earlier), a magma ocean deeper than 500 km is therefore required (Longhi 2006). A magma ocean at depths of ~1,000 km is supported by geophysical data (Taylor and Jakes 1974) with a partially molten zone potentially present below 1,000 km (Toksoz 1979) or a lower Mg# value for the lunar mantle >1,000 km (Goins et al. 1979). This latter point indicates that this depth could represent the extent of the LMO-derived cumulate pile or primitive lunar mantle (Fig. 2, Hollister 1975; Longhi 1981; Shearer and Papike 1999).

Crystallization of a primordial lunar magma ocean was also recently revisited by Elardo et al. (2011) through a two-stage crystallization approach, with an initially 100 % molten Moon followed by fractional crystallization of a residual magma ocean. In this scenario, an olivine-pyroxene cumulate pile is established from the Moon core-mantle boundary at ~1,440 to ~335 km (Fig. 3). Current work by Rapp and Draper (2013, 2014, 2016) aims to further evaluate LMO crystallization and assumes an initial LMO depth of 1,100 km (Fig. 3).

Geophysical Constraints on the Depth of the LMO

As mentioned earlier, estimates of the depth of the primary LMO range from an initially completely molten Moon, to <20 km, to layers of melt trapped between partially molten zones (magmafiers). From the above discussion, chemical constraints largely favor an LMO >500 km and potentially as deep as 1,000 km. From seismic data reported in Goins et al. (1981), the present-day lunar structure has been constrained to the following: (1) a lunar upper mantle from 60 to 400 km depth, (2) a transition zone between 400 and 480 km, and (3) a lower lunar mantle from 480 km to at least 1,100 km depth. Electrical conductivity data imply near-solidus temperatures >500 km depth, whereas seismic data imply incipient melting >1,000 km depth (Hood and Sonett 1982; Toksöz 1979; Nakamura et al. 1982; Warren 1985).

Following the Apollo missions, Kopal (1977) argued against the theory that the bulk Moon was ever covered by a magma ocean which extended to several 100 km, suggesting instead that lunar differentiation and crystallization was local (on the scale of the lunar maria) and that solidification occurred at depths of <20 km. These conclusions were based on the idea that the interior of the Moon was out of hydrostatic equilibrium and the report of observed differences in "moments of inertia of the lunar globe about its principal axes." The idea that a primordial LMO was local and/or as shallow as ~20 km has however been largely abandoned throughout the scientific literature. For example, large impact basins like Imbrium have been inefficient in sampling ultramafic cumulates from depth within the lunar interior and are rare throughout the lunar crust. Their absence, to date, can therefore be used to constrain a minimum LMO depth of at least 50 km (Shearer and Papike 1999).

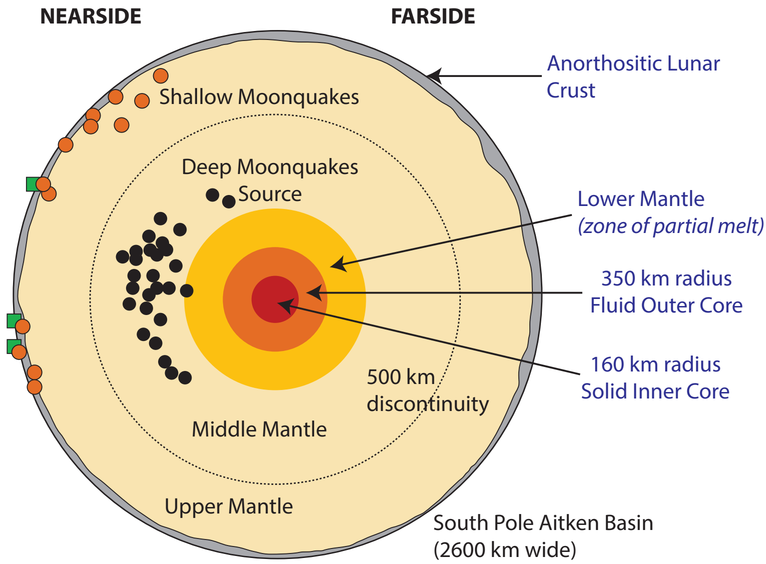
Following the Moon-forming Giant Impact between Proto-Earth and the Mars-sized object Theia, the Moon began to cool (Figs. 2 and 3). As planetary bodies cool their volumes decrease and as a result of this interior contraction, diagnostic tectonic features are predicted to form on a planets surface, such as have been observed on Mercury (Solomon and Chaiken 1976). However, for over 55 years, it has been known that the Moon lacks structural features diagnostic of significant volume change associated with contraction during primordial cooling (MacDonald 1960). This apparent lack of contractional features argues for limited secular cooling, implying a relatively cold initial Moon and thus limiting the depth of a primordial LMO (Elkins-Tanton and Bercovici 2014). For example, there are no observable large-scale, horizontal displacements associated with the rays or ridges emanating from Tycho (Fig. 1b). Evidence for vertical displacement on steep crater walls are observed but are likely associated with collision craters and hence not related to the internal cooling architecture of the Moon. The overall lack of faulting on the lunar surface, but principally strike-slip faulting, was interpreted by MacDonald (1960) as implying that the Moon's volume and surface dimensions have been almost constant since its formation.

