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Mantle

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Introduction

The Earth-Moon system resulted from a giant impact between proto-Earth and a small Mars-sized object (Theia). Solidification of the coalesced material resulted in internal differentiation and the establishment of a core-mantle-crust structure (e.g., Smith et al. 1970; Wood et al. 1970; Gagnepain-Beyneix et al. 2006; Weber et al. 2011; Trønnes et al. 2019). As it exists today, the lunar mantle is the likely result of differentiation of a Moon-wide magma ocean, the lunar magma ocean (or LMO; Smith et al. 1970; Wood et al. 1970). The depth of this primordial LMO has been extensively investigated over the past half century with estimates ranging from several 100 km (i.e., a shallow LMO) to scenarios in which the whole Moon was completely molten following the giant impact event (e.g., Minear and Fletcher 1978; Charlier et al. 2018; Steenstra et al. 2020). Despite a lack of consensus regarding depth, most Moon-forming models do at least agree that the LMO existed Moon-wide. These models also generally agree that the onset of LMO solidification was marked by crystallization and settling of dense Mg-rich olivine and orthopyroxene, thus establishing early mafic mantle cumulates at the base of the LMO (Fig. 1, e.g., Charlier et al. 2018; Li et al. 2019; Moriarty III et al. 2021). As cooling and crystallization continued, clinopyroxene and anorthitic plagioclase feldspar became liquidus phases. The relatively denser pyroxene crystals continued sinking, while the relatively less dense feldspar crystals rose (or "floated") to establish a primordial flotation crust (e.g., Walker and Hays 1977; Warren 1990; Dygert et al. 2017). As solidification progressed, the residual LMO melt became relatively enriched in incompatible elements (e.g., K, REEs, P, Th, and U,) leading to the generation of an urKREEP reservoir (Fig. 1; Warren and Wasson 1979), and KREEP-rich clinopyroxenes. Further differentiation led to the crystallization of trace phosphates, and the formation of oxides, along with dense, late-stage cumulates, including ilmenite-bearing cumulates (IBCs; e.g., Zhao et al. 2019). The establishment of these late-stage cumulates prior to complete LMO solidification is proposed to have then initiated overturn (or "gravitational restructuring"; Moriarty III et al. 2021) of the lunar mantle due to density instabilities. This reorganization may have involved as much as 50–70% of the IBCs sinking diapirically toward the earlier-formed mafic mantle cumulates, largely without disturbing the shallow urKREEP reservoir (Zhao et al. 2019). These early LMO processes were later followed by partial melting of the olivine and pyroxene cumulates to produce the younger mare basalts. Collectively, these

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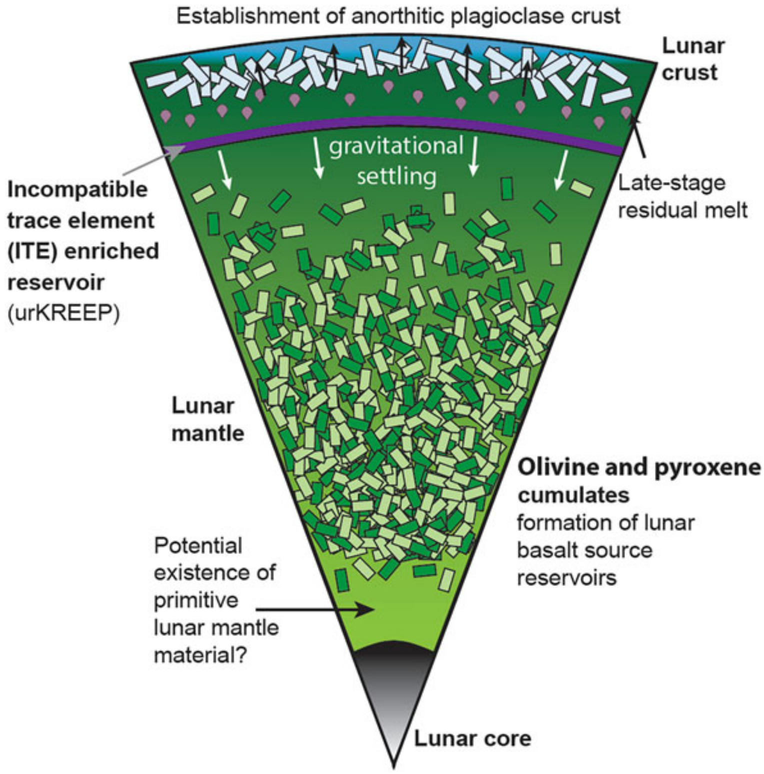
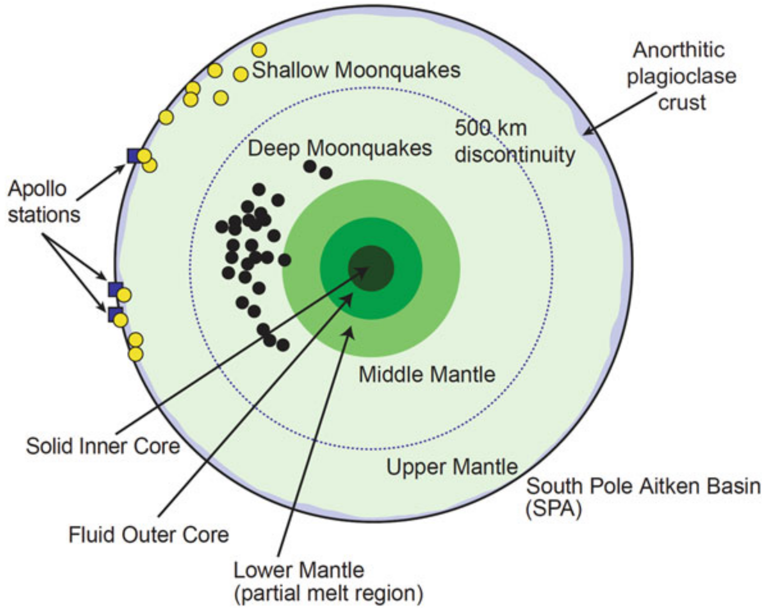
Mantle, Fig. 1 Summary of LMO processes leading to the establishment of olivine-pyroxene (± ilmenite) -bearing cumulates at depth, an anorthitic, primary flotation crust, and an incompatible trace element-enriched reservoir (urKREEP)



events have led to the broadly stratified lunar mantle that exists today.

Geophysical Constraints

To date, no samples of the deep lunar interior have been unequivocally confirmed to exist either as exposed sections on the Moon, or within the sample collections (Moriarty III et al. 2021; Qian et al. 2021). Knowledge of the Moon's interior structure is therefore primarily derived from geophysical observations (see summary in Fig. 2). The collection of lunar geophysical data began during the Apollo program through the Apollo Lunar Surface Experiment Package (ALSEP). A variety of datasets were collected both via active source experimentation and passive listening and are summarized in detail by Nunn et al. (2020). Instruments deployed included a surface gravimeter, heat-flow probes, retroreflectors, seismometers, and surface magnetometers (e.g., Wieczorek 2009; Garcia et al. 2019). The seismometers yielded the highest resolution data and operated from 1969 to 1977 during which they detected 28 shallow moonquakes, ~7000 deep moonquakes, and ~1800 meteoroid impacts (Wieczorek 2009; Garcia et al. 2019). As summarized in Civilini et al. (2021), there are four, naturally occurring, primary sources of lunar seismicity. These include moonquakes at shallow and deep levels (see Fig. 2), impacts, and thermal events associated with significant changes in temperature between day and night conditions. Nunn et al. (2020) and Garcia et al. (2019) further distinguish impacts into (1) artificial impacts on the lunar surface and (2) meteoroid strikes. Interestingly, at the time of the Apollo missions, scientists did not expect to catalog moonquakes, hence their existence was a discovery in itself (see Nakamura 2015, for a historical summary). From the recent work of Watters et al. (2019), the origin of eight shallow moonquakes detected via ALSEP was attributed to fault activity on young thrust faults, with six of these occurring during times when the Moon was close to its apogee (and thus likely



Mantle, Fig. 2 Simplified cross section of the lunar interior. Note the presence of shallow and deep moonquakes on the lunar nearside and the largest impact basin in the Solar System, the South Pole-Aitken Basin, on the lunar farside. (Modified from Wieczorek et al. 2006)

experiencing maximum compressional stress). In contrast, deep moonquakes at ~800–1200 km depth were the most commonly detected seismic events during ALSEP. These events have since been correlated to tidal stresses and to an increase in the brittle-ductile transition temperature within the lunar mantle (Kawamura et al. 2017).

From data acquired on the lunar surface during the Apollo missions, scientists have also determined that the lunar mantle is probably anhydrous and has temperatures well below the mantle solidus for depths <1000 km (e.g., Karato 2013; Khan et al. 2014; Kuskov and Kronrod 2009; Longhi 2006; Garcia et al. 2019). Garcia et al. (2019) provide a recent review of the understanding of lunar structure at the time of writing. Current models suggest that the lunar mantle extends from the base of the crust at ~40 km depth (a crust which includes a ~10 km zone of brecciation at a depth of ~30 km), to the potential small lunar core at ~1400 km depth (Gagnepain-Beyneix et al. 2006; Wieczorek et al. 2013; Khan et al. 2014; Garcia et al. 2019). Seismic velocities for P and S waves in the lunar upper mantle range from 7.6 to 7.8 km/s and 4.35 to 4.45 km/s, respectively, while at depths associated with deep moonquakes (>740 km), seismic velocities increase to ~8.15 km/s and ~4.15 km/s, respectively (Gagnepain-Beyneix et al. 2006). Since the deployment of the Apollo network, scientists have debated the existence of a midmantle discontinuity corresponding to an upper depleted mantle of potential pyroxenite composition, a lower, magnesian-rich, primitive mantle, and a discontinuity at ~1200 km depth (≥1600 °C) where melt may reside (e.g., Nakamura 1983, 2005; Gagnepain-Beyneix et al. 2006; Nimmo et al. 2012; Khan et al. 2014; Wieczorek et al. 2006; Garcia et al. 2019).

Remote Sensing Constraints

From high resolution gravity data retrieved via the GRAIL (Gravity Recovery and Interior Laboratory) spacecraft, numerous regions of mass concentration associated with large positive gravity anomalies have been identified. These so-called “mascons” are associated with the Moon’s impact basins including those that are, and are not, infilled with basaltic lava (Melosh et al. 2013). From the recent work of Zhao et al. (2021), 3-D inversion of GRAIL data was used to propose that following an impact event and the collapse of a transient crater, lunar mantle material upwelled to fill the crater and establish high-density lithologies beneath the Moon's basins as observed today. The materials that comprise impact basin rings have the potential to originate from a variety of depths within the lunar interior with Lemelin et al. (2019) recently demonstrating that the innermost rings are often dominated by anorthosite. However, from their detailed mineralogical assessment of data acquired via the SELENE (Kaguya) Multiband Imager, a "mantle component" was also proposed to exist with ultramafic material potentially also present below the single pixel scale (<62 m in this case).

The largest impact structure on the Moon is the South Pole-Aitken (SPA) Basin, which has a 2900 km diameter stretching from the lunar south pole up to Aitken Crater at the equatorial region of the lunar farside. At its deepest point, the SPA Basin is approximately 8 km deeper than the average lunar surface and 12 km deeper than the surrounding feldspathic highlands crust (Smith et al. 2010). It has therefore been proposed that either the SPA-forming impact event or a subsequent crater generated within the SPA may have exposed lunar mantle material. However, to date, no such lithologies have been confirmed. From analysis of data retrieved via the Moon Mineralogy Mapper (M^3) on board the Lunar Reconnaissance Orbiter (LRO), an olivine-rich exposure was identified within the central peak ring of Schrödinger Crater; however, Kramer et al. (2013) argue that this exposure represents either a dunitic or troctolitic intrusion. Unloading of crust in this region due to the SPA basin-forming impact has occurred, thus creating a gravity anomaly in the SPA region (Pieters et al. 1997; James et al. 2019). This uplifting has likely led to more mafic lithologies existing within the SPA center (e.g., Borst et al. 2012). Nonetheless, no lunar mantle-derived materials have been unambiguously confirmed to exist in the lunar surface via remote sensing (Moriarty III and Pieters 2018).

