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Investigating the Influences of Crustal Thickness and Temperature on the Uplift of Mantle Materials Beneath Large Impact Craters on the Moon

Min Ding^{1,2,3}, Jason M. Soderblom³, Carver J. Bierson⁴, Maria T. Zuber³

¹State Key Laboratory of Lunar and Planetary Sciences, Macau University of Science and Technology, Macau, China.

²CNSA Macau Center for Space Exploration and Science, Macau, China.

³Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139, USA.

⁴Department of Earth and Planetary Sciences, University of California Santa Cruz, Santa Cruz, CA 95064.

Corresponding author: Min Ding (miding@must.edu.mo)

Key Points:

- The minimum crater diameter for which impacts result in mantle uplift is positively correlated with local crustal thickness.
- Gravity signatures of mare craters suggest that the mare deposits account for ~ 25–35% of the thickness of post-impact deposits.
- Statistical indistinguishability between the SPA and FHT craters provides an upper limit on the effect of thermal difference.

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Abstract

In this work, we examine variations in the mantle uplift associated with large lunar impact craters and basins between major terranes. This study is based on Bouguer gravity anomalies of 100–650-km diameter impact craters using Gravity Recovery and Interior Laboratory (GRAIL) observations and the Lunar Orbiter Laser Altimeter (LOLA) crater database. The Bouguer gravity anomalies of 324 large impact craters analyzed herein are primarily controlled by the uplifted crust-mantle (Moho) interface in the central region of these impact craters, while post-impact mare deposits that represent ~ 25–35% of the post-impact deposit column also contribute to the gravity anomalies of mare craters. The central uplift of the Moho interface is primarily controlled by impact energy and increases to ~ 30 km for a 650-km diameter crater. Analyses of non-mare craters in the Feldspathic Highlands Terrane (FHT) with varied crustal thickness (T_c) reveal that the onset crater diameter (D_{min}) with an uplifted Moho interface is dependent on the local T_c : $D_{min} \sim 150 + 1.3T_c$ (in a unit of km). This equation also provides a quantification of the depth-dependent attenuation of impact-induced structural uplift, using the Moho uplift as a proxy for the structural uplift. The Moho uplift of large craters in the South Pole-Aitken (SPA) Terrane is not statistically different from the FHT craters, consistent with limited overall thermal difference between these two terranes and compensational thermal effects at crater collapse and viscoelastic relaxation stages.

Plain Language Summary

Gravitational signatures of impact basins reveal notable mantle uplifts under the basin floor. The mantle uplift is significant for lunar basins with a crater diameter larger than ~ 200 km and is one of the main characteristics for peak-ring and multi-ring basins. The magnitude of the mantle uplift increases with impact energy, which also correlates with crater diameter. It has been suggested that the target properties, including the crustal thickness and thermal state, affect the crater mantle uplift as well. In this study, we find that increasing crustal thickness reduces basin mantle uplift. But there is no statistical difference between the mantle uplifts of impact basins formed in the highlands and those in the South Pole-Aitken basin. We attribute this to limited thermal differences between the two terranes in the pre-Nectarian period, which is consistent with the hypothesized early lunar thermal evolution and impact chronology. The compensational thermal effects at the transient crater collapse and viscoelastic relaxation stages also help explain this statistical indistinguishability.

1 Introduction

It has long been recognized that impact basins are associated with notable uplift of mantle materials (e.g., Pilkington & Grieve, 1992; Cintala & Grieve, 1998). The impact-induced mantle uplift has been investigated by both laboratory (e.g., Schmidt & Housen, 1987) and numerical experiments (e.g., Potter et al., 2013; Milbury et al., 2015). On the Moon, the onset of mantle (or more precisely the crust-mantle boundary, Moho) uplift coincides with the morphologic transition from complex craters to peak-ring basins. This coincidence has strong implications for the formation mechanism of the peak ring. Dynamic collapse models (Collins et al., 2002; Baker et al., 2016; Morgan et al., 2016) show that the peak ring is formed by convergence of the outward collapse of initially uplifted deep crustal rocks and the inward collapsing rim materials of the transient cavity.

While impact conditions (e.g., impact energy and impactor density) are the primary factor in determining the magnitude of Moho uplift, target properties (e.g., crustal thickness and temperature) have measurable influences on Moho uplift from which valuable information can be gained. Moho uplift occurs more easily in regions of thinner crust, resulting in an inverse relationship between the crustal thickness and the onset of Moho uplift (Milbury et al., 2015). Hotter thermal states weaken target materials by reducing the yield strength and enhancing impact melt, which reduces viscosity, resulting in greater Moho uplift (Potter et al., 2013; Miljković et al., 2016).

Post-impact viscoelastic relaxation of impact structures occurs on timescales of ~ 10 Ma. Isostatic adjustment of impact basins and thermal contraction of impact melts during this relaxation results in the uplift of the entire basin and the formation of gravitational mascon (Melosh et al., 2013; Montesi, 2013; Freed et al., 2014). The temperature of the lithosphere influences this process, with colder targets more rapidly freezing the crustal structure of impact basins in the lithosphere. In contrast, warmer targets have enhanced viscous creep of crustal materials, which tends to flatten the uplifted Moho interface (Kamata et al., 2015). This mechanism acts in an opposite sense as the effect of temperature during the transient crater collapse. The net influence of the temperature requires further investigation.

High resolution and precision gravity data from the Gravity Recovery and Interior Laboratory (GRAIL) mission (Zuber et al., 2013; Goossens et al., 2020) provide an unprecedented opportunity to investigate the internal structure of impact craters and to infer crater formation and modification processes. Unlike surface topography, which degrades over time, leading to the loss of evidence of ancient impact basins, the underlying Moho interface is better preserved and can be inferred from gravity data. GRAIL gravity data, for example, have been used to identify previously unknown impact basins (Neumann et al., 2015). GRAIL gravity also provides constraints on the onset crater diameter (D_{min}) when the central Moho uplift starts to exist. A global D_{min} of ~ 200 km has been found (Soderblom et al., 2015; Baker et al., 2017). However, there has yet to be a systematic study to investigate the effect of target properties on this onset of Moho uplift.

