AN INVESTIGATION OF EARLY LUNAR HISTORY USING GRAVITY DATA

by

Weigang Liang

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We respectfully acknowledge the University of Arizona is on the land and territories of Indigenous peoples. Today, Arizona is home to 22 federally recognized tribes, with Tucson being home to the O’odham and the Yaqui. Committed to diversity and inclusion, the University strives to build sustainable relationships with sovereign Native Nations and Indigenous communities through education offerings, partnerships, and community service.
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Abstract

The Moon has fascinated humanity since ancient times. Recent technological and scientific advances have allowed us to gain significant information on the state of the lunar interior, revolutionizing our understanding of the Moon and its history. One such key method in studying the lunar interior is using gravity data. Gravity data can reveal anomalous structures that have minimal topographical expressions on the surface, regardless of whether the anomalous structure is located at a few kilometers beneath the surface, or beneath the crust-mantle boundary. In this thesis, I present new advances in the understanding and interpretation of the Moon and its history through the analysis of gravity measurements.

The Gravity Recovery and Interior Laboratory (GRAIL) mission has revealed scattered linear to arcuate gravity anomalies in both the lunar near and far sides, as well as a polygonal anomaly in the nearside that surrounds the Procellarum KREEP Terrane (PKT), an area of high radioactive element abundance. My investigations in Chapters 2 and 4 revealed fundamental differences in nature between the two types of anomalies, as the linear anomalies are found to be most consistent with dike-like features located at roughly 15 km beneath the crust, while the PKT anomalies are most consistent with remnants of ilmenite-bearing cumulates (IBCs) that did not sink during last stages of lunar mantle overturn. The global mantle overturn is a critical theory that has been used to address fundamental questions about the Moon such as the degree-1 asymmetry in radioactive elements, and my work provides the first ever physical evidence of the global mantle overturn.

In Chapter 3, I investigated the mechanisms responsible for the deficit in the crater size frequency distributions of the nearside mare accounting for buried craters observed in the gravity data, in comparison to those of the farside highlands. This deficit is proposed to be related to the missing gravitational signatures of the rings of the Imbrium basin where they are buried beneath the maria. Thermal erosion of substrate topography due to mare flooding was found to be the explanation that is most consistent with the observations. These findings greatly contribute to our understanding of the Moon's thermal and tecto-magmatic evolution, shedding light on crucial parts of lunar history.
Chapter 1: Introduction

The conventional view of the history of the Moon after its accretion is that it experienced a global magma ocean (Elkins Tanton et al., 2002; Elkins-Tanton et al., 2011; Hess & Parmentier, 1995). The first solids to crystallize from the global magma ocean are olivines and pyroxenes, which are denser than the liquid and sink towards the base of the magma ocean. Later in the sequence, plagioclase begins to crystallize, which is less dense than the liquid and rises to the top to form a floatation crust. As the olivines and plagioclases approach complete crystallization, the remaining liquids will be saturated in ilmenite and Fe-rich clinopyroxenes (e.g., Lin et al., 2017; Snyder et al., 1992), as well as the incompatible elements including potassium, rare-earth element, and phosphorus, or KREEP, including the heat generating element thorium.

The solids that precipitate from the ilmenite-saturated liquid are designated as “ilmenite-bearing cumulates” (IBCs), and a global IBC layer forms at the end of magma ocean crystallization. Since ilmenite and clinopyroxenes are denser than the olivine orthopyroxene mantle, the configuration is gravitationally unstable. The gravitational instability is thought to have resulted in global mantle overturn (Hess & Parmentier, 1995; Zhong et al., 2000). The overturn and sinking of the IBCs has been proposed to address fundamental lunar features including the present-day nearside seismicity (Nakamura, 2003), cessation of the lunar dynamo (Stegman et al., 2003), and with more relevance to this work, the degree one nearside-farside asymmetry in KREEP distribution (Zhang et al., 2022) and the widespread nearside high-Ti volcanism (Zhang et al., 2013; Zhong et al., 2000).
Commencing from at least ~3.6-3.7 Ga (Hiesinger et al., 2011), widespread volcanism is thought to have occurred in the nearside mare region. Many of these basalt flows are high in titanium abundance (Lucey et al., 1994; Sato et al., 2017) (Fig. 1.1b) and/or thorium abundance (Prettyman et al., 2006) (Fig. 1.1c). The nearside mare region is co-located with the Procellarum KREEP Terrane (Jolliff et al., 2000), which is a region of especially high KREEP abundance. Such a region is absent in the farside region, which presents a fundamental question of why there is an asymmetry in KREEP and ilmenite, when the initial IBC layer which is enriched in KREEP is predicted to be global. Models of a global mantle overturn (e.g., Jones et al., 2022; Zhang et al., 2022) can be used to explain the degree 1 asymmetry in KREEP elements and the high-Ti basalts, as the mantle overturn would result in the sinking of the dense IBC into the mantle, after which the thermal gradient generated from the heat of the radioactive elements would overcome the negative density gradient and result in the ascension of Ti-rich basalts (e.g., Zhang et al., 2013). However, while global mantle overturn has been proposed to address numerous fundamental questions in lunar evolution, currently there is a definite lack in identified physical evidence for the global mantle overturn.

At the same time as the magma ocean crystallization and the subsequent mantle overturn, the surface of the Moon was bombarded by impacts. Currently, the oldest known basin is the South Pole Aitken (SPA) Basin, with an estimated age of 4.2-4.3 Ga (Garrick-Bethell et al., 2020; Orgel et al., 2018). Due to the combination of the declining rate of bombardment and the size-frequency distribution of the impactors, the largest impacts occurred during early lunar history. Impactors resulting in craters greater than roughly 500 kilometer in diameter impacted the lunar near and farsides throughout 4.2-3.8 Ga (Orgel et al., 2018), forming basins such as
Serenitatis and Imbrium, which are still prominent features when observing the Moon. On the other hand, smaller impactors bombarded the lunar surface throughout lunar history until present day, resulting in lunar highland surfaces reaching saturation at numerous diameter bins (Head et al., 2010; Povilaitis et al., 2018). Orientale and Imbrium are the youngest major basins, with Imbrium being the largest impact basin in the lunar nearside region.

Subsequent to the decline in bombardment, the much of the nearside was flooded by eruptions of mare basalt. The widespread nearside volcanism has resulted in the mare flooding of major basins such as Imbrium, Serenitatis, and Humorum, and the plains within the nearside mare contain flows which surfaced as recent as 1 Ga (Hiesinger et al., 2011). The extensive mare flooding of the Imbrium basin also cause it to appear much different from Orientale, despite their similarities in age and dimensions. The typical mare thickness in the nearside is ~1 km (DeHon, 1979; Evans et al., 2016; Gong et al., 2016; Head, 1975), and thus it is difficult to gain insight into the topography and surface features of the ancient nearside mare region before the mare flooding events.

In this work, I use gravity data to shed light on the history of the Moon. Gravity measurements can provide insight into the buried craters, missing Imbrium rings, and the global mantle overturn, as gravity measurements are able to detect the gravitational signal from the subsurface, and any deviation from the expected gravity implies structures of mass excess or mass deficits within the subsurface.

Gravity data is commonly represented as spherical harmonics:

\[
gr = \frac{GM}{r^2} \sum_{l=0}^{\infty} \sum_{m=-l}^{l} \left( \frac{R_0}{r} \right)^l (l + 1)C_{lm}Y_{lm} \tag{1.1}\]
\( Y_{lm} = \begin{cases} \bar{P}_{lm} \cos(\theta) \cos(m\varphi) & \text{if } m \geq 0 \\ \bar{P}_{|l|m|} \cos(\theta) \sin(|m|\varphi) & \text{if } m < 0 \end{cases} \) (1.2)

where \( P_{lm} \) are the associated Legendre polynomials (Wieczorek 2015). The \( l \) term, or the spherical harmonic degree, is correlated with the horizontal wavelength of the anomalies. Free-air gravity contains information on mass excesses or deficits from both the surface and the subsurface structures. Then, Bouguer gravity can be calculated from free-air gravity by subtracting the gravity arising from the surface topography assuming a density for the surface material. Bouguer gravity contains the gravity variations, and consequently the mass variations, from only the subsurface structures. Features such as mantle uplifts due to large impacts, craters buried by mare, as well as intrusive volcanic features such as dikes and sills are prominent positive (>\( \sim \)150 mGal, where 1 Gal = 0.01 m/s\(^2\)) features in the lunar Bouguer gravity (Fig. 1.1d). In this study, Bouguer gravity gradients (Fig. 1.1e), which are the maximum amplitude eigenvalues of the tensor of the second horizontal derivatives of the potential, and which highlight density variations at shorter wavelengths, are used as the primary expression of data (Andrews-Hanna et al., 2013).
Figure 1.1: Cylindrical map projections centered at the nearside-farside separation of the lunar (a) topography, (b) TiO$_2$ abundance (Sato et al., 2017), (c) Thorium abundance (Prettyman et al., 2006), (d) Bouguer gravity in units of mGal (10$^{-5}$ m/s$^2$), and (e) Bouguer gravity gradient in units of Eotvos (10$^{-9}$ s$^{-2}$). The gravity maps were calculated at a Bouguer correction density of 2550 kg/m$^3$, and low-passed at spherical harmonic degree 150 to highlight the low-degree features, which are the focus of this work. Units are in mGal is and E.
The gravity data used in this work was obtained by the Gravity Recovery and Interior Laboratory (GRAIL) mission (Zuber et al., 2013). The Gravity Recovery and Interior Laboratory (GRAIL) mission (Zuber et al., 2013) returned gravity data of unprecedented precision and resolution. A major feature revealed by the GRAIL gravity maps is the population of linear to arcuate positive gravity anomalies and negative gravity gradients, implying a region of higher density in comparison to the surroundings within the subsurface (Fig. 1.1e). The linear gravity anomalies can be categorized into two types. The first type is the scattered linear anomalies that are present in both the nearside and farside regions. The second type is the linear anomalies surrounding the PKT region that form a polygonal structure.

Chapter 2 focuses on investigating the gravity power spectra as a function of spherical harmonic degree of the two types of linear anomalies, in order to ascertain if they share a common origin. Those analyses find spectral evidence indicating a difference in origin between the two types of anomalies. I then use Monte Carlo methods and the power spectra of the linear anomalies to constrain the physical dimensions as well as depths of the linear anomalies. Chapter 4 will address the second type of linear gravity anomalies, the PKT anomalies, and investigate the hypothesis that the PKT anomalies are remnants of the IBC downwellings at the end stages of the global mantle overturn. I compare observed gravity data of the anomalies to the forward-modelled gravity of predicted remnant IBC layers in mantle overturn models (Zhang et al., 2022). In addition, I use Monte Carlo methods, as well as inversions of observed topography and gravity, to constrain the dimensions of the proposed IBC structures.

Another major feature revealed by GRAIL gravity are craters whose topographical signatures have been erased (Evans et al., 2016) due to burial by overlaying mare. As the
nearside and farside surfaces are expected to be exposed to the same population of impactors for roughly the same period of time, the crater size-frequency distributions (CSFD) of lunar craters on both the near and far sides should be equal if the signal of craters buried beneath the mare were also included. However, a deficit in CSFD for craters of less than 90 km in diameter is observed (Evans et al., 2016, 2018) for the nearside mare in comparison to the farside highlands. In addition, the rings of Imbrium, which have minimal signal in topography due to the mare flooding, similarly have minimal signal in gravity, in stark contrast to the topography and gravity of Orientale and its rings. This observation is surprising since simple mare flooding of ring structures resembling those around Orientale which are up to 4 km in height and 100 km in width, should result in strong gravity anomalies that are lacking in the data. Chapter 3 will propose and test the possible mechanisms resulting in the deficit in nearside craters, and the missing rings of Imbrium. The proposed mechanisms include an effect of a greater mare thickness than previously assumed, a lower density contrast between the mare and substrate, a dense pre-mare layer reducing the gravity from buried craters, or some manner of erosional event affecting the buried crater and basin ring topography. The results favor the latter mechanism, which would constitute erasure of an unprecedented scale.

This work focuses on the quantification and interpretation of features in lunar gravity data including the prominent gravity anomalies, basin rings, and buried craters, and uses them to address key questions in lunar science. Gravity offers a unique perspective into the subsurface of the Moon as it reveals structures not visible on the surface, and as a result new understandings of lunar history and evolution can be obtained. The implications of my work provide further insight into key moments in lunar history, such as the end of the lunar global mantle overturn, the early
farside magmatism, as well as the widespread nearside mare flooding whose results are still visible today.

References


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Chapter 2: Probing the source of ancient linear gravity anomalies on the Moon


2.1 Abstract

A large set of linear to arcuate gravity anomalies which are not associated with any surface expressions was revealed by lunar gravity data from the GRAIL mission. The anomalies can be categorized into two types: a set of large anomalies that border the Procellarum KREEP terrane (PKT) region, and the linear gravity anomalies (LGAs) that are scattered throughout the lunar highlands. In this study, we use band-passed gravity gradient maps and localized power spectral analyses to characterize the nature of and the differences between the two types of anomalies. The results show that the PKT border anomalies are primarily long wavelength features whose power spectra are not clearly distinguishable from the background spectrum of the surrounding region, while the linear gravity anomalies are short wavelength features whose power spectra in at least one case rises significantly above the background spectrum over a wide range of degrees. This difference in gravitational signature suggests a fundamental difference in the nature of the sources of the two types of anomalies. We then used a Markov chain Monte Carlo model to test different interpretations for the most prominent LGA by comparing the observed power spectrum to modeled spectra for elliptical, triangular, and T-shaped intrusive geometries. We find multiple geometries are able to fit the power spectrum of the LGA equally.
well. Analogs of comparable scale to the LGAs originate from extensional stress regimes, supporting an inferred period of global expansion in the early Moon, though the cause of that expansion remains uncertain. If the depth extent of the intrusions were governed by neutral buoyancy, we find that intrusions on the nearside similar to those on the farside would have been eruptive and may have contributed to the earliest mare volcanism. The total volume of the intrusions is ~20% of the mare volume, revealing a lower magma production rate on the farside than on the nearside.

2.2 Introduction

The thermal and geodynamic evolution of the Moon is expressed in the record of volcanic, magmatic, and tectonic activity. