Title: Excavation of the lunar mantle by basin-forming events on the Moon

Abstract: Global maps of crustal thickness on the Moon, derived from gravity measurements obtained by NASA’s Gravity Recovery and Interior Laboratory (GRAIL) mission, have shown that the lunar crust is thinner than previously thought. Hyperspectral data obtained by the Kaguya mission have also documented areas rich in olivine that have been interpreted as material excavated from the mantle by some of the largest lunar impact events. Numerical simulations were performed with the iSALE-2D hydrocode to investigate the conditions under which mantle material may have been excavated during large impact events and where such material should be found. The results show that excavation of the mantle could have occurred during formation of the several largest impact basins on the nearside hemisphere as well as the Moscovienne basin on the farside hemisphere. Even though large areas in the central portions of these basins were later covered by mare basaltic lava flows, surficial lunar mantle deposits are predicted in areas external to these maria. Our results support the interpretation that the high olivine abundances detected by Kaguya spacecraft could indeed be derived from the lunar mantle.
The Moon is a differentiated body composed of a geochemically distinct crust, mantle and core. Analysis of lunar basalts indicates that the mantle is composed predominantly of the minerals olivine and pyroxene, in contrast to the crust that is composed mostly of anorthite. Direct sampling of the lunar mantle, or observation of mantle exposures on the surface from orbit, would provide important constraints towards understanding not only early differentiation processes, but also the mechanism by which the Moon formed.

Global maps of crustal thickness on the Moon, recently derived from gravity measurements obtained by NASA's Gravity Recovery and Interior Laboratory (GRAIL) mission, have shown that the lunar crust is thinner than previously thought. At a similar time, hyperspectral data obtained by JAXA's Kaguya mission have documented areas on the surface of the Moon that are rich in olivine. These data have been interpreted as material excavated from the mantle by some of the largest lunar impact events. The most prominent olivine-rich areas detected surround the Crisium and Moscoviense basins. These basins also have a near-zero crustal thickness in their centers, according to the GRAIL-derived crustal thickness maps, which supports the interpretation that the mantle of the Moon was excavated during the formation of these basins.

The locations of the olivine-rich sites observed by Kaguya were compared with the locations of the mantle exposures modeled in this study to test whether such deposits are compatible with a mantle origin. Our numerical impact simulations of the basin formation process are made in the iSALE-2D hydrocode and they show that several basins on the nearside hemisphere, and at least one basin on the farside hemisphere, likely excavated material from the upper mantle of the Moon. Our results also support the interpretation that the high olivine abundances detected by Kaguya spacecraft could indeed be derived from the lunar mantle.
(Final answers to the reviewers are shown in green)

Reviewer 2:

Page 5 line 132: Why was the vertical impact speed set to 17 km s⁻¹? Also, describe how the impact speed affects (or does not affect) your results.

Reviewer’s comments: You didn’t respond to my previous comment “Also, describe how the impact speed affects (or does not affect) your results.”

L145-150: “and the vertical impact speed was set to 17 km s⁻¹. This choice for the impact speed represents an average vertical component of the velocity vector at moderately oblique angles of incidence (40°–70°) given that the mean asteroid impact speed during the basin-formation epoch (i.e., the Late Heavy Bombardment) was 20.9 km s⁻¹ (Bottke et al., 2012; Le Feuvre and Wieczorek, 2011).”

We added in L153-154: “For the proposed range of impact energies, the distribution of excavated mantle material exposed within the main rim is largely independent of the selected combination of projectile diameter and speed.”

Page 8 line 201:

Authors reply: ...we chose to fix the impact velocity to a reasonable and uncontroversial value, and then to vary the projectile diameter to fit the basin diameter.

...for the proposed range of impact energies, the amount of the mantle exposure within the main rim is present irrespective of the chosen projectile-speed combination

Reviewer’s comments: Then please state that you chose the impactor diameter and velocity to fit the observed basin diameter and that the amount of the mantle exposure does not depend on the chosen projectile-speed combination

We added in L242-244: “The model produced a basin with a diameter of crustal thinning equal to 480 km, which is in close agreement with the GRAIL-determined value of 476 km (Miljković et al., 2013).”

Page 9 line 218-219: Indicate the location of mantle material exposure in Fig. 3 bottom.

Authors reply: We have clarified the locations of mantle exposures in both Fig 3 and 4 (bottom) in the figure captions.

Reviewer’s comments: Are you saying that the location of the mantle exposure (dark grey) is not visible in Fig. 3 and Fig. 4 at the locations you described in the new caption, but that it is present in your simulation results?
Yes. We additionally clarified the caption of Figure 3: “The mantle exposures extend outward from the basin center to a radial distance of about 270 km and possibly beyond. Present day mare basalts (top panel) could have covered mantle exposures within the central portion of the basin (bottom panel), which could be the reason why Kaguya observations are limited to regions surrounding the mare deposits and beyond (top panel).”

If so, please mention it because otherwise readers will misunderstand that you are saying that the melt pool location corresponds to mantle material and that the solidified melt pool should be olivine-rich.

This is what we are saying. See L263-265: “All mantle material exposed in, and just exterior to, the central portion of the basin is predicted to have been in an initial state of partial to complete melting.” Also, in the Figure 3 caption: “The melt pool, which is composed almost entirely of mantle, is shown by dark yellow.”