In this study, we use the Bouguer gravity anomalies of 100–650-km diameter craters to infer the central uplifts of the underlying Moho interface. We analyze the effects of crater diameter and crustal thickness on the Moho uplift using craters in the Feldspathic Highlands Terrane (FHT). We constrain the thickness and gravitational contribution of post-impact mare deposits by comparing mare and non-mare craters in the FHT. By comparing the FHT craters with the craters in the South Pole-Aitken Terrane (SPA), we infer the effects of thermal difference. Because we consider a wide range of impact morphologies, from complex craters to proto-basins, and from peak-ring basins to multi-ring basins (e.g., Baker et al., 2012, 2016, 2017), for simplicity we refer to them all as impact craters.

2 Crater Parameters and Data Analysis Method

2.1 Central Bouguer gravity anomaly

We analyze 324 impact craters with rim diameters (D_c) of 100–650 km identified in the Lunar Orbiter Laser Altimeter (LOLA) crater database (Head et al., 2010; Kadish et al., 2011; Figure 1). We use the GRAIL free-air gravity anomaly model JGGRAIL_1200C12A (Konopliv et al., 2014), and derive a Bouguer gravity model by subtracting the gravitational contribution of

topography. Our Bouguer correction assumes the laterally varying crustal density model of Wieczorek et al. (2013), with the unconstrained density of the mare region interpolated from existing data. The effect of using a constant crustal density is tested in Model (b) (Table 1 and Section 3.4). Spherical harmonic degrees > 600 (corresponding to a block-size resolution of < 9 km) are excluded due to their low signal-to-noise ratio. In Models (c), we test the application of a high-pass filter $l_{min} = \frac{2\pi R_0}{4D_c}$ (corresponding to a block-size resolution of $2D_c$) to each crater in order to remove gravity signals that are significantly larger than, and therefore irrelevant to the crater; this is the same approach employed by Bierson et al. (2016). Here R_0 is the reference radius of the Moon and is set to 1,738 km.

The local Bouguer gravity model for each crater is then referenced to an elevation corresponding to 60 km above the local Moho interface (measured as an average in a reference annulus from the outer rim flank, as quantified by Pike, 1977, to one crater radius beyond the crater rim) (Figure 2). This 60 km is the largest crustal thickness from the reference global model. Using these varied local references ensures that, for a given central Moho uplift (the focus of this study), variations in the local crustal thickness do not influence the gravity anomaly. It also guarantees the gravity anomaly to be measured above local topography. In this manner, the local Bouguer gravity anomaly for most craters is less than that in the global Bouguer gravity model with a constant reference radius of 1,738 km (Model d), particularly for those craters within regions of thin crust. The effect of using local topography as the reference (Model e) is similar (Figure 2). We use the most recent topography model from the Lunar Orbiter Laser Altimeter (LOLA) (Smith et al., 2016) and the reference crustal thickness model of Wieczorek et al. (2013) with a mean crustal thickness of 35 km. The effect of using a different crustal thickness model is tested in Model (f) (Table 1 and Section 3.4).

For each crater, the Bouguer gravity signature is characterized by a single measurement, the central Bouguer anomaly (CBA). Following Soderblom et al. (2015), CBA is defined as the difference between the area-weighted average Bouguer anomaly of the central region with a radial distance less than $0.2R$ (R is the crater radius) and that of the annular region from 0.5 to $1R$ (Figure 2). To assist the data analysis, we estimate the uncertainty of the crater CBA values as the standard deviation of the Bouguer gravity data points (with a block size of 9 km) within the central circular region. The corresponding p -value measures the probability for a two-sample t -test statistic to be more extreme than the observation under the null hypothesis that the mean Bouguer anomaly within the central circular region is less than or equal to that in the reference annulus. A small p -value casts doubt on the validity of the null hypothesis and therefore implies the CBA is indeed larger than zero. For a given significance level α , the craters can be divided into two groups: a positive CBA group with p -value $< \alpha$ and a non-positive group with p -value $\geq \alpha$. The α value is tested in a range of 0.001–0.1 with a higher α classifying more craters into the positive group. These parameters, as well as the other crater parameters, are all included in the crater parameter database (Dataset S1; Ding, 2020).

2.2 Central Moho uplift and mare deposits

The crater CBA is primarily influenced by mantle uplift and post-impact mare deposits (Figure 3). Other internal structures, from the impact-induced melt and porosity change, to post-impact breccia infills, all have effects that are more or less uniform from the crater center to rim and thus will not contribute significantly to the crater CBA, which identifies differences between the central region within $0.2R$ and the surrounding crater floor region. The negligibility of the

impact-induced porosity change is justified by the low correlation of -0.08 between the crustal porosity and crater CBA (Figure S1) and likely reflects the fact that these impacts are sufficiently large to reset the local porosity (Huang et al., 2020). The impact-induced melt has been observed ubiquitously within basin interiors (e.g., Cintala & Grieve, 1998) with a spatial scales of $\sim$ one crater diameter, and thus will not greatly influence the crater CBA.