As an early Moon cooled, compressive stresses would have developed in the lunar crust, but based on the relative lack of fault surfaces as discussed above, Solomon and Chaiken (1976) estimated a <1 km change in lunar radius since the termination of the late heavy bombardment, or lunar cataclysm, ~3.8 Ga. This near constant lunar radius has been accounted for in a scenario where contraction of the LMO was balanced by radioactive heating and associated expansion of an interior that was initially cold. From this, the LMO was inferred to never have been >200 km deep (±100 km, with a well-defined boundary separating a completely molten zone from a melting-absent zone). In a later study, Solomon (1980) updated this estimate to 300 ± 100 km. These initial studies which utilized the observation of the lack of fault scarps on the lunar surface were later challenged in Binder and Lange (1980) and Binder (1982). These subsequent studies advocated that as a result of contraction following a state of complete Moon melting, the decrease in the lunar radius was a factor of three lower than that applied by Solomon and Chaiken (1976) resulting in a change of the lunar radius of >1 km. The fault scarps of the lunar highlands have been suggested as supporting evidence of this lunar contraction. In addition, these fault scarps have been interpreted as evidence for a once totally molten Moon (Binder 1982). In later calculations, Kirk and Stevenson (1989) suggested that the volume increase associated with heating and melting of the primitive lunar interior, below an LMO of 600 km depth, had the potential to counterbalance the contraction that would be otherwise expected on a cooling Moon. However, the timing of this predicted melting event within the Moon has since been associated with mare basalt generation, and mare basalts do not represent melts derived from undifferentiated, primitive lunar material (as discussed earlier).

More recently, gravity data reported by the GRAIL (Gravity Recovery and Interior Laboratory) mission, and discussed in Elkins-Tanton and Bercovici (2014), was applied to solidification models of a primordial LMO which demonstrated potential expansion, not contraction. This inferred expansion is attributed to dike intrusions within the lower lunar crust, as imaged by GRAIL (Andrews-Hanna et al. 2013). However, models presented in Elkins-Tanton and Bercovici (2014) demonstrated that the degree of expansion or contraction on a young Moon cannot to be used to constrain the depth of a primordial LMO. Models reported a wide range of values for the amount of expansion and contraction for a single LMO depth. In addition, no lunar crust exists that would record the earliest volume changes as the plagioclase flotation crust, which typifies the lunar crust, did not form until ~80 vol.% solidification of the LMO (Fig. 3; Elkins-Tanton et al. 2011).