Page 11 line 267:

Authors reply: The tracer particles in Figure 3 show that the uppermost 10-15 km of the mantle is exposed in the areas around the rim of the basin, which is where Kaguya observed olivine-rich signatures, and we acknowledge that this is difficult to see in the image.

Reviewer’s comments: If so, just say in the text that it is difficult to see in the image, but your simulation results indicate the uppermost 10 to 15 km of the mantle is exposed in the areas for clarification.

We added in L313-314: “(as shown by tracer particles in Figs. 3 and 4, although these tracers may be difficult to see clearly),”

Page 15 line 364-372:

Reviewer’s comments: The second pyroxene-rich mantle prior to a global overturn seems consistent with the recent compositional study reported by Ohtake et al. (2014) and their interpretation.

We add reference to Ohtake et al., 2014 in L422, and also reference their earlier work Ohtake et al., 2012 in L415-418: “On the basis of surface reflectance measurements, nearside–farside differences in the history of solidification of the lunar magma ocean, could have caused the farside lunar highlands to be more magnesian than the nearside highlands (Ohtake et al., 2012).”
Reviewer #1: I have reviewed the revised manuscript; the authors have done an excellent job of responding to and incorporating the suggestions that I made in the initial review. I think is now acceptable for publication in essentially its present form.

Minor notes/typos/clarity suggestions:

Line 187-191. "Of course ... million-year time-scales (Melosh et al., 2013)". I suggest rewording this somehow. My suggested version: "GRAIL measures basin gravity signatures that exist today. This is modified from initial impact structure modeled as a result of long-term cooling and viscous relaxation processes, which uplift the central basin floor by a few kilometers over million-year time-scales subsequent to the impact (Melosh et al., 2013)".

Implemented.

Line 329. "By to this", delete "to".

Corrected.

Line 335. An->A

Corrected.
Excavation of the lunar mantle by basin-forming events on the Moon

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Abstract

Global maps of crustal thickness on the Moon, derived from gravity measurements obtained by NASA's Gravity Recovery and Interior Laboratory (GRAIL) mission, have shown that the lunar crust is thinner than previously thought. Hyperspectral data obtained by the Kaguya mission have also documented areas rich in olivine that have been interpreted as material excavated from the mantle by some of the largest lunar impact events. Numerical simulations were performed with the iSALE-2D hydrocode to investigate the conditions under which mantle material may have been excavated during large impact events and where such material should be found. The results show that excavation of the mantle could have occurred during formation of the several largest impact basins on the nearside hemisphere as well as the Moscoviense basin on the farside hemisphere. Even though large areas in the central portions of these basins were later covered by mare basaltic lava flows, surficial lunar mantle deposits are predicted in areas external to these maria. Our results support the interpretation that the high olivine abundances detected by Kaguya spacecraft could indeed be derived from the lunar mantle.

Keywords: Moon; impact cratering; mantle

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1. Introduction

The Moon is a differentiated body composed of a geochemically distinct crust, mantle, and core. Most current theories for lunar formation involve a giant impact into the proto-Earth (Hartmann and Davis, 1975; Benz et al., 1986, 1987, 1989; Canup and Asphaug, 2001; Canup, 2004; Ćuk and Stewart, 2012; Canup, 2012) and rapid accretion of material from a hot circum-terrestrial disk, yielding a Moon with a global magma ocean that subsequently crystallized and differentiated (for a review, see Shearer et al., 2006). Analyses of the mare basalts (e.g., Longhi, 1992) and simulations of lunar magma ocean crystallization (e.g., Elkins-Tanton et al., 2011) indicate that the mantle is composed predominantly of the minerals olivine and pyroxene, in contrast to a crust that consists mostly of anorthite (e.g., Wieczorek et al., 2006).

Results of inversions of seismic travel-time data acquired by the Apollo seismic network are also consistent with the view that the lunar mantle is composed primarily of olivine and pyroxene (Khan et al., 2007; Kronrod and Kuskov, 2011), but there is no consensus on which of these two minerals is dominant. That no samples of the lunar mantle have been found to date in the lunar sample collection limits our understanding of the Moon’s early differentiation.

Initial analyses of the Apollo seismic data indicated that the crust of the Moon was about 60 km thick on the central nearside, and even thicker on the farside (Toksöz et al., 1974). Given such a thickness, only the largest impact basins would have been capable of excavating through the crust and into the underlying mantle, such as the Imbrium and Serenitatis basins on the nearside (Spudis, 1993). Geochemical analyses of impact melt breccias believed to be derived from the Imbrium impact have been modeled as a mixture of anorthositic crustal materials, KREEP-rich materials, and a highly forsteritic olivine component that may have been derived from the upper mantle (Korotev, 2000).

Crustal thickness models derived from gravity and topography datasets acquired by the Clementine and Lunar Prospector missions resolved regions of extremely thin crust beneath the
largest nearside impact basins (Zuber et al., 1994; Neumann et al., 1996). These models
demonstrated that the largest basins excavated several tens of kilometers into the crust and that
the crust is nearly absent beneath a few impact basins, notably Crisium (Wieczorek and
Phillips, 1998, 1999). Updated gravity models models from the Kaguya mission of the Japan
Aerospace Exploration Agency (JAXA) confirmed these results and further suggested that the
crust is also nearly absent beneath the Moscoviense basin on the lunar farside (Ishihara et al.,
2009).

Recent global maps of lunar crustal thickness derived from gravity measurements obtained
by NASA’s Gravity Recovery and Interior Laboratory (GRAIL) mission (Zuber et al., 2013)
show that the crust is substantially thinner on average than previously thought and that the
average density of the crust has been markedly lowered by impact fracturing (Wieczorek et al.,
2013). GRAIL-derived crustal thickness models are the first to satisfy revised interpretations of
the Apollo seismic data (Khan and Mosegaard, 2002; Lognonné et al., 2003; Chenet et al.
2006) and imply an average crustal thickness that is somewhere between 34 and 43 km
(Wieczorek et al., 2013). With crustal thicknesses that are less than previously thought by as
much as a factor of two, the likelihood that the mantle was excavated by large impacts is
increased. As with previous models, the GRAIL crustal thickness models predict that the crust
is nearly absent beneath both the nearside Crisium and farside Moscoviense basins. Other
basins, such as Humboldtianum, Apollo, and Poincaré, also have predicted interior crustal
thickness values less than 5 km.

Surface reflectance data collected by the spectral profiler on the Kaguya spacecraft have
detected regions on the lunar surface with anomalously high abundances of olivine (Yamamoto
et al., 2010, 2012). The spectra of these exposures are consistent with olivine-dominated
lithologies (i.e., dunite), though small quantities of admixed plagioclase cannot be excluded
(Yamamoto et al., 2010, 2012). These olivine-rich exposures have horizontal extents of several
kilometers and are, for the most part, located within relatively narrow regions along the
prominent innermost rings of the largest basins. There are no olivine-rich sites observed in the central regions of these basins, though most have been flooded by mare basalts, or exterior to the main topographic rims.

The reports from the Kaguya mission are not the first time that olivine-rich areas were detected on the surface of the Moon, but they do represent the first detections of nearly pure olivine from orbit. Prior to the Kaguya mission, some of the highest abundances of olivine were detected in the central peak of Copernicus (e.g., Pieters and Wilhelms, 1985; Isaacson et al., 2011), with analyses of Clementine spectral reflectance data suggesting a troctolitic composition of about 72% olivine and 22% plagioclase (Cahill et al., 2009). High abundances of olivine (up to nearly 50%) have also been detected in some mare basalts (e.g., Lucey, 2004).

The most prominent olivine-rich areas detected by the Kaguya spacecraft surround the Crisium basin on the nearside and the Moscoviense basin on the farside, with less prominent areas associated with the Imbrium, Serenitatis, Nectaris, Humorum, and Humboldtianum basins on the nearside and Shrödinger on the farside (Fig. 1). Observations by the Moon Mineralogy Mapper (M3) on the Chandrayaan-1 spacecraft have confirmed many instances of olivine-rich areas reported by Yamamoto et al. (2010) and added several additional olivine-rich locations within the Crisium (Powell et al., 2012), Moscoviense (Isaacson et al., 2011), and Nectaris (McGovern et al., 2013) basins. Because olivine may be an important mineral in the upper mantle of the Moon, particularly if the crystallization products of the magma ocean overturned after solidification (Hess and Permentier, 1995; Elkins-Tanton et al., 2011), the observed olivine-rich deposits may have been brought to the surface during large basin-forming events (Yamamoto et al., 2010, 2012).

Previous hydrocode simulations of the impact basin formation process have predicted that a substantial amount of mantle material was excavated during the largest impact events on the Moon (Ivanov et al., 2010; Potter et al., 2012a, 2012b, 2013). Moreover, Potter et al. (2013),
showed that the maximum excavation depth differed by less than 2% for two identical impacts into targets with thermal profiles that differed by ~500 K at a depth of 100 km. However, whereas the pre-impact temperature–depth profile has little effect on crater excavation, it is an important parameter that governs the final morphology of an impact basin, including the final diameter of the region of crustal thinning (Miljković et al., 2013). In part because of the complex nature of the basin formation process, these previous studies did not investigate the locations where exposed mantle should be found at the surface, the expected thickness of the deposits, or whether these materials would have been initially molten or solid.

In this study, we investigate the conditions under which mantle material can be excavated during large impacts, and we quantify the distribution of excavated material. The numerical simulations were fit to observational constraints from high-resolution Lunar Reconnaissance Orbiter (LRO) topographic data (Smith et al., 2010) and GRAIL-derived crustal thickness models (Wieczorek et al., 2013) in the vicinity of lunar basins. The locations of the olivine-rich sites observed by Kaguya were compared with the locations of the mantle exposures modeled in this study to test whether such deposits are compatible with a mantle origin.

2. Numerical modeling of large impacts

The formation of lunar impact basins has been modeled using iSALE-2D, a multi-material, multi-rheology shock physics hydrocode (Amsden et al., 1980; Collins et al., 2004; Wünnemann et al., 2006). A half-space target mesh in cylindrical coordinates was divided into crustal and mantle layers. The crust and mantle layers were represented with material models for basalt and dunite, respectively. These consisted of equation-of-state tables derived using ANEOS (ANalytical Equation Of State) for basalt (Pierazzo et al., 2005) and dunite (Benz et al., 1989), combined with a strength and failure model (Collins et al., 2004) for which the model parameters were adopted from Pierazzo et al. (2005) and Ivanov et al. (2010) for basalt and dunite, respectively. These material models have been used previously for modeling basin
formation on the Moon and Mars (Pierazzo et al., 2005; Ivanov et al., 2010; Miljković et al., 2013) and are similar (in terms of strength parameters and reference density) to those used in other lunar impact basin modeling studies (Potter et al., 2012b; Melosh et al., 2013). It was found in this study (by replacing basalt with granite) and in other studies (e.g., Melosh et al., 2013) that the choice of equation of state for the crustal material does not make a substantial difference to the outcome of a basin-forming event, for a given reference density.

The cell size in all simulations was 1.5 × 1.5 km; the thickness of the pre-impact crustal layer was fixed at 30, 45, or 60 km; the impactor diameter ranged from 15 to 120 km; and the vertical impact speed was set to 17 km s\(^{-1}\). This choice for the impact speed represents an average vertical component of the velocity vector at moderately oblique angles of incidence (40°–70°) given that the mean asteroid impact speed during the basin-formation epoch (i.e., the Late Heavy Bombardment) was 20.9 km s\(^{-1}\) (Bottke et al., 2012; Le Feuvre and Wieczorek, 2011). This range of impactor properties spans impact events expected to form craters of a size range from small peak-ring basins (~150-km diameter) to large multi-ring basins (300–1000-km diameter). For the proposed range of impact energies, the distribution of excavated mantle material exposed within the main rim is largely independent of the selected combination of projectile diameter and speed. The numerical setup and the material, thermal, and acoustic fluidization models used in the simulations were identical to those of Miljković et al. (2013).

There are major geological differences between the lunar nearside and farside hemispheres. The nearside is dominated by the compositionally distinctive Procellarum KREEP Terrane (PKT), which is highly enriched in heat-producing and other incompatible elements likely concentrated during the late stages of magma-ocean crystallization (Jolliff et al., 2000; Korotev, 2000). Combined with the evidence from the preponderance of mare basalt deposits on the nearside hemisphere and the viscous relaxation of topographic relief of nearside basins (Kamata et al., 2013), this hemispherical asymmetry in crustal and upper mantle heat
production indicates higher-than-average subsurface temperatures within and surrounding the PKT for a considerable interval of lunar history (Wieczorek and Phillips, 2000; Zhong et al., 2000; Hess and Permentier, 2001; Laneuville et al., 2013).

In our numerical simulations, the PKT region, the rest of the lunar nearside, and the farside hemisphere were represented by different temperature profiles adopted from the three-dimensional (3D) lunar thermal evolution models of Laneuville et al. (2013) at 4.0 Ga, a time near the formation times of most lunar basins. The crust and upper mantle of the PKT region had considerably higher temperatures at that time than the other regions as a result of the enrichment in heat-producing elements (i.e., KREEP) in the crust. Therefore, the temperature profile for the PKT region is here denoted as “hot.” The high heat production in the PKT affects the surrounding regions, so the rest of the nearside as well as the regions on the limbs between the nearside and farside hemispheres are modeled with an “intermediate” temperature profile. The farside is unaffected by the PKT region and is considerably cooler than the nearside hemisphere; its temperature profile is denoted as “cold” (Fig. 2).

The morphology of the final basin is highly dependent on the assumed temperature profile (Potter et al., 2012a; Miljković et al., 2013). In addition to the nominal temperature profiles, we also employed two variations for the hot, intermediate, and cold temperature profiles, which arose from two different initial temperature conditions for the mantle in the lunar thermal evolution models of Laneuville et al. (2013). For the nominal model, the initial temperature profile followed the solidus to a depth of 350 km and an adiabatic gradient below. For a cooler variant, the initial temperature was assumed to follow an adiabatic gradient for the entire mantle. We also used the temperature profiles at 3.5 Ga ago, an age younger than the last large impact basin, Orientale (Le Feuvre and Wieczorek, 2011), to test the sensitivity of our simulations to the time of impact (Fig. 2).
Numerical modeling results shown here represent the first 2 hours after impact. In contrast, observations of gravity and topography are for the modern Moon. The gravity field and topography of a lunar basin have been modified from those of its initial structure in response to long-term cooling and viscous relaxation processes, which uplift the central basin floor by several kilometers over million-year time scales following the impact (Melosh et al., 2013). As this relaxation involves nearly vertical displacements, it should not affect markedly the crustal thickness profile or the distribution of the excavated mantle, which are the primary observations that we use here to validate our simulations.

As this study required a large number of impact simulations, calculations not presently feasible with a full 3D shock physics code, we employed two-dimensional (2D, cylindrical geometry) impact models that enforced an impact incidence angle of 90° from the horizontal. Although vertical impacts are rare, 75% of all impacts occur at an impact angle larger than 30° from the horizontal. Several numerical modelling studies (e.g., Elbeshausen et al., 2009) have shown that the morphology of large impact craters is axially symmetric for impacts at angles larger than 30°, implying that the vertical impact approximation is adequate. Moreover, although crater size decreases with decreasing impact angle (e.g., Elbeshausen et al., 2009), the maximum depth of excavation remains a constant fraction of the transient crater diameter (for angles >30°; Artemieva et al., 2013). Thus, although oblique impact would result in an asymmetric distribution of ejecta, with more material distributed in the downrange direction, the provenance of the deepest ejected material is expected to be similar to that for vertical impact.

Given that temperature is the main controlling factor in our simulations, the projections of crustal thickness in Fig. 1 are centered over the PKT and the opposing hemisphere. Since the center of the PKT, at (20°N, 335°E), is close to the sub-Earth point, throughout this paper we will refer to these two hemispheres for simplicity as the nearside and farside. In using this
projection, the Orientale basin is located on the PKT hemisphere, and the Smythii basin is located on the opposite hemisphere.

3. Simulations of mantle excavation

Impact basins form via the growth of a deep, bowl-shaped transient crater followed by a complex collapse process to produce the final basin morphology. The collapse of the gravitationally unstable transient crater occurs by a combination of inward motion of the crater walls and uplift of the crater floor (Melosh, 1989; O’Keefe and Ahrens, 1993), both of which depend sensitively on the shear strength and temperature of the crust and upper mantle (Potter et al., 2012a; Miljković et al., 2013). During crater formation mantle materials can become exposed on the surface in one of two ways: (i) as part of the ballistic ejecta that is deposited outside the transient crater rim and (ii) as part of the outwardly collapsing central uplift that is thrust up over the inwardly collapsing transient crater rim (e.g., Potter et al., 2012b, 2013).

As shown in Fig. 1, the most prominent olivine-rich areas detected from spectral reflectance surround the Crisium impact basin on the lunar nearside and the Moscoviense basin on the farside. As the GRAIL-derived crustal thickness values in these two basins approach zero at their centers, we consider these two basins to be the most likely candidates to have excavated mantle material during their formation. We therefore simulated the formation of the Crisium and Moscoviense basins in detail and investigated the properties of any excavated mantle material. Numerical simulations were validated with GRAIL crustal thickness data, and the locations of mantle exposures were compared with the olivine-rich locations documented by Kaguya. Numerical simulations of other impact basins are discussed in Section 4.

Both the Crisium and Moscoviense basins were modeled with the iSAFE-2D hydrocode as vertical impacts of 60-km-diameter dunite projectiles at 17 km/s impact velocity, parameters that provided basin profiles well matched with GRAIL-derived crustal thickness profiles. The pre-impact crustal thickness in the case of Crisium was 30 km, and that in the case of
Moscoviense was 45 km. The Crisium basin is located outside the PKT region on the nearside hemisphere, and we therefore used the intermediate temperature profile. The Moscoviense basin is located in the farside highlands, so the cold temperature profile (v2 for 4.0 Ga, Fig. 2) was used for this basin.

The result of our iSALE-2D simulation of the Crisium basin is shown in Fig. 3. The model produced a basin with a diameter of crustal thinning equal to 480 km, which is in close agreement with the GRAIL-determined value of 476 km (Miljković et al., 2013). Note that the diameter of crustal thinning is defined as twice the radial distance from the basin center to the point at which the crustal thickness first reaches its pre-impact value. The diameter of the inner ring of Crisium has been estimated to be approximately 540 km and that of the topographic rim to be 740 km (Pike and Spudis, 1987). The inner “peak ring” is found to be located just exterior to the region of crustal thinning, whereas the main topographic rim crest is seen to be located near the hinge line where crustal materials were displaced downward.

[Fig. 3]

The extent of mantle uplift, the exposure of upper mantle material at the surface, and the extent of crustal overturn are illustrated in Fig. 3. The numerical simulations show that mantle material was exposed at the surface in the central portion of the basin, where the crust was (nearly) completely excavated, and just outward of the region of crustal thinning. All mantle material exposed in, and just exterior to, the central portion of the basin is predicted to have been in an initial state of partial to complete melting. The extent of the melt pool, formed within the first 2 hours after impact, is defined as material that is more that 50% partially molten (Fig. 3), under the assumption that melt fraction increases linearly with temperature between the solidus and liquidus (Ivanov et al., 2010; Potter et al., 2012b).

More than 2 hours after impact, i.e., at times after our simulations were stopped, several processes are expected to have further modified the basin structure. First, as described by Melosh et al. (2013), the impact melt pool would have cooled and contracted, and the sub-
isostatic state of the crust just outward of the melt pool would have caused the entire central
portion of the basin to have slowly rebounded upwards. This process would give rise to a
positive gravity anomaly in the center of the largest basins, consistent with observations.
Second, our simulations show that within the impact melt pool less than 1.5% of crustal
material was mixed in with the molten mantle material. If perfectly segregated during
crystallization, this admixture would form a crustal layer only 2 km thick, a small figure
compared with the uncertainties in the seismic constraints employed in generating the GRAIL
crustal thickness models (Wieczorek et al., 2013). However, such a new crustal layer could
have been sufficient to mask a spectral signature of pure olivine. Third, mare basaltic lavas,
derived from partial melting of the mantle up to hundreds of millions of years later, were
emplaced in this basin (Hiesinger et al., 2003; Fernandes et al., 2013). Mare basaltic material in
central Crisium is believed to be about 1–2 km thick (e.g., Wieczorek et al., 2006) and covered
almost the entire region of crustal thinning. The locations of olivine-rich exposures in Crisium,
as observed by Kaguya, are just exterior to the mare basalts and are in the vicinity of the basin
peak ring. As shown in Fig. 3, these locations match those where our simulations predict that
mantle material would have been exposed at the surface.

In addition to the mantle exposures found exterior to the mare basaltic lava flows, craters
Picard and Pierce, with diameters of 23 and 19 km, respectively, should have excavated
material from up to 2.5 km depth (Kalynn et al., 2013). Therefore, these two craters may have
evacuated through the thin mare basalt deposits and into the compositionally distinctive
underlying layer. X-ray fluorescence data from the Apollo 15 mission have shown these craters
to have atypical magnesium-rich compositions (Andre et al., 1978), perhaps consistent with a
contribution from excavated mantle material (Wieczorek and Phillips, 1998).

The numerical simulation results for the Moscovienne basin are shown in Fig. 4. With
identical impact conditions, but with a cooler thermal profile, the iSALE-2D model produced a
basin with a crustal thinning diameter of 320 km, which is nearly identical to the GRAIL-
derived value of 319 km. The diameter of the peak ring of Moscoviense is estimated to be 192 km and that of its main rim 421 km (Baker et al., 2011). The melt pool is composed of less than 1% crustal material, which could have given rise to a crustal layer no more than 1 km thick on top of the melt pool if this material were perfectly segregated. The locations of olivine-rich exposures in Moscoviense, as observed by Kaguya, are located just exterior to the mare basalts, approximately between the basin peak ring and the main topographic rim. As shown in Fig. 4, these locations are in good agreement with those where our simulations predict mantle material to be exposed at the surface. [Fig. 4]

In this study we modeled the Moscoviense basin as a single impact. However, the final morphology of the Moscoviense basin could have originated from two nearly overlapping large impacts separated in time; by such an interpretation, the first impact formed the outermost ring 640 km in diameter, and the second impact formed the 420-km-diameter main rim and 190-km-diameter inner ring that are offset from the outermost ring (Ishihara et al., 2011). Under this hypothesis of Ishihara et al. (2011), we essentially model only the second impact that occurred onto a target with the ambient crustal thickness observed by GRAIL. There is a possibility that this ambient crust was thinned by the first impact, but given the lack of evidence constraining the size of the first impact and the axial symmetry of the 2D model, we cannot speculate whether the first impact was capable of excavating any mantle material.

Our numerical simulations have shown the extent and locations of mantle exposures at the surface of the Moon for the case of Crisium and Moscoviense basins. The surface exposures of mantle are in agreement with the locations of olivine-rich areas observed by Kaguya within these basins. Furthermore, our simulations also suggest that the mantle exposures mainly originate from the uppermost 10–15 km of the mantle (as shown by tracer particles in Figs. 3 and 4, although these tracers may be difficult to see clearly), but some contribution from the deeper mantle cannot be excluded. Together, these results support the interpretation of
Yamamoto et al. (2010) that the observed olivine-rich deposits are indeed from an olivine-dominated lunar upper mantle.

4. Mantle excavation threshold

The previous section illustrated that two impacts of the same energy (same impactor diameter and velocity) on different parts of the Moon with different pre-impact thermal states both exposed mantle materials yet resulted in very different crustal thinning diameters. To expand our study to other lunar impact basins, a suite of iSALE-2D simulations was performed to determine the threshold crater size for exposing mantle at the surface as a function of impact and target conditions. The target conditions considered were the temperature profile and pre-impact crustal thickness. The projectile size, which is directly related to the diameter of the region of crustal thinning in the final basin (for a given pre-impact temperature profile), was also varied.

The results of all simulations are presented in Tables S1 and S2. These tables also include diameters of the transient craters for all simulations. The transient crater size was found to be nearly independent of the temperature profile and the pre-impact crustal thickness. Following Potter et al. (2012b), we defined the time of transient crater formation as the moment the opening cavity in the target was at its largest volume, which occurred shortly after the crater floor began to rebound. By this definition, the transient crater diameter scales with impactor parameters according to standard scaling laws (Ivanov et al., 2010; Potter et al., 2012b). Previous numerical models of impact basin formation showed that the maximum depth of excavation (defined as the pre-impact depth of the deepest material exposed at the surface in the final crater) is about one-tenth of the transient crater diameter (Potter et al., 2012b, 2013), independent of pre-impact target temperature, planetary curvature, impactor size, or impact velocity (within the range 10–20 km/s). A ratio of excavation depth to diameter of ~0.1 is also consistent with experimental, analytical, geological, and geophysical investigations of impact
craters of all sizes (Croft, 1980; Melosh, 1989; Spudis, 1993; Wieczorek and Phillips, 1999; Hikida and Wieczorek, 2007).

The conditions for mantle exposure on the lunar nearside and farside hemispheres are shown in Figs. 5 and 6, respectively, as functions of crustal thinning diameter and ambient crustal thickness. The hot temperature profile representative of the nearside PKT was used for generating Fig. 5, whereas the cold temperature profile representative of the farside was used for Fig. 6. Each symbol represents one simulation, and for a given crustal thickness, each crustal thinning diameter corresponds to a different impactor size (with larger impactors creating larger craters). Blue symbols denote basins that did not show any mantle exposures at the surface at the end of the simulation (both within and exterior to the region of crustal thinning), and red symbols denote basins in which the mantle was exposed within the region of the crustal thinning, and/or where mantle material was exposed on top of other portions of the crust (e.g., as seen in Figs. 3 and 4).

The red-shaded regions in the upper portion of these figures mark the range of parameter space for which mantle should be found at the surface of the nearside and farside hemispheres, respectively. The threshold crustal thinning diameter for exposing mantle material is smaller for the hot target (Fig. 5) than for the cold target (Fig. 6), which implies that, for the same crustal thickness, there is a range of basin size (crustal thinning diameter) that should expose mantle on the farside but not on the nearside. We emphasize that this difference in threshold crustal thinning diameter is a consequence of the strong influence of temperature on crustal thinning diameter. Consistent with previous work (e.g., Potter et al., 2013), target temperature has only a minor influence on the maximum depth of excavation. For the same impactor properties, transient crater size and maximum excavation depth are nearly independent of the target conditions. However, the final crustal structure of the basin is highly sensitive to the assumed temperature profile. Hotter upper mantle temperatures lower the strength of the target, which enhances uplift and produces a broader zone of uplifted mantle and thinned crust.
Hence, whereas impacts with the same impactor properties, regardless of the crustal thickness, would form transient craters similar in size on both hemispheres, the resultant crustal thinning diameter could be as much as twice as large on the nearside than on the farside (Miljković et al., 2013).

[Fig. 5]  
[Fig. 6]  

The thresholds for excavating mantle material as derived from the iSALE-2D simulations can be compared with actual lunar basins. The GRAIL-derived diameters of crustal thinning and the pre-impact crustal thicknesses are plotted in Figs. 7 and 8 for the nearside and farside basins, respectively. Here, the South Pole-Aitken basin was excluded, as it is the oldest and largest impact structure on the Moon and likely formed under conditions that are not represented by our simulations. As indicated in Fig. 7, mantle material should have been excavated during the formation of the largest basins on the nearside, and the majority of these basins (Imbrium, Serenitatis, Crisium, Nectaris, Humorum) are indeed associated with olivine-rich exposures. The basin Humboldtiyanum is also associated with olivine-rich exposures, and this basin plots close to, but just below, the expected transition. Those basins that are associated with exposures of pure anorthosite (Yamamoto et al., 2012) are predicted to have excavated only crustal material. All basins that are associated with olivine exposures are Nectarian in age or younger.

[Fig. 7]  
[Fig. 8]  

As indicated in Fig. 8, on the farside hemisphere, which has a thicker average crustal thickness than the nearside, only the Moscovyiense and Smythii basins could have excavated mantle material. (Symthii is located on the nearside hemisphere, but it is on the hemisphere opposite to the PKT). Though the Moscovyiense basin shows evidence of olivine exposures, there are no olivine or pure anorthosite detections associated with Smythii. As this is one of the
oldest basins (pre-Nectarian in age) to show a crustal thinning anomaly in its interior, it is
plausible that such deposits were covered by ejecta from younger basins. Similar to the
situation on the nearside hemisphere, basins associated with pure anorthosite are predicted to
have excavated only crustal materials. The only basin that is anomalous is the Schrödinger
basin, which shows evidence for both pure anorthosite and olivine deposits in its interior, even
though our simulations predict that only crustal material should have been excavated during the
formation of this basin.

That Schrödinger is an exception may be related to the fact that this basin formed on the
dge of the South Pole-Aitken basin. Although Schrödinger should not have excavated mantle
material under normal conditions, it is possible that it excavated ejecta from the older South
Pole-Aitken basin that contained mantle material. Alternatively, it is possible that the South
Pole-Aitken impact melt sheet differentiated, forming a thick olivine layer (Vaughan and Head,
2014) and that Schrödinger later excavated portions of this solidified melt sheet.

Numerical impact models of the giant South Pole-Aitken basin show that this impact event
should have exposed a substantial amount of mantle material at the lunar surface (Potter et al.,
2012b; Melosh et al., 2014). Those results are consistent with simple extrapolation of the
trends presented here (Figures 5 and 6). However, as of yet, olivine-rich deposits have not been
conclusively associated with the South Pole-Aitken basin, and the composition of the interior
of this basin is notably rich in pyroxene (Lucey, 2004; Ohtake et al., 2014). As this is the oldest
recognizable impact structure on the Moon, it is possible that this basin could have formed
prior to the lunar mantle overturn event when the upper mantle was still composed of
pyroxene-rich materials. Given the great age of the basin, it is even possible that it formed
before the magma ocean completely crystallized. On the basis of surface reflectance
measurements, nearside–farside differences in the history of solidification of the lunar magma
ocean could have caused the farside lunar highlands to be more magnesian than the nearside
highlands (Ohtake et al., 2012). Under any of these scenarios, the composition of the ejecta
from this basin would have contained a relatively high normative abundance of pyroxene.

Alternatively, it is possible that the ejecta of this basin was in fact rich in olivine, but that these materials were mixed, diluted, and buried by subsequent impact events.

5. Conclusions

The identification of lunar mantle material, either in the sample collection or from orbital observations, would be of great importance for understanding the lunar formation process. Our simulations show that several basins on the nearside hemisphere, and at least one basin on the farside hemisphere, likely excavated material from the upper mantle of the Moon. Many of these basins are associated with deposits of nearly pure olivine in their interiors, and our results are consistent with the interpretation of Yamamoto et al. (2010) that these deposits are mantle-derived material.

If the upper mantle of the Moon is indeed olivine rich, this inference has important implications for the processes involved during the Moon’s initial differentiation. As the lunar magma ocean crystallized, magnesium-rich olivine would have been one of the first minerals to have crystallized and such crystals would have sunk to the bottom of the mantle. Later-crystallizing material would have been richer in both iron and pyroxene, and the latest-crystallizing fraction would have had high abundances of the dense mineral ilmenite (e.g., Shearer et al., 2006; Elkins-Tanton et al., 2011). Because the solidified products of magma ocean crystallization would have been gravitationally unstable, it has been hypothesized that the entire mantle would have overturned (Hess and Permentier, 1995). The identification of nearly pure olivine deposits derived from the upper lunar mantle is consistent with this hypothesis. A further prediction is that these deposits should be considerably more magnesian than their ferroan anorthosite counterparts.

Impact simulations and remote sensing data show that mantle material should be found within the interiors of several large basins on the Moon. Mantle material should constitute a
large fraction of these deposits and should be easily accessible at the lunar surface. As samples of the mantle of a terrestrial planetary body, they should be considered a high priority for any future sample return mission to the Moon.

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**Figure captions**

**Fig. 1.** Crustal thickness of the Moon derived from GRAIL gravity data (Wieczorek et al., 2013), updated with the gravity model of Konopliv et al. (2013), and locations of olivine-rich exposures (stars) as documented by Kaguya (Yamamoto et al., 2010). The largest expanses of olivine are observed around the Crisium and Moscoviense basins, which have crustal thickness values close to zero. The maps are Lambert azimuthal equal-area projections centered over (left) the Procellarum KREEP Terrane (20°N, 335°E) and (right) the opposite hemisphere; grid lines are spaced every 30°.

**Fig. 2.** Temperature profiles used for modeling lunar impact basin formation. “Hot,” “intermediate,” and “cold” profiles are representative of the PKT, off-PKT nearside, and farside, respectively. The “hot” profile includes enhanced heat production at the base of the nearside crust, and the “intermediate” profile includes enhanced heat production that is distributed uniformly within the nearside crust. The designations v1 and v2 denote two different initial conditions for the mantle temperature profile; for v1, the initial temperature profile is set to the solidus for the upper 350 km and follows an adiabatic gradient below, and for v2 the initial temperature profile is set to an adiabat for the entire mantle. The profiles are adopted from the 3D lunar thermal model of Laneuville et al. (2013). Similar temperature profiles were used by Miljković et al. (2013).
Fig. 3. (top) iSALE-2D simulation of Crisium basin formation 2 h after impact (black) compared with modern topography and the location of the crust-mantle interface inferred from GRAIL and LOLA measurements (gray). The extent of the mare basalts is denoted in dark orange (as observed in present day), and the magenta stars indicate locations of olivine-rich areas documented by Kaguya. Blue arrows denote the location of the peak ring and main rim, and the approximate locations of the Picard and Pierce impact craters are shown by red circles. (bottom) Location of crustal (light gray) and mantle (dark gray) material from the iSALE-2D simulation (2 h after impact), with tracer particles demonstrating mantle uplift (green) and crustal overturn (blue). The melt pool, which is composed almost entirely of mantle, is shown by dark yellow. The mantle exposures extend outward from the basin center to a radial distance of about 270 km and possibly beyond. Present day mare basalts (top panel) could have covered mantle exposures within the central portion of the basin (bottom panel), which could be the reason why Kaguya observations are limited to regions surrounding the mare deposits and beyond (top panel).

Fig. 4. Same as Fig. 3, but for the Moscoviense basin. The iSALE-2D results (bottom) show that the mantle is exposed in the central uplift region (to about 130 km radial distance from the basin center) and between 150 km and 250 km radial distance from the basin center.

Fig. 5. Exposure of mantle material on the lunar surface by impact basin formation as a function of the basin’s crustal thinning diameter and pre-impact crustal thickness. In this plot, “hot” temperature profiles were used for the crust and mantle that are representative of the PKT. Each symbol represents one simulation with different impact conditions (projectile size and velocity). Red and blue symbols are for simulations that exposed, and did not expose, mantle material at the surface, respectively. The red-shaded region denotes those conditions, on the basis of an extrapolation from the simulation results, for which mantle material is expected to be exposed at the surface.
Fig. 6. Same as Fig. 5, but for simulations of impacts into a cold target, representative of the lunar farside.

Fig. 7. Exposures of olivine and pure anorthosite in lunar nearside impact basins as a function of the basin’s crustal thinning diameter and pre-impact crustal thickness. Red shading denotes the range of impact conditions that should have led to the excavation of mantle material on the nearside hemisphere of the Moon, on the basis of iSALE simulations. Red symbols denote basins with olivine-rich deposits (Yamamoto et al., 2010); in order of decreasing size, they are Imbrium, Serenitatis, Crisium, Nectaris, Humorum, and Humboldtianum. Blue symbols denote basins with pure anorthosite deposits (Yamamoto et al., 2012). Unfilled symbols denote basins where neither was detected.

Fig. 8. Same as Fig. 7, but for impact basins that formed in the lunar farside highlands. The red circle denotes the Moscoviense basin, and the circle half in red and half in blue denotes the Schrödinger basin. The largest basin shown is Smythii.
Highlights

- Excavation of the mantle occurred during formation of the largest lunar basins.
- Mantle material should be found within the interiors of several basins on the Moon.
- Predicted mantle exposures agree with areas of high olivine abundances detected by Kaguya.
- Samples of the mantle material should be easily accessible at the lunar surface.
Figure
Click here to download Figure: Figure6-revised.pdf

- Cold temperature profile
- ▲ No mantle exposures
- ▼ Mantle exposed on surface

Crustal thinning diameter (km) vs. Pre-impact crustal thickness (km)
Figure
Click here to download Figure: Figure8.pdf

Excavation of mantle

Lunar farside basins
- Olivine
- Pure anorthosite
- No detections

Crustal thinning diameter (km)

Pre-impact crustal thickness (km)