For impact craters without mare deposits, we derive the magnitude of Moho uplift (d_M) after inverting for the relief of the crust-mantle (Moho) interface. The inversion for the Moho relief is conducted using the opensource software SHTools (Wieczorek & Phillips, 1998; Wieczorek & Meschede, 2018; Wieczorek et al., 2018) and the variable crustal density model (i.e., variable grain density with a crustal porosity of 12%) from Wieczorek et al. (2013). For the reference model, we assume a mean crustal thickness of 35 km. This crustal thickness is the same with the first crustal thickness model of Wieczorek et al. (2013) except that we update the gravity model with higher precision and resolution, which permits a higher $\lambda_{1/2}$ value of 120 (in comparison with 80 in Wieczorek et al., 2013) for the high-frequency filter. This high-frequency filter, which is required for the inversion algorithm to converge, is characterized by $\lambda_{1/2}$, the spherical harmonic degree at which the high-frequency filter reaches a value of 0.5. A higher $\lambda_{1/2}$ value permits more spherical harmonic degrees to be used to invert for crustal thickness variations at shorter wavelengths., which improves our sensitivity to the smaller craters considered in our study. We find, however, that our results are insensitive to the $\lambda_{1/2}$ value (Model g in Table 2 and Section 3.4).

The Moho uplift d_M is then measured as the difference between the area-weighted Moho relief in the central region within a radial distance $< 0.2R$ and that of the reference annular region from 0.5 to $1R$, spatially similar to the definition of the crater CBA (Figure 2). The inversion for the Moho relief also provides crustal thickness models, similar with Wieczorek et al. (2013).

For impact craters with mare deposits, quantification of this layer requires a spatial distribution map of mare basalts (Nelson et al., 2014). For a typical mare crater, Poincaré basin (Figure 4), we estimate the mare coverage within the central region (inside $0.2R$) to be 85% and mare coverage within the crater rim to be 27% (denoted as P_{mare} hereafter). The thickness of the mare basalts is uncertain, but its upper bound (referring to as maximum possible thickness, H_{max} , hereafter) can be determined by the difference between the observed crater depth and the expected fresh-crater depth (using scaling relationship from Pike, 1977; Kalynn et al., 2013; see details in Ding et al., 2018). Although the fresh crater depth/diameter scaling relationship of Kalynn et al. (2013) is derived from complex craters, this relationship is consistent with results of Williams & Zuber (1998) and Dobb & Kiefer (2015) for larger impact basins and thus is used for larger impact basins here. The lower bound of the mare basalt thickness is zero.

Assuming a mare density of $3,150 \text{ kg/m}^3$ and varied local crustal density ($2,690 \text{ kg/m}^3$ for Poincaré basin in Dataset S1, based on the crustal density model of Wieczorek et al., 2013), we calculate the gravitational attraction due to mare deposits in the spherical domain using SHTools. The corresponding maximum possible gravity anomaly in the central region of Poincaré basin (with a radial distance $< 0.2R$) is 48 mGal, and the gravity anomaly in the reference annulus region (from 0.5 to $1R$) is 7 mGal (Figure 4c). We then consider two possible scenarios: Model I is when both the central region and the reference annulus region are associated with the maximum possible mare thickness, while Model II corresponds to a scenario with the maximum possible mare thickness in the central region but zero mare thickness in the reference annulus

region. For the Poincaré basin, Model *I* provides a mare-induced CBA of 41 mGal, while Model *II* yields a larger mare-induced CBA of 48 mGal. Model *II* gives the maximum possible mare-induced CBA, although the CBA estimates of the two models are not significantly different. The modeled gravity attraction (with a reference radius of 1,738 km) is then upward continued to the same reference radius as the crater CBA calculation (i.e., 60 km above the local Moho interface, corresponding to 1,780 km for Poincaré basin), yielding mare-induced CBA values of 16 mGal for Model *I* and 22 mGal for Model *II*. This derivation has been conducted for all craters in our study, providing maximum possible CBA estimates due to mare deposits. By comparing the CBA values of mare craters with those of non-mare craters, we provide constraints on the thickness of mare deposits (Section 3.2).

Similar to mare deposits, cryptomare is expected to contribute to the gravity observations. However, as cryptomare deposits are commonly located in the vicinity of mare deposits, we only unambiguously identify two non-mare craters (Balmer and Milne) to be associated with cryptomare deposits, based on the cryptomare distribution of Whitten & Head (2015). Considering post-impact cryptomare deposits thus does not influence our statistics.

2.3 Crustal thickness and temperature

The candidate control parameters for the central Moho uplift d_M include the impact energy and target properties, most notably crustal thickness and target temperature (Figure 3). While impact energy (affected by a combination of impact velocity, size/density and angle) cannot be directly measured, it is known to correlate with D_c . At the same time, d_M is observed to correlate with D_c (e.g., Soderblom et al., 2015; Baker et al., 2017), suggesting that impact energy is a primary control of d_M .

The influences of target properties on d_M are more nuanced. To examine the effects of crustal thickness, we calculate local crustal thickness values T_c (as the average crustal thickness in a surrounding annulus from the outer rim flank (Pike, 1977) to one crater diameter from the rim), which range from 15 to 60 km for the craters in the reference model. Results are described in Section 3.1. Model (f) assumes a larger mean crustal thickness of 45 km and this sensitivity test is discussed in Section 3.4.

The thermal effects at the transient crater collapse (Potter et al., 2013) and post-impact viscoelastic relaxation (Kamata et al., 2015) stages likely compensate for each other. The net outcome of these two stages can be revealed after correcting for the thermal effect on the crater diameter (Miljković et al., 2013, 2016) (Section 3.3). To consider different thermal states, we compare the FHT and SPA craters. The SPA boundary is from Jolliff et al. (2000) and Garrick-Bethell & Zuber (2009). Choosing either the former thorium content-based boundary or the latter elliptical boundary for the SPA basin does not influence our results, and here we combine these two boundaries to associate more craters with SPA (Figure 1a). While the nearside Procellarum KREEP Terrane (PKT) almost certainly has yet another thermal state, the paucity of PKT craters makes it impossible to conduct a statistical analysis for this region.