Lithological Constraints

In addition to potential exposure via impact processes, physical samples of lunar mantle lithologies may have also been transported to the surface as xenoliths in mare basalts (e.g., Shearer et al. 2015). However, Shearer et al. (2015) note that even examples of deep crustal material are rare in the lunar collection and mantle-derived material has never been found.

In terms of the returned Apollo, Luna, and Chang'e-5 sample collections, and the lunar meteorite suites, mantle lithologies have not been unambiguously confirmed. However, dunitic clasts have been identified within several samples and have been proposed as potentially representing the lunar mantle (Moriarty III et al. 2021). In terms of samples collected on the lunar surface, Shearer et al. (2015) provide a comprehensive overview of olivine-rich samples including the dunite within the high-Ti Apollo mare basalt sample 74,275, and the dunitic clasts and olivine megacrysts within the Apollo 17 high-Mg rocks (samples 72,415–72,418). At present, the geochemical signatures of these samples are inconsistent with an unequivocal lunar mantle origin but may instead represent (primary LMO) cumulate olivine and/or Mg-suite magmatism (e.g., Dymek et al. 1975; Elardo et al. 2011; Wang et al. 2015). Dunite fragments have also been identified in several of the Apollo 14 breccias, the origins of which have been attributed to the deep lunar crust, potentially as cumulate materials (e.g., Warren et al. 1987). From the meteorite collections, a dunitic clast within impact melt breccia sample Northwest Africa (NWA) 11,421 has most recently been proposed as being derived from the lunar mantle, in particular from a region within the mantle which has been significantly tectonized and recrystallized (Treiman and Semprich 2019, 2021).

Geochemical Constraints

Additional insights into the structure and composition of the lunar mantle have been gained via geochemical investigations of the lunar mare basalts and volcanic glasses. These materials originated as partial melts of the mantle and thus have been used to model LMO crystallization and assess the establishment of geochemically distinct reservoirs within the lunar interior (e.g., Neal

2001; Shearer et al. 2006; Taylor et al. 2006; Wieczorek et al. 2006; Elardo et al. 2011; Tartèse et al. 2019). As summarized by Wieczorek et al. (2006) and Tartèse et al. (2019, and references therein), the lunar mare source regions are generally low in alkali and siderophile elemental abundances, have nonchondritic Sm-Nd and Lu-Hf isotopic compositions, and exhibit low-volatile budgets relative to their terrestrial counterparts, including low-oxygen fugacities (see McCubbin et al. 2015 for a review of lunar volatile budgets). Nonchondritic trace element abundances indicate that the lunar mantle experienced differentiation early in the Moon's history. As described previously, plagioclase feldspar became a liquidus phase during LMO solidification and crystallized out, establishing the anorthositic primary flotation crust. As the Moon is a reducing environment, europium (Eu) exists primarily in the divalent state and therefore readily substitutes for Ca in the crystal structure of anorthite; thus, the crystallization of an anorthite-rich feldspathic crust depleted the lunar mantle in Eu. This process is evidenced by a strong negative Eu anomaly in the chondrite-normalized rare earth element patterns of the mantle-derived lunar mare basalts and volcanic glasses (e.g., Shearer et al. 2006; Grove and Krawczynski 2009). Beyond this, the mantle is laterally (and vertically) geochemically heterogeneous (e.g., Grove and Krawczynski 2009; Shearer et al. 2006; Elkins-Tanton et al. 2011; Hallis et al. 2014). This heterogeneity is exemplified by the variable titanium (Ti) contents of the mare basalts which range in TiO_2 from <1 wt. % to >12 wt. %: the very low-Ti, low-Ti, and high-Ti basalts (e.g., Neal and Taylor 1992; Shearer et al. 2006), while the volcanic glasses range from <0.25 wt. % to >16 wt. % (Shearer and Papike 1993; Grove and Krawczynski 2009). The relatively low Ti contents of some of these lunar materials reflect partial melting of early formed lunar mantle cumulates which were dominated by olivine and pyroxene, and generally lacked Ti-bearing phases (e.g., ilmenite). Dense Ti-bearing oxide minerals crystallized during the later stages of LMO differentiation forming Ti-rich source regions, i.e., IBCs, within the lunar mantle (e.g., Shearer et al. 2006; Elardo et al. 2011; Elkins-Tanton et al. 2011, and references therein). This ultimately led to the mantle restructuring described earlier and generating lateral geochemical heterogeneities within at least the shallow mantle (e.g., Elkins-Tanton et al. 2011; Hallis et al. 2014). A discussion of the bulk composition of the whole lunar mantle and the challenges associated with determining this are provided in Shearer et al. (2006), Taylor et al. (2006), Tartèse et al. (2019).