Seismometers were deployed during each of NASAs Apollo missions in order to investigate the Moon's interior structure, and the seismicity detected is demonstrably distinguishable from that measured on Earth (Wieczorek et al. 2006). Deep moonquakes, ~700–1,200 km (Fig. 4), occur monthly and display near identical wave forms indicating a common source region (Lammlein et al. 1974; Nakamura 1978). This monthly signature indicates a relationship with the tides raised by the Earth and Sun. Shallow moonquakes, ~50–220 km (Fig. 4), are less abundant and average five events per year (Nakamura 1977; Khan and Mosegaard 2002). Seismic velocity models of the Moon are characterized by a velocity increase at ~500 km (Fig. 4, Nakamura et al. 1982; Nakamura 1983; Khan et al. 2000; Khan and Mosegaard 2002). At this depth, P-wave velocities were shown to increase from 7.46 ± 0.25 km/s to 8.26 ± 0.40 km/s, and S-wave velocities were shown to increase from 4.25 ± 0.10 km/s to 4.65 ± 0.16 km/s (Nakamura 1983). This discontinuity has been attributed to change in composition and a potential phase transition, possibly the spinel to garnet transition which may occur between 300 and 550 km (Green and Ringwood 1967; Kuskov 1995, 1997). Khan et al. (2006) suggested this transition could occur at depths as shallow as ~200 km. While potentially consistent with the ~500 km discontinuity, this phase change would result in a small velocity change ($\leq$0.1–0.4 km/s for S-waves, Hood and Jones 1987; Nakamura 1983) and could not therefore account for the observed ~500 km discontinuity. A change in composition is therefore required (Wieczorek et al. 2006). One potential explanation is that the mantle was compositionally homogeneous and primary melting and differentiation of the LMO only occurred to depths of ~500 km. This process would have to efficiently partition Al_2O_3 from the upper mantle into the lunar crust leaving behind a primordial lower lunar mantle. However, the implied compositional change at ~500 km for given bulk Al_2O_3 contents in the lower lunar mantle, which controls the proportion of garnet present, cannot sufficiently account for the velocity change (Hood and Jones 1987; Mueller et al. 1988).

The 500 km discontinuity may also represent primary lunar mantle zonation with aluminous phases present in the lower lunar mantle (Fig. 4). If the primary LMO differentiated over a 500 km depth, with this structure, seismic velocity increases compatible with velocity models are possible (Nakamura 1983; Hood and Jones 1987; Mueller et al. 1988). Both of these scenarios invoke an initial LMO at a depth of 500 km. However, this would be inconsistent with Moon-forming Giant Impact scenarios which demonstrate that the Moon could have potentially formed in a 100 % molten state (Pritchard and Stevenson 2000; Canup 2004; Elkins-Tanton et al. 2011). The 500 km discontinuity could therefore be intrinsic to the primordial crystallization of a primary, global LMO. Early crystallization of the LMO is dominated by olivine and subsequent orthopyroxene crystallization to form the olv-opx cumulates (Figs. 2 and 3). The ~500 km discontinuity could therefore represent the boundary between olivine-dominated and orthopyroxene-dominated LMO cumulates (Wieczorek et al. 2006). However, the potential problem with these three scenarios is that post LMO density-driven cumulate overturn would act to eradicate features associated with primary crystallization (Snyder et al. 1992; Elkins-Tanton et al. 2011). One other possibility is that this discontinuity represents the maximum depth of the mare basalt source region. From the thermal



Lunar Magma Ocean, Size, Fig. 4 Schematic illustration of the Moon's internal architecture highlighting the location of deep moonquakes (800–1000 km), shallow moonquakes (50–220 km), the ~500 km discontinuity as revealed by changes in P and S wave velocities, and a likely Crust-Mantle-Core structure. The crustal thickness on the Moon varies from 60 to 100 km and is schematically oversimplified here to illustrate the relatively thinner crust at the location of the South Pole Aitken Basin, the Moon's largest crater (~2600 km in diameter, ~13 km in height from crater floor to rim). The size of the solid inner core is poorly constrained but thermal considerations support at least partial solidification (Wieczorek et al. 2006). The solid green squares represent Apollo landing sites. Image has been modified from Wieczorek et al. (2006).

evolution model of Wieczorek and Phillips (2000), lunar mantle melting is constrained to beneath the Procellarum KREEP Terrane (PKT) with melting depth increasing with time from ~200 to ~600 km. The lateral extent of this mantle melting is currently not well constrained due the position of three of the four Apollo seismic station being located within the PKT.

In order to further constrain the lunar structure at present day (Figs. 1a and 4), Khan et al. (2000) reported results from inverse Monte Carlo calculations applied to P- and S-wave arrival times. From the lunar surface to 45 ± 5 km depth, velocity increased to the base of the crust. Below the crust, a constant velocity upper mantle was inferred over a 560 ± 15 km depth indicating a homogenous upper lunar mantle (Fig. 4). At *c*. 560 km depth, a prominent increase in P- and S-wave velocities was observed from 8.5 ± 1.5 km/s and 4.8 ± 1.1 km/s, respectively, to 9.9 ± 1.9 km/s and 5.9 ± 0.9 km/s, respectively. These observed increases are broadly consistent with data reported in Nakamura (1983) where P-wave velocities were shown to increase from 7.46 ± 0.25 km/s to 8.26 ± 0.40 km/s and S-wave velocities to increase from 4.25 ± 0.10 km/s to 4.65 ± 0.16 km/s. At ~780 km, Khan et al. (2000) also observed increases in P- and S-wave velocities from 9.0 ± 1.9 km/s and 5.5 ± 0.9 km/s to 11.0 ± 2.1 km/s and 6.0 ± 0.7 km/s, respectively, and a high velocity zone between 800 and 1,000 km. This deeper zone coincides with the source region for deep moonquakes (Fig. 4).