Crater ages from Losiak et al. (2009) are also included in our crater parameter database. Although crater age is expected to correlate with the target temperature (at the time of crater formation) and thus the Moho uplift, the lack of precise crater dating (except for large impact basins with $D_c > 300$ km by Fassett et al., 2012 and Orgel et al., 2018) prohibits quantification of

crater age effects here. The age of our investigated craters is dominated by the pre-Nectarian period.

3 Results and Discussions

3.1 FHT: Effect of crustal thickness and implications

In order to examine the effects of crustal thickness on the onset and magnitude of Moho uplift, we consider 191 non-mare craters in the FHT. We fit these data to a two-slope model that assumes the CBA values are equal to zero for a crater diameter (D_c) less than the onset diameter D_{min} , but linearly increase when $D_c > D_{min}$ (Figure 5a): $CBA = a(D_c - D_{min})$. Although a higher-order polynomial fit is numerically more precise (Baker et al., 2017; Ding et al., 2020), this two-slope model provides one single proxy, D_{min} , to quantify the transition from zero to positive CBA. We estimate $D_{min} = 197 \pm 30$ km and $a = 0.58 \pm 0.15$ mGal/km (Table 1). The best-fit value and uncertainty are given by the mean and standard deviation of 1,000 resampled datasets using a bootstrap re-sampling method. This best-fit D_{min} value is smaller than 218 ± 41 km when fitting both the mare and non-mare craters (second row of Table 1), whereas the latter closely matches 218 ± 17 km provided by Soderblom et al. (2015). Our uncertainty is larger than that of Soderblom et al. (2015) because a smaller number of craters is considered. Because of the linear relationship assumed for $D_c > D_{min}$, the best-fit D_{min} is sensitive to the CBA values of impact craters much greater than 200 km.

A more accurate way to quantify the onset of $CBA > 0$ is by using a limited portion of craters with CBA values close to zero. For statistical robustness, we use the p -values of the crater CBAs to divide the craters into two groups: positive CBA group (p -value less than a significance level α of 0.01; red dots in Figure 5c) and non-positive CBA group (other craters; gray dots). We apply a linear discriminant analysis to find a linear decision boundary between the two groups, using T_c and D_c as control parameters. The coefficients of the decision boundary $D_{min} = cT_c + d$ are estimated to be $c = 1.3 \pm 1.0$ and $d = 150 \pm 39$ km. The D_{min} values estimated by the linear discriminant analysis and the two-slope model agree within error. In addition to uncertainty estimation, bootstrap re-sampling shows that c and d are highly correlated with a correlation coefficient of -0.97. The positive best-fit value for c is consistent with physical consideration that it is harder for impacts to induce Moho uplift in a thick crust than thin crust, though this is only a 1-sigma detection and should be considered with appropriate caution. The decision boundary is also sensitive to the arbitrarily chosen significant level α : A less conservative α of 0.05 yields a smaller c estimate of 0.7 ± 1.2 (Figure 5d). Therefore, this decision boundary provides an upper limit for the effects of crustal thickness on the onset of central Moho uplift.

For the craters with $CBA > 0$, we analyze the magnitude of the uplift d_M . The usage of $D_c - cT_c - d$ as the predictor parameter ensures that the central Moho uplift d_M starts from zero, and that the effects of T_c and D_c are consistently included. We find a regression coefficient, e , of 0.075 ± 0.007 (Figure 5b). It is worth noting that a direct two-variable linear regression for D_c and T_c yields a positive regression coefficient of 0.24 ± 0.16 for T_c (with a regression coefficient of 0.04 ± 0.01 for D_c and an intercept of -9.55 ± 5.87), which means that our assumed negative correlation between T_c and d_M is not directly supported by the data. However, a relatively high p -value of 0.15 for T_c suggests that the effect of T_c is insignificant. The lack of negative correlation between T_c and d_M (as well as CBA, Figure S1) is likely due to complex formation processes of the central Moho uplift, which may initially be uplifted together with the over-

heightened central peak materials and then collapse downward (Baker et al., 2016; Kring et al., 2016; Morgan et al., 2016). The final Moho uplift is also influenced by inward migration of crustal materials (Trowbridge et al., 2020) and crystallization of impact melt. These mechanisms and large data scattering due to individual impact characteristics, as well as limited number of large impact basins, together preventing a simple linear relationship between T_c and d_M .

The outliers in Figures 5a-b are craters with gravity signatures that are not well explained by the overall FHT trend. Among them, Szilard North crater (with a crater ID number of 259) overlaps with a smaller crater, Szilard (132), and thus the CBA is not reliable. The low-CBA Mutus-Vlacq crater (313) is probably not an impact crater but rather topographic depressions surrounded by thick impact ejecta (Byrne, 2016). The other three outliers, Janssen (274), Orientale Southwest (289) and Fitzgerald-Jackson (321), may be associated with anomalous impact characteristics, although the effect of local variations in target properties and impact age cannot be excluded.

