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Title: 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.

Cover letter

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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36 1. Introduction

37 The Moon is a differentiated body composed of a geochemically distinct crust, mantle, and
38 core. Most current theories for lunar formation involve a giant impact into the proto-Earth
39 (Hartmann and Davis, 1975; Benz et al., 1986, 1987, 1989; Canup and Asphaug, 2001; Canup,
40 2004; Ćuk and Stewart, 2012; Canup, 2012) and rapid accretion of material from a hot circum-
41 terrestrial disk, yielding a Moon with a global magma ocean that subsequently crystallized and
42 differentiated (for a review, see Shearer et al., 2006). Analyses of the mare basalts (e.g.,
43 Longhi, 1992) and simulations of lunar magma ocean crystallization (e.g., Elkins-Tanton et al.,
44 2011) indicate that the mantle is composed predominantly of the minerals olivine and
45 pyroxene, in contrast to a crust that consists mostly of anorthite (e.g., Wieczorek et al., 2006).
46 Results of inversions of seismic travel-time data acquired by the Apollo seismic network are
47 also consistent with the view that the lunar mantle is composed primarily of olivine and
48 pyroxene (Khan et al., 2007; Kronrod and Kuskov, 2011), but there is no consensus on which
49 of these two minerals is dominant. That no samples of the lunar mantle have been found to date
50 in the lunar sample collection limits our understanding of the Moon's early differentiation.

51 Initial analyses of the Apollo seismic data indicated that the crust of the Moon was about
52 60 km thick on the central nearside, and even thicker on the farside (Toksöz et al., 1974).
53 Given such a thickness, only the largest impact basins would have been capable of excavating
54 through the crust and into the underlying mantle, such as the Imbrium and Serenitatis basins on
55 the nearside (Spudis, 1993). Geochemical analyses of impact melt breccias believed to be
56 derived from the Imbrium impact have been modeled as a mixture of anorthositic crustal
57 materials, KREEP-rich materials, and a highly forsteritic olivine component that may have
58 been derived from the upper mantle (Korotev, 2000).

59 Crustal thickness models derived from gravity and topography datasets acquired by the
60 Clementine and Lunar Prospector missions resolved regions of extremely thin crust beneath the

61 largest nearside impact basins (Zuber et al., 1994; Neumann et al., 1996). These models
62 demonstrated that the largest basins excavated several tens of kilometers into the crust and that
63 the crust is nearly absent beneath a few impact basins, notably Crisium (Wieczorek and
64 Phillips, 1998, 1999). Updated gravity models from the Kaguya mission of the Japan
65 Aerospace Exploration Agency (JAXA) confirmed these results and further suggested that the
66 crust is also nearly absent beneath the Moscoviense basin on the lunar farside (Ishihara et al.,
67 2009).

68 Recent global maps of lunar crustal thickness derived from gravity measurements obtained
69 by NASA's Gravity Recovery and Interior Laboratory (GRAIL) mission (Zuber et al., 2013)
70 show that the crust is substantially thinner on average than previously thought and that the
71 average density of the crust has been markedly lowered by impact fracturing (Wieczorek et al.,
72 2013). GRAIL-derived crustal thickness models are the first to satisfy revised interpretations of
73 the Apollo seismic data (Khan and Mosegaard, 2002; Lognonné et al., 2003; Chenet et al.
74 2006) and imply an average crustal thickness that is somewhere between 34 and 43 km
75 (Wieczorek et al., 2013). With crustal thicknesses that are less than previously thought by as
76 much as a factor of two, the likelihood that the mantle was excavated by large impacts is
77 increased. As with previous models, the GRAIL crustal thickness models predict that the crust
78 is nearly absent beneath both the nearside Crisium and farside Moscoviense basins. Other
79 basins, such as Humboldtianum, Apollo, and Poincaré, also have predicted interior crustal
80 thickness values less than 5 km.

81 Surface reflectance data collected by the spectral profiler on the Kaguya spacecraft have
82 detected regions on the lunar surface with anomalously high abundances of olivine (Yamamoto
83 et al., 2010, 2012). The spectra of these exposures are consistent with olivine-dominated
84 lithologies (i.e., dunite), though small quantities of admixed plagioclase cannot be excluded
85 (Yamamoto et al., 2010, 2012). These olivine-rich exposures have horizontal extents of several
86 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).

[Fig. 1]

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),

113 showed that the maximum excavation depth differed by less than 2% for two identical impacts
114 into targets with thermal profiles that differed by ~ 500 K at a depth of 100 km. However,
115 whereas the pre-impact temperature–depth profile has little effect on crater excavation, it is an
116 important parameter that governs the final morphology of an impact basin, including the final
117 diameter of the region of crustal thinning (Miljković et al., 2013). In part because of the
118 complex nature of the basin formation process, these previous studies did not investigate the
119 locations where exposed mantle should be found at the surface, the expected thickness of the
120 deposits, or whether these materials would have been initially molten or solid.

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

128 **2. Numerical modeling of large impacts**

129 The formation of lunar impact basins has been modeled using iSALE-2D, a multi-material,
130 multi-rheology shock physics hydrocode (Amsden et al., 1980; Collins et al., 2004;
131 Wünnemann et al., 2006). A half-space target mesh in cylindrical coordinates was divided into
132 crustal and mantle layers. The crust and mantle layers were represented with material models
133 for basalt and dunite, respectively. These consisted of equation-of-state tables derived using
134 ANEOS (ANalytical Equation Of State) for basalt (Pierazzo et al., 2005) and dunite (Benz et
135 al., 1989), combined with a strength and failure model (Collins et al., 2004) for which the
136 model parameters were adopted from Pierazzo et al. (2005) and Ivanov et al. (2010) for basalt
137 and dunite, respectively. These material models have been used previously for modeling basin

138 formation on the Moon and Mars (Pierazzo et al., 2005; Ivanov et al., 2010; Miljković et al.,
139 2013) and are similar (in terms of strength parameters and reference density) to those used in
140 other lunar impact basin modeling studies (Potter et al., 2012b; Melosh et al., 2013). It was
141 found in this study (by replacing basalt with granite) and in other studies (e.g., Melosh et al.,
142 2013) that the choice of equation of state for the crustal material does not make a substantial
143 difference to the outcome of a basin-forming event, for a given reference density.

144 The cell size in all simulations was 1.5×1.5 km; the thickness of the pre-impact crustal
145 layer was fixed at 30, 45, or 60 km; the impactor diameter ranged from 15 to 120 km; and the
146 vertical impact speed was set to 17 km s^{-1} . This choice for the impact speed represents an
147 average vertical component of the velocity vector at moderately oblique angles of incidence
148 (40° – 70°) given that the mean asteroid impact speed during the basin-formation epoch (i.e., the
149 Late Heavy Bombardment) was 20.9 km s^{-1} (Bottke et al., 2012; Le Feuvre and Wieczorek,
150 2011). This range of impactor properties spans impact events expected to form craters of a size
151 range from small peak-ring basins (~150-km diameter) to large multi-ring basins (300–1000-
152 km diameter). For the proposed range of impact energies, the distribution of excavated mantle
153 material exposed within the main rim is largely independent of the selected combination of
154 projectile diameter and speed. The numerical setup and the material, thermal, and acoustic
155 fluidization models used in the simulations were identical to those of Miljković et al. (2013).

156 There are major geological differences between the lunar nearside and farside hemispheres.
157 The nearside is dominated by the compositionally distinctive Procellarum KREEP Terrane
158 (PKT), which is highly enriched in heat-producing and other incompatible elements likely
159 concentrated during the late stages of magma-ocean crystallization (Jolliff et al., 2000;
160 Korotev, 2000). Combined with the evidence from the preponderance of mare basalt deposits
161 on the nearside hemisphere and the viscous relaxation of topographic relief of nearside basins
162 (Kamata et al., 2013), this hemispherical asymmetry in crustal and upper mantle heat

163 production indicates higher-than-average subsurface temperatures within and surrounding the
164 PKT for a considerable interval of lunar history (Wieczorek and Phillips, 2000; Zhong et al.,
165 2000; Hess and Permentier, 2001; Laneuville et al., 2013).