Another component of bulk lunar mantle chemistry (and mineralogy) which has received a lot of attention over the past half century is the role (or not) of garnet (e.g., Anderson 1975; Neal 2001; Jing et al. 2022). As a phase rich in aluminum (Al), it plays a critical role in governing the Moon's Al budget, which is critical for anorthitic flotation crust production. As summarized in Jing et al. (2022), various approaches have been previously taken to evaluate the presence of garnet at depth within the Moon from high-pressure experiments to seismology, to geochemical modeling. Using an experimental approach and a fixed starting bulk Moon composition, Jing et al. (2022) recently demonstrated the stability of garnet at depth, specifically in the lower mantle at up to 20 wt. % in deep mantle cumulates. From this, the REE characteristics of high KREEP compositions could also be matched during late stage LMO crystallization. While no garnet has been physically identified in lunar samples, the geochemical signatures of lunar volcanic products, namely a subset of the volcanic glasses, are consistent with the presence of garnet at depth (Neal 2001). In this case however, a primitive garnet-bearing reservoir which remained unmelted during the LMO is proposed as the source.

Additional geochemical complexity within the lunar interior is associated with the proposed urKREEP reservoir (Fig. 1; Warren and Wasson 1979; Warren 1985). This incompatible trace element-enriched reservoir is proposed to have formed during the final stages of the LMO, specifically following ~99% solidification, and is understood to exist at the base of the lunar crust (e.g., Wieczorek et al. 2006). While the spatial extent of this reservoir at depth remains poorly constrained, the chlorine isotopic compositions of lunar apatites were recently used to propose that the urKREEP reservoir experienced significant degassing as a result of the primordial lunar crust being punctured via impact events prior to final LMO solidification (Barnes et al. 2016). Evidence for the existence of the urKREEP reservoir is derived from the KREEP-rich nature of select mare basalt suites. From bulk-elemental geochemistry, it is evident that some lunar mare magmas likely assimilated KREEPy material (i.e., Apollo 11 group A basalts, Apollo 14 very high potassium basalts, and Apollo 15 KREEP basalts); however, the full extent of KREEP assimilation by lunar basaltic and plutonic magmas remains unknown (e.g., Taylor et al. 2006; Shearer et al. 2015, Elardo et al. 2020). Thus, at the time of writing it is not fully understood whether the trace element enrichment of some mare basalt groups represents contamination by urKREEP-derived material, a distinct mantle source generated during gravitational restructuring, or a combination of both.

Cross-References

- ▶ Differentiation of the Lunar Interior
- ▶ Internal Structure/Mantle Motions of the Moon
- ▶ Mantle Convection

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