Using T_c as a proxy for the varied crustal layer depth (z) beneath a crater, the Moho uplift d_M represents the depth-dependent structural uplift due to this impact event. For a given impact diameter D_c (e.g., 300 km in Figure 6a), we can predict structural uplift using $d_M = e(D_c - cz - d)$. The onset depth of structural uplift is determined by the decision boundary $D_c = cz + d$. Near the surface, highly elevated deep materials may be exposed and provide insights into the lunar composition and stratification at depth (e.g., Cintala & Grieve, 1998; Tompkins & Pieters, 1999; Kring, 2009). In order to directly compare with numerical modeling results of Potter et al. (2013), we normalize the depth and structure uplift by the transient crater radius (half of the transient crater diameter D_t) and the maximum structural uplift d_{max} , respectively, where $D_t = 1.54D_c^{0.84}$ and $d_{max} = 0.13D_c^{1.07}$ is the maximum amount of structural uplift at a depth of 15–30% of the transient crater radius from Potter et al. (2013). Figure 6b shows non-mare FHT craters as gray dots, which are well described by a two-slope model (blue lines and shaded region). The estimated depth for the onset of structural uplift here is consistent with our previous estimation using linear discriminant analysis (black horizontal line), but significantly less than provided by Potter et al. (2013). The numerical models of Potter et al. (2013) also overestimate the structural uplift at least by a factor of 5 (Figure 6b), likely due to underestimated yield strength or overestimated acoustic fluidization in their hydrocode models. For a 320-km diameter peak-ring impact basin, Schrödinger basin, the peak-ring materials have found to be originated from a depth of 20–30 km (Kring et al., 2016) by combining spectral and photogeological analysis and impact hydrocode simulation. This is consistent with the gravity-based Moho uplift values derived in this work. While we find consistency between our results and the original depth of peak-ring materials, direct extrapolation of our results to near-surface structures or deep materials require further justification (by observations or numerical modeling) as it is merely based on gravity-derived Moho relief.

3.2 Effect of mare deposits

Comparison between mare and non-mare craters in the FHT reveals the gravitational effect of mare deposits. On average, mare craters (triangles in Figure 7a) are associated with larger CBA than the non-mare crater model (blue curve and shaded region). In order to quantify the contributions of mare deposits to crater CBA, the non-mare crater CBA model is subtracted from the observed CBA for mare craters. The residual crater CBA, $CBA^{Residual}$, is expected to be due to mare deposits.

The $CBA^{Residual}$ of mare craters are then compared with results of gravity forward modeling. Assuming the maximum possible mare thickness H_{max} , Model *II* (thick mare in the central region only, yellow bars in Figure 7b) predicts slightly larger mare-induced CBA than Model *I* (thick mare in both the central and annulus region, red bars). For mid-sized craters with D_c less than 250 km, both models predict larger CBA than $CBA^{Residual}$ (gray dots). The mare deposits thus cannot occupy the entire column (i.e., from the observed crater depth to its initial depth from fresh crater scaling relationship) of post-impact deposits. The remainder of the deposit column is likely breccia infill and ejecta generated by long-term bombardment and landslides following the emplacement of the crater (i.e., crater degradation, Craddock and Howard, 2000; Fassett and Thomson, 2014; and Du et al., 2019).

Assuming a real mare thickness h_{mare} (less than H_{max}), its contribution to the crater CBA is $\sim 2\pi\Delta\rho h_{mare}$ according to horizontal thin sheet approximation, where $\Delta\rho$ is the density contrast between the mare and surrounding crustal materials. Therefore, the thickness ratio of mare deposits over the entire deposit column, h_{mare}/H_{max} , can be estimated by the ratio between the corresponding $CBA^{Residual}$ and the modeled maximum potential mare-induced CBA (Figure 7c). The average ratio is $\sim 34\%$ for Model *I* and $\sim 26\%$ for Model *II*. Large scatter in this ratio from zero to 100% reveals local variations in the mare eruption and deposition processes. Considering that the average H_{max} of our analyzed mid-sized mare craters is $\sim 2.1 \pm 0.9$ km, the corresponding average thickness of mare deposits is $\sim 0.63 \pm 0.28$ km. This estimation is based on a mare density of $3,150 \text{ kg/m}^3$, while a higher mare density of $3,250 \text{ kg/m}^3$ (e.g., Evans et al., 2016) yields slightly lower ratio of 25–30% and mare thickness of 0.58 ± 0.25 km. Our estimated h_{mare}/H_{max} is consistent with the previous estimate of $\sim 50\%$ (e.g., Du et al., 2019) to first order.

Large impact basins with mare deposits are associated with extremely large crater CBA that cannot be explained by post-impact mare deposits of only a few kilometers (Figure 7b). It is more likely that the mare basins preferentially sample and represent the basins with enhanced Moho uplift and crustal thickness reduction, because post-impact magma tends to erupt in regions with thin crust (e.g., Taguchi et al., 2017).

3.3 SPA: Effect of temperature and implications

We apply the same statistical analysis in Section 3.1 to the 51 SPA craters. Because there is a limited number of SPA craters, however, we need to include both mare and non-mare craters for statistical robustness. To ensure the comparability of SPA and FHT craters, results are shown for all the FHT craters in Figure 8. There is no statistical difference between the SPA and FHT craters, although the mean CBA for $D_c < D_{min}$ is -4.2 ± 1.6 mGal for SPA craters (Figure 8a). This slightly negative CBA is likely due to lower target porosity in the SPA (Ding et al., 2018) and impact-induced pore space under the crater floor (Milbury et al., 2015). The similar size-dependent behavior for SPA and FHT crater CBAs indicates that either the SPA thermal state is not significantly distinct from FHT, or the thermal effects are limited.

To further interpret this indistinguishability, we consider the thermal history of the SPA region. Comparing with FHT, this region's thermal state is influenced by two factors: impact heating within ~ 100 Ma after the SPA impact (Rolf et al., 2017) and long-term lack of radiogenic heating due to impact excavation of crustal materials (Figure 9a). The decrease in radiogenic heating is estimated by the time-dependent radiogenic heating of uranium (U), thorium (Th) and potassium (K), assuming a Th concentration of 1 ppm, a Th/U ratio of 3.7 and a K/Th ratio of 460 for the lunar crust (Laneuville et al., 2018). Heat generation rates and half-

lives of the radiogenic elements are from Turcotte & Schubert (2014). The radiogenic heat rate is then multiplied by the average thickness (T_{exc}) of the excavated crustal materials (i.e., ejected materials from the excavation zone of the transient crater) in SPA to yield the surface heat flux. We consider T_{exc} ranging from 15 to 35 km; the lower limit is from gravity analyses (Wieczorek et al., 2013; James et al., 2019), while the upper limit is derived from hydrocode simulations (Potter et al., 2012; Melosh et al., 2017). Recent hydrocode simulation of Trowbridge et al. (2020) explains the SPA basin interior as resulting from the inward migration of crustal materials during transient crater collapse, yielding T_{exc} that is similar to the gravity-based estimate.