From Khan et al. (2004), the observation of strong shear wave arrivals, diagnostic of deep moonquakes, was hypothesized to indicate the presence of partial melts within the lunar interior at present day. These moonquakes had bottoming depths of ~1,100 km. No prominent shear wave arrivals were associated with waves that bottomed deeper and were therefore inferred to have traveled through a deeper partially molten zone (Fig. 1, Nakamura et al. 1973; Nakamura 2005; Weber et al. 2011; Khan et al. 2014). Whether this partially molten zone is associated with the lunar core or the lunar deep mantle is currently debated (Figs. 1 and 4, Williams et al. 2001, 2012; Khan et al. 2004; Nimmo et al. 2012; Harada et al. 2014). From Khan et al. (2014), the melt zone was recently modeled at depths of ≥1,200 km at 100–150 km thickness and was interpreted as a molten region surrounding the lunar core with temperatures at 1,600–1,800°K at 1,300 km depth. Crucially, the presence of this melt layer within the Moon at present-day places fundamental constraints on the thermal evolution of a primordial LMO including (1) the potential sinking of late-formed, ilmenite-bearing cumulates through the lunar interior and the transport of radiogenic heat-producing elements and (2) these cumulates forming a stable layer at depth.

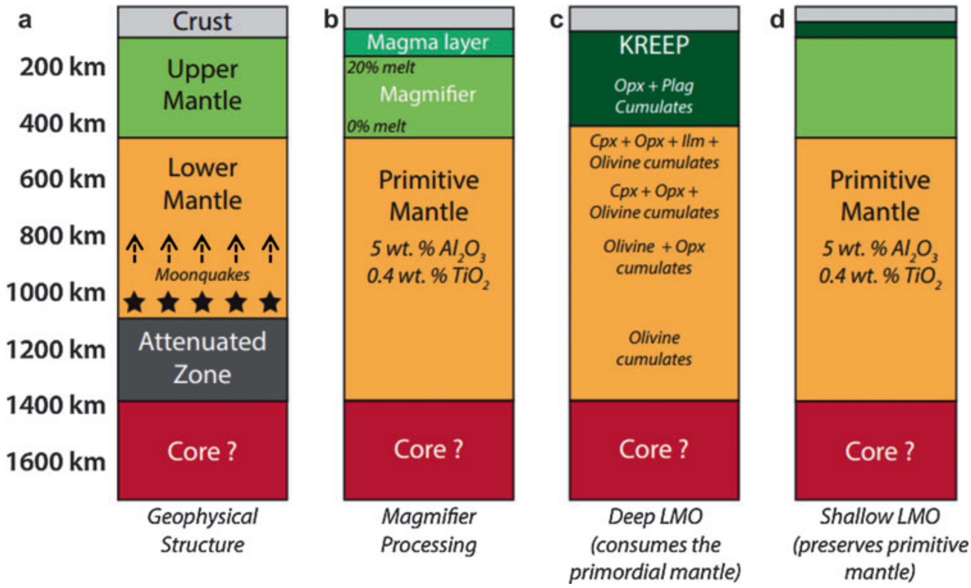
Figure 5 shows a schematic summary of the different geochemical and geophysical constraints that have been discussed regarding the depth of a primordial LMO. From a geophysical perspective, a significant change in seismic velocity occurs at ~500 km below which there is a zone of deep moonquakes (~700–1200 km) and a potentially molten zone surrounding the lunar core (Panel (a), Fig. 5, (Lammlein et al. 1974; Nakamura 1978). Broadly coincident with the ~500 km discontinuity are geochemical constraints that advocate for a shallow LMO <500 km ± magmafier processes (Panels (b) and (d), Fig. 5, Shirley 1985; Taylor 1978; Warren 1985). In both of these scenarios, the lower lunar mantle is characterized by primordial mantle compositions (~5 wt.% Al_2O_3), i.e., undifferentiated. Panel (c) in Fig. 5 represents a scenario in which a primordial LMO was ~100 km, as derived from certain geochemical constraints: the depth of the source region to the mare basalts, for example (Longhi 2006). In this scenario, no undifferentiated, primitive lunar mantle exists at depth.

Conclusions

To date, differentiation and crystallization of a magma ocean on an early-formed Moon are arguably the most elegant (Shearer and Papike 1999) framework in which current lunar lithological, mineralogical, geochemical, geochronological, and geophysical data should be evaluated. Constraints on the extent of melting associated with a primordial LMO on a young planetary body have however, not been consistent, ranging from <20 km, to shallow zones of partial melting to whole Moon melting. The most significant difference between a shallow and a deep LMO is whether or not the lower lunar mantle today represents undifferentiated primitive material, i.e., a shallow LMO (<500 km), or is characterized by olivine-pyroxene cumulates as a result of early differentiation, i.e., a deep LMO (~1,000 km).

From a geochemical perspective, the wt. % Al_2O_3 and K concentrations of lunar crustal rocks can be used to calculate the depth of melting using mass balance calculations and constraints on the thickness of the plagioclase-rich lunar crust. Depth estimates range from 250 km (Warren 1985) to 500 km (Taylor 1978) and support a shallow LMO. Elkins-Tanton et al. (2011) argued that if all the Al_2O_3 is accommodated in the anorthitic flotation crust, an LMO <400 km would not be able to generate observed crustal thicknesses. Instead, a deep LMO at ~1,000 km is favored and is consistent with the calculated source depth for the green glasses within the Apollo mare basalts (700–1,000 km, Longhi 2006). Recent and current efforts to model the crystallization sequence of a primordial LMO have invoked initial depths of 1,000–1,100 km and whole Moon melting (Elardo et al. 2011; Elkins-Tanton et al. 2011; Rapp and Draper 2013, 2014, 2016).

From a geophysical perspective, the lack of contractional features on the lunar surface has been used to imply a relatively cold initial Moon. The lack of these physical structures has been used to infer that the LMO was never >300 ± 100 km deep (Solomon and Chaiken 1976; Solomon 1980). In a later study, Kirk and Stevenson (1989) advocated that volume expansion below a depth of 600 km in the LMO could potentially counterbalance expected contraction. However, a more recent study by Elkins-Tanton and Bercovici (2014) argued that constraints on



Lunar Magma Ocean, Size, Fig. 5 Summary LMO schematic showing (**a**) Internal architecture of the Moon as constrained from geophysical studies with a zone at ~800–1000 km characterized by deep Moonquakes. (**b**) Processing of the lunar mantle through magmafier processes where the whole Moon is not molten. The shallow upper lunar mantle is characterized by zones of partial melting and primordial lunar mantle remains at depth (magmafiers). (**c**) Deep LMO to depths of >1000km where the lower mantle is characterized by Olivine ± Orthopyroxene ± Clinopyroxene ± Ilmenite cumulates and the upper mantle by late-stage KREEP melts and Orthopyroxene + Plagioclase cumulates. In this scenario no primordial lunar mantle remains. (**d**) Shallow LMO (~400–500 km) where the differentiated upper mantle is characterized by Olivine ± Orthopyroxene ± Clinopyroxene ± Ilmenite cumulates. At depth, > 500 km, primordial lunar mantle is present (Image is modified from Shearer and Papike (1999))

the extent of lunar expansion or contraction could not be used to calculate the depth of a primordial LMO. On the present-day Moon, one of the most characteristic features of the lunar interior is the 500 km discontinuity. The distinct P- and S-wave velocity change at ~500 km has been suggested as representing the extent of the LMO, which would be consistent with geochemical constraints supporting a shallow LMO. However, other interpretations exist including preservation of a primary lunar mantle zonation feature, the boundary between olivine and pyroxene cumulates, and the mare basalt source region (Wieczorek et al. 2006).

From the discussion presented here, the debate of whether the primordial LMO was shallow (<500 km) or deep (≥1,000 km) remains active. The scenarios in which an early magma ocean was <20 km, or that due to slow rates accretion the shallow Moon was characterized by magmafiers, have largely been removed from discussion. Future results from LMO crystallization studies are eagerly anticipated (Draper et al. 2016), and the potential for expanding a lunar seismic network outside of the Apollo landing site is encouraged (Wieczorek et al. 2006). From these approaches, further insights into Moon's structure, thermal evolution, and potential depth of a primary LMO will be provided.

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