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

177 **[Fig. 2]**

178 The morphology of the final basin is highly dependent on the assumed temperature profile
179 (Potter et al., 2012a; Miljković et al., 2013). In addition to the nominal temperature profiles, we
180 also employed two variations for the hot, intermediate, and cold temperature profiles, which
181 arose from two different initial temperature conditions for the mantle in the lunar thermal
182 evolution models of Laneuville et al. (2013). For the nominal model, the initial temperature
183 profile followed the solidus to a depth of 350 km and an adiabatic gradient below. For a cooler
184 variant, the initial temperature was assumed to follow an adiabatic gradient for the entire
185 mantle. We also used the temperature profiles at 3.5 Ga ago, an age younger than the last large
186 impact basin, Orientale (Le Feuvre and Wieczorek, 2011), to test the sensitivity of our
187 simulations to the time of impact (Fig. 2).

188 Numerical modeling results shown here represent the first 2 hours after impact. In contrast,
189 observations of gravity and topography are for the modern Moon. The gravity field and
190 topography of a lunar basin have been modified from those of its initial structure in response to
191 long-term cooling and viscous relaxation processes, which uplift the central basin floor by
192 several kilometers over million-year time scales following the impact (Melosh et al., 2013). As
193 this relaxation involves nearly vertical displacements, it should not affect markedly the crustal
194 thickness profile or the distribution of the excavated mantle, which are the primary
195 observations that we use here to validate our simulations.

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

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

213 projection, the Orientale basin is located on the PKT hemisphere, and the Smythii basin is
214 located on the opposite hemisphere.

215 **3. Simulations of mantle excavation**

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

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

234 Both the Crisium and Moscoviense basins were modeled with the iSALE-2D hydrocode as
235 vertical impacts of 60-km-diameter dunite projectiles at 17 km/s impact velocity, parameters
236 that provided basin profiles well matched with GRAIL-derived crustal thickness profiles. The
237 pre-impact crustal thickness in the case of Crisium was 30 km, and that in the case of

238 Moscoviense was 45 km. The Crisium basin is located outside the PKT region on the nearside
239 hemisphere, and we therefore used the intermediate temperature profile. The Moscoviense
240 basin is located in the farside highlands, so the cold temperature profile (v2 for 4.0 Ga, Fig. 2)
241 was used for this basin.

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

251 **[Fig. 3]**

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

261 More than 2 hours after impact, i.e., at times after our simulations were stopped, several
262 processes are expected to have further modified the basin structure. First, as described by
263 Melosh et al. (2013), the impact melt pool would have cooled and contracted, and the sub-

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

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

287 The numerical simulation results for the Moscoviense basin are shown in Fig. 4. With
288 identical impact conditions, but with a cooler thermal profile, the iSALE-2D model produced a
289 basin with a crustal thinning diameter of 320 km, which is nearly identical to the GRAIL-

290 derived value of 319 km. The diameter of the peak ring of Moscoviense is estimated to be 192
291 km and that of its main rim 421 km (Baker et al., 2011). The melt pool is composed of less
292 than 1% crustal material, which could have given rise to a crustal layer no more than 1 km
293 thick on top of the melt pool if this material were perfectly segregated. The locations of
294 olivine-rich exposures in Moscoviense, as observed by Kaguya, are located just exterior to the
295 mare basalts, approximately between the basin peak ring and the main topographic rim. As
296 shown in Fig. 4, these locations are in good agreement with those where our simulations
297 predict mantle material to be exposed at the surface.

298 **[Fig. 4]**

299 In this study we modeled the Moscoviense basin as a single impact. However, the final
300 morphology of the Moscoviense basin could have originated from two nearly overlapping large
301 impacts separated in time; by such an interpretation, the first impact formed the outermost ring
302 640 km in diameter, and the second impact formed the 420-km-diameter main rim and 190-
303 km-diameter inner ring that are offset from the outermost ring (Ishihara et al., 2011). Under
304 this hypothesis of Ishihara et al. (2011), we essentially model only the second impact that
305 occurred onto a target with the ambient crustal thickness observed by GRAIL. There is a
306 possibility that this ambient crust was thinned by the first impact, but given the lack of
307 evidence constraining the size of the first impact and the axial symmetry of the 2D model, we
308 cannot speculate whether the first impact was capable of excavating any mantle material.

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

316 Yamamoto et al. (2010) that the observed olivine-rich deposits are indeed from an olivine-
317 dominated lunar upper mantle.

318 **4. Mantle excavation threshold**

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

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

341 craters of all sizes (Croft, 1980; Melosh, 1989; Spudis, 1993; Wieczorek and Phillips, 1999;
342 Hikida and Wieczorek, 2007).

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

354 The red-shaded regions in the upper portion of these figures mark the range of parameter
355 space for which mantle should be found at the surface of the nearside and farside hemispheres,
356 respectively. The threshold crustal thinning diameter for exposing mantle material is smaller
357 for the hot target (Fig. 5) than for the cold target (Fig. 6), which implies that, for the same
358 crustal thickness, there is a range of basin size (crustal thinning diameter) that should expose
359 mantle on the farside but not on the nearside. We emphasize that this difference in threshold
360 crustal thinning diameter is a consequence of the strong influence of temperature on crustal
361 thinning diameter. Consistent with previous work (e.g., Potter et al., 2013), target temperature
362 has only a minor influence on the maximum depth of excavation. For the same impactor
363 properties, transient crater size and maximum excavation depth are nearly independent of the
364 target conditions. However, the final crustal structure of the basin is highly sensitive to the
365 assumed temperature profile. Hotter upper mantle temperatures lower the strength of the target,
366 which enhances uplift and produces a broader zone of uplifted mantle and thinned crust.

367 Hence, whereas impacts with the same impactor properties, regardless of the crustal thickness,
368 would form transient craters similar in size on both hemispheres, the resultant crustal thinning
369 diameter could be as much as twice as large on the nearside than on the farside (Miljković et
370 al., 2013).

371 **[Fig. 5]**

372 **[Fig. 6]**

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

386 **[Fig. 7]**

387 **[Fig. 8]**

388 As indicated in Fig. 8, on the farside hemisphere, which has a thicker average crustal
389 thickness than the nearside, only the Moscoviense and Smythii basins could have excavated
390 mantle material. (Smythii is located on the nearside hemisphere, but it is on the hemisphere
391 opposite to the PKT). Though the Moscoviense basin shows evidence of olivine exposures,
392 there are no olivine or pure anorthosite detections associated with Smythii. As this is one of the

393 oldest basins (pre-Nectarian in age) to show a crustal thinning anomaly in its interior, it is
394 plausible that such deposits were covered by ejecta from younger basins. Similar to the
395 situation on the nearside hemisphere, basins associated with pure anorthosite are predicted to
396 have excavated only crustal materials. The only basin that is anomalous is the Schrödinger
397 basin, which shows evidence for both pure anorthosite and olivine deposits in its interior, even
398 though our simulations predict that only crustal material should have been excavated during the
399 formation of this basin.

400 That Schrödinger is an exception may be related to the fact that this basin formed on the
401 edge of the South Pole-Aitken basin. Although Schrödinger should not have excavated mantle
402 material under normal conditions, it is possible that it excavated ejecta from the older South
403 Pole-Aitken basin that contained mantle material. Alternatively, it is possible that the South
404 Pole-Aitken impact melt sheet differentiated, forming a thick olivine layer (Vaughan and Head,
405 2014) and that Schrödinger later excavated portions of this solidified melt sheet.

406 Numerical impact models of the giant South Pole-Aitken basin show that this impact event
407 should have exposed a substantial amount of mantle material at the lunar surface (Potter et al.,
408 2012b; Melosh et al., 2014). Those results are consistent with simple extrapolation of the
409 trends presented here (Figures 5 and 6). However, as of yet, olivine-rich deposits have not been
410 conclusively associated with the South Pole-Aitken basin, and the composition of the interior
411 of this basin is notably rich in pyroxene (Lucey, 2004; Ohtake et al., 2014). As this is the oldest
412 recognizable impact structure on the Moon, it is possible that this basin could have formed
413 prior to the lunar mantle overturn event when the upper mantle was still composed of
414 pyroxene-rich materials. Given the great age of the basin, it is even possible that it formed
415 before the magma ocean completely crystallized. On the basis of surface reflectance
416 measurements, nearside–farside differences in the history of solidification of the lunar magma
417 ocean could have caused the farside lunar highlands to be more magnesian than the nearside
418 highlands (Ohtake et al., 2012). Under any of these scenarios, the composition of the ejecta

419 from this basin would have contained a relatively high normative abundance of pyroxene.
420 Alternatively, it is possible that the ejecta of this basin was in fact rich in olivine, but that these
421 materials were mixed, diluted, and buried by subsequent impact events.

422 **5. Conclusions**

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

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

442 Impact simulations and remote sensing data show that mantle material should be found
443 within the interiors of several large basins on the Moon. Mantle material should constitute a

444 large fraction of these deposits and should be easily accessible at the lunar surface. As samples
445 of the mantle of a terrestrial planetary body, they should be considered a high priority for any
446 future sample return mission to the Moon.

447

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456

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 626

627 **Figure captions**

628 **Fig. 1.** Crustal thickness of the Moon derived from GRAIL gravity data (Wieczorek et al.,
629 2013), updated with the gravity model of Konopliv et al. (2013), and locations of olivine-rich
630 exposures (stars) as documented by Kaguya (Yamamoto et al., 2010). The largest expanses of
631 olivine are observed around the Crisium and Moscoviense basins, which have crustal thickness
632 values close to zero. The maps are Lambert azimuthal equal-area projections centered over
633 (left) the Procellarum KREEP Terrane (20°N, 335°E) and (right) the opposite hemisphere; grid
634 lines are spaced every 30°.

635 **Fig. 2.** Temperature profiles used for modeling lunar impact basin formation. “Hot,”
636 “intermediate,” and “cold” profiles are representative of the PKT, off-PKT nearside, and
637 farside, respectively. The “hot” profile includes enhanced heat production at the base of the
638 nearside crust, and the “intermediate” profile includes enhanced heat production that is
639 distributed uniformly within the nearside crust. The designations v1 and v2 denote two
640 different initial conditions for the mantle temperature profile; for v1, the initial temperature
641 profile is set to the solidus for the upper 350 km and follows an adiabatic gradient below, and
642 for v2 the initial temperature profile is set to an adiabat for the entire mantle. The profiles are
643 adopted from the 3D lunar thermal model of Laneuville et al. (2013). Similar temperature
644 profiles were used by Miljković et al. (2013).

645 **Fig. 3.** (top) iSALE-2D simulation of Crisium basin formation 2 h after impact (black)
646 compared with modern topography and the location of the crust-mantle interface inferred from
647 GRAIL and LOLA measurements (gray). The extent of the mare basalts is denoted in dark
648 orange (as observed in present day), and the magenta stars indicate locations of olivine-rich
649 areas documented by Kaguya. Blue arrows denote the location of the peak ring and main rim,
650 and the approximate locations of the Picard and Pierce impact craters are shown by red circles.
651 (bottom) Location of crustal (light gray) and mantle (dark gray) material from the iSALE-2D
652 simulation (2 h after impact), with tracer particles demonstrating mantle uplift (green) and
653 crustal overturn (blue). The melt pool, which is composed almost entirely of mantle, is shown
654 by dark yellow. The mantle exposures extend outward from the basin center to a radial distance
655 of about 270 km and possibly beyond. Present day mare basalts (top panel) could have covered
656 mantle exposures within the central portion of the basin (bottom panel), which could be the
657 reason why Kaguya observations are limited to regions surrounding the mare deposits and
658 beyond (top panel).

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

662 **Fig. 5.** Exposure of mantle material on the lunar surface by impact basin formation as a
663 function of the basin's crustal thinning diameter and pre-impact crustal thickness. In this plot,
664 "hot" temperature profiles were used for the crust and mantle that are representative of the
665 PKT. Each symbol represents one simulation with different impact conditions (projectile size
666 and velocity). Red and blue symbols are for simulations that exposed, and did not expose,
667 mantle material at the surface, respectively. The red-shaded region denotes those conditions, on
668 the basis of an extrapolation from the simulation results, for which mantle material is expected
669 to be exposed at the surface.

670 **Fig. 6.** Same as Fig. 5, but for simulations of impacts into a cold target, representative of the
671 lunar farside.

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

680 **Fig. 8.** Same as Fig. 7, but for impact basins that formed in the lunar farside highlands. The
681 red circle denotes the Moscoviense basin, and the circle half in red and half in blue denotes the
682 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.

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