Combining the effects of impact heating and lack of radiogenic heating, the surface heat flux in SPA was greater than that in the FHT for the first ~90 Ma after the SPA formation, but less than FHT afterwards (Figure 9a). The reference FHT thermal evolution path is from Laneuville et al. (2013). It is noted that if a mafic lower layer with substantial radiogenic content exists under the anorthosite crust (e.g., Ryder & Wood, 1977; Taylor & Wieczorek, 2014), a possibility that is still under debate (Laneuville et al., 2018), the lunar surface heat flux would be greatly increased. But as this mafic layer should exist in both FHT and SPA, our comparison between the two terranes still holds. This simplified calculation ignores the secondary effect of crustal radiogenic heating on the decay of SPA impact heating, which requires further thermal modeling.

To relate these thermal states to target properties for individual impact craters requires precise dating of impact craters, which is not yet available. We instead use the standard cumulative crater density function (i.e., monotonic declining function by Neukum et al. 2001; solid curve in Figure 9b) to normalize the surface heat flux, yielding average heat fluxes sampled by the SPA and FHT craters in the entire pre-Nectarian period (Figure 9c). Here we consider the pre-Nectarian period to be from the formation of SPA to 3.9 Ga. While isotopic dating and crater counting put the age of SPA at 4.25–4.3 Ga (Orgel et al., 2018; Garrick-Bethell et al., 2020), debates regarding the source of the dated samples and uncertainty in the early cratering chronology cast doubt on this age. To be conservative, herein we test a range of 4.1 to 4.45 Ga for the SPA formation.

We next consider the temperature-induced D_c variability. Impact hydrocode modeling by Miljković et al. (2016) suggests that while D_c is sensitive to the thermal state, the transient crater diameter D_t depends only on the impact energy. The scaling relationship $D_t = fD_c^g$ is thus useful for estimating D_c in varied target thermal state. Miljković et al. (2016) find $f_1 = 2.92$ and $g_1 = 0.77$ for the nearside with a surface temperature gradient of 20 K/km (corresponding to a surface heat flux of 60 mW/m² assuming a crustal heat conductivity of 3 W/m/K), and $f_2 = 2.48$ and $g_2 = 0.84$ for the FHT with a surface temperature gradient of 10 K/km (corresponding to 30 mW/m²). We estimate the change of D_c with surface heat flux by further assuming that the coefficients f and g are linearly related to the surface heat flux. These linear relationships are assumed only for computational purpose, and further numerical simulations are required to validate these relationships. Figure 9d shows the expected crater diameter D_c^* (this asterisk means *counterpart* crater diameter) for an impact into SPA with the same impact energy that would form a FHT crater with $D_c = 300$ km, as a function of varied SPA formation age. A larger T_{exc} and an earlier SPA formation age predicts D_c^* to be closer to 300 km. But the key result is that the difference between D_c^* and 300 km is less than 30 km, which is within the uncertainty range of the CBA– D_c relationship (e.g., Figure 5a, Table 1). Therefore, the overall temperature-induced D_c difference is limited between FHT and SPA craters.

Although the monotonic declining crater chronology in Figure 9b is naturally expected as the continuous disposal of planetesimals after the formation of planets (Morbidei et al., 2018), there are severe debates on the crater chronology in the pre-Nectarian period due to lack of direct constraints. We thus test another scenario, the sawtooth-like weak-cataclysm model by Morbidelli et al. (2012). Applying the similar calculation as for the monotonic declining model yields even less average thermal difference between SPA and FHT, and D_c^* closer to 300 km. It confirms that the current thermal models and crater chronology, although highly uncertain in the pre-Nectarian period, predict limited D_c difference between the FHT and SPA craters.

After confirming the limited thermal sensitivity of D_c , the statistical indistinction in the CBA– D_c relationship between FHT and SPA craters implies limited thermal effects on the Moho uplift process. It is possible that the thermal difference between FHT and SPA, normalized by crater chronology, is not significant enough to yield statistical distinction in the Moho uplift. More specifically, the thermal effects at the transient crater collapse (Potter et al., 2013) and post-impact viscoelastic relaxation (Kamata et al., 2015) stages are expected to compensate for each other, yielding limited net thermal effect on Moho uplift. Quantification of these two mechanisms requires additional numerical simulations and tests against observations. Analysis of individual impact craters with precise dating will also be important for revealing the effect of target temperature.

3.4 Model sensitivity and other influential factors

Although individual CBA and d_M values depend on the details of the gravity modeling and inversion, our statistical results are robust within the uncertainty level. Table 1 shows the sensitivity of the estimated parameters in the two-slope model and linear decision boundary to the Bouguer gravity models. Using a uniform crustal density of 2,550 kg/m³ (Model b) instead of a variable crustal density model to derive the Bouguer gravity does not influence the results. Applying high-pass filters (Bierson et al., 2016, Model c) and adopting a thicker crust (Model f) only influences the decision boundary coefficients c and d . Adopting a global reference radius of 1,738 km (Model d) or a varied reference radii that depends on local topography (Model e), both less than the mean reference radius of 1,762 km in the reference Model (a), yields systematically larger crater CBA (Figure 2), and thus lower D_{min} and d but larger a .

Coefficient c is the most important parameter in this analysis as it quantifies the effect of crustal thickness on the onset crater diameter with a Moho uplift. Considering that the decision boundary is also highly sensitive to the chosen significance level α (for classifying the positive and non-positive CBA groups, Figure 5d), the listed α , c and d values are for models that predict the largest positive c . A lower α corresponds to a more strict criterion to choose the positive CBA group, and thus the chosen α values are smaller in Models c–f in order to detect a decision boundary with the largest c . Comparing with these test models, our estimated c of 1.3 ± 1.0 for the reference model should be considered an upper limit.

For the coefficient e (Table 2), which quantifies the increase of the Moho uplift with crater size, the effect of using a uniform crustal density (Model b) is negligible. Assuming a greater mean crustal thickness (Model f), the inverted central Moho uplifts become larger, as the amplitude of inverted Moho undulations is highly sensitive to its assumed depth. The estimated coefficient e becomes accordingly larger. Application of lower $\lambda_{1/2}$ (Model g) in the gravity inversion algorithm filters out more high-frequency gravity information and thus lowers the inverted Moho uplift and coefficient e .

In addition to crater diameter, crustal thickness and thermal state, impact characteristics, including impactor energy (size, velocity and impact angle) and composition also influence the crater CBA and Moho uplift. Local variability in the target material properties, thermal state and crustal thickness not included in our analysis may contribute to the CBA variability. It is possible that the FHT, as a broad terrane that covers the majority of the lunar surface, experienced spatially variable thermal states (e.g., Zhang et al., 2014). The effects of crater age and non-homogenous thermal evolution on the Moho uplift requires more comprehensive numerical simulations and more precise dating of individual impact craters.

4 Conclusions

By linking the gravity signatures of large impact craters to the lunar crustal thickness, presence of mare deposits, and a thermal model that considers the difference between FHT and SPA terranes, we investigate the effect of crustal thickness and thermal state on the onset and magnitude of impact-induced Moho uplift.

The onset of the Moho uplift occurs at larger crater diameter in the context of thicker crust. The dependence on the crustal thickness implies structural uplift attenuation with depth. But the gravity observations reveal less structural uplift than previous numerical results of Potter et al. (2013), likely due to underestimated strength parameters in the hydrocode models.

The statistical similarity between the FHT and SPA craters suggests that the thermal difference of these two terranes is not significant and does not introduce a noticeable net difference in the relationship between Moho uplift and crater diameter. We find limited thermal effects on the crater diameter after considering the temperature-dependent crater diameter scaling (Miljković et al., 2016), SPA thermal models, and crater chronology functions. The thermal effect directly on the Moho uplift is thus required to be indistinguishable between FHT and SPA terranes as well.

Higher CBA observations exist for mare craters, and the difference between mid-sized mare and non-mare craters require 25–35% of the post-impact deposit column to be mare deposits in these craters.

Acknowledgments, Samples, and Data

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Figure 1. (a) Craters considered in this study located in the FHT (blue) and SPA (green) projected on a gray-scale LOLA topography map in a Mollweide pseudo-cylindrical projection centered on the farside. Dashed red circles denote mare craters and darkest patches are mare basalts. Dashed black curves outline the three major crustal terranes, including Procellarum KREEP Terrane (PKT). (b) Map of the first crustal thickness model of Wieczorek et al. (2013). (c) Measured central Bouguer anomalies (CBA) of craters. (d) Inverted Moho uplift (d_M) in the crater central region. Grid spacing of displayed meridians is 30° and that of parallels is 15° .

Figure 2. Observed Bouguer gravity anomaly (red curves) referenced to two different altitudes. The crustal structure (gray region) is an azimuthally averaged profile for the Balmer-Kapteyn crater to illustrate the definition of the crater CBA Moho uplift d_M , and local crustal thickness T_c . The reference altitudes are 60 km above the local Moho interface (Reference model a) and the local topography (Test model e). R is the crater radius.

Figure 3. Flow chart showing candidate control parameters on the crater CBA and central Moho uplift. Gray colors indicate observed or derived variables that are included in the crater database of this study, while white colors indicate parameters that can only be inferred. Parentheses of the symbols indicate slight difference between the word description and the chosen representative parameters. Solid arrows mean direct causal effects quantified in this study, whereas dashed arrows indicate potential influential factors that are either less significant (crustal porosity) or impossible to quantify (temperature and crater age).

Figure 4. (a) Topography and mare basalt distribution (black contours) for Poincaré basin; (b) Estimated maximum possible thickness of mare deposits; (c) Modeled gravity attraction due to (b) referenced to the local surface. The dark-gray circle indicates the central region ($<0.2R$) for CBA calculation. The surrounding annular region extends from 0.5 to $1R$ and is indicated by light-gray circles.

Figure 5. For non-mare FHT craters, (a) Two-slope model (blue curve and shaded region) for the crater CBAs (gray dots). (b) Zero-intercept linear regression (blue curve and shaded region) for central Moho uplifts (gray dots). Notable outliers are marked by their crater ID numbers in the crater database. (c) Linear discriminant analysis for the crater CBA: red dots are positive CBA group (p -value less than a significance level of 0.01); gray dots are non-positive CBA group. Black line is the estimated decision boundary with shaded uncertainty region. Cross marks indicate mis-classified points by the decision boundary. For example, a gray dot with cross mark is associated with non-positive CBA value but mis-classified in the positive CBA group by the decision boundary. (d) Sensitivity of decision boundary parameters to the sensitivity level, where a sensitivity level of 0.01 is the best for detecting a positive relationship between D_{min} and T_c (with the largest positive c value). Model and parameter uncertainties are standard deviations (i.e., 68% confidence intervals).

Figure 6. (a) Predicted structural uplifts at varied layer depths for a typical 300-km diameter crater. Structural uplifts for near-surface and deep materials are extrapolated (dashed curves). (b) Comparison between gravity-inverted crater Moho uplifts (dots), its corresponding two-slope model (blue line), and numerical results (dashed curves) from Potter et al. (2013). Structural

uplift of peak-ring materials for the Schrödinger basin and onset of structural uplift corresponding to the estimated decision boundary in Figure 5c are also shown.

Figure 7. (a) Comparison between mare (triangles) and non-mare (gray dots) FHT craters with a two-slope model for non-mare craters (blue curve and shaded region). The notable low-CBA outlier, Kohlschutter-Leonov (316), is probably not an impact crater but rather topographic depressions surrounded by thick impact ejecta (Byrne, 2016). (b) Residual CBA values (gray dots) and modeled maximum possible crater CBAs due to post-impact mare deposits: Red bars considers mare contributions to the gravity all over the crater region (Model *I*), while yellow bars correspond to the CBA due to mare within the central region only (Model *II*). Only positive residual CBAs are shown, and gray dots without corresponding color bars are mare craters with too limited mare coverage (e.g., $P_{\text{mare}} \sim 1\%$) within the impact basin to contribute significantly to the gravitational signature. (c) Residual CBA divided by the modeled CBA, which provides an estimation for the thickness ratio of the mare deposits.

Figure 8. Similar to Figure 5 but for SPA craters. Gray dots indicate non-mare craters, while gray triangles are mare craters. (a) Two-slope model and (b) zero-intercept linear model for SPA craters are shown in magenta color, which matches well with the results for FHT craters (blue color). (c) Decision boundary for SPA craters (dashed black line and light gray region) is not well determined due to the limited number of SPA craters. The decision boundary for FHT craters (solid black line and dark gray region) also well separate the positive and non-positive CBA groups of SPA craters.

Figure 9. (a) Modeled SPA thermal evolution (gray) that includes the increased heat from the SPA impact and the loss in radiogenic heating of a 15–35 km crust from the FHT thermal evolution path (blue), assuming a SPA formation age of 4.25 Ga. (b) Cumulative crater density from the specified time to present (with a crater diameter larger than 1 km per area of 1 km²) for monotonic declining chronology (solid) and sawtooth-like cataclysm chronology (dashed). (c) Average surface heat flux sampled by impact craters in the pre-Nectarian period as a function of the formation age of SPA, ranging from 4.1 to 4.45 Ga. (d) Corresponding crater diameter D_c^* for an impact with the same energy that would form a 300-km diameter FHT crater, in the hotter SPA region, plotted against the SPA formation age.

Table 1. Model sensitivity to Bouguer gravity calculation parameters

Model Settings	Terrane	a , mGal/km	D_{min} , km	α	c	d , km
<i>a. Reference Model</i> (Ref. to 60 km above local Moho)	<i>FHT (non-mare)</i>	0.58±0.15	197±30	0.01	1.3±1.0	150±39
	<i>FHT</i>	0.86±0.20	218±41	0.01	1.3±1.2	166±48
	<i>SPA</i>	0.81±0.11	218±19	0.01	2.0±1.0	175±31
<i>b. Uniform crustal density</i>	<i>FHT (non-mare)</i>	0.58±0.15	197±34	0.01	1.2±0.9	151±38
	<i>FHT</i>	0.85±0.21	220±44	0.01	1.3±1.2	167±48
	<i>SPA</i>	0.80±0.11	218±19	0.01	2.0±1.1	175±31
<i>c. High-pass filter $l_{min} = \frac{2\pi R_0}{4D_c}$</i>	<i>FHT (non-mare)</i>	0.59±0.15	195±32	0.002	0.6±1.0	178±41
	<i>FHT</i>	0.86±0.19	215±38	0.002	1.0±1.2	180±47
	<i>SPA</i>	0.79±0.10	214±17	0.002	1.9±1.4	172±37
<i>d. Ref. to 1,738 km</i>	<i>FHT (non-mare)</i>	0.67±0.10	152±16	0.038	0.8±2.2	127±83
	<i>FHT</i>	0.99±0.12	166±13	0.038	2.6±2.7	78±99
	<i>SPA</i>	0.95±0.05	145±11	0.038	4.2±1.3	87±33
<i>e. Ref. to (1 km above) local topo</i>	<i>FHT (non-mare)</i>	0.63±0.09	153±15	0.038	1.2±2.2	133±81
	<i>FHT</i>	0.96±0.14	167±17	0.038	3.3±2.5	53±93
	<i>SPA</i>	0.94±0.07	143±16	0.038	3.7±1.3	98±39
<i>f. Ref. to 70 km above local Moho (Mean crustal thickness = 45 km)</i>	<i>FHT (non-mare)</i>	0.57±0.13	193±30	0.007	0.6±0.9	172±44
	<i>FHT</i>	0.85±0.22	218±42	0.007	0.7±1.1	181±56
	<i>SPA</i>	0.80±0.10	219±18	0.007	2.0±1.0	156±39

Table 2. Model sensitivity to gravity inversion parameters

Model Settings	Terrane	e , 10^{-2}
<i>a. Reference Model</i> (Mean crustal thickness = 35 km)	<i>FHT (non-mare)</i>	7.5±1.0
	<i>FHT</i>	7.8±0.8
	<i>SPA</i>	9.1±1.1
<i>b. Uniform crustal density</i>	<i>FHT (non-mare)</i>	7.5±1.0
	<i>FHT</i>	7.9±0.8
	<i>SPA</i>	9.6±1.7
<i>f. Mean crustal thickness = 45 km</i>	<i>FHT (non-mare)</i>	8.7±1.2
	<i>FHT</i>	8.9±1.0
	<i>SPA</i>	12.5±2.3
<i>g. $\lambda_{1/2} = 80$</i>	<i>FHT (non-mare)</i>	6.8±0.7
	<i>FHT</i>	7.5±0.6
	<i>SPA</i>	9.2±1.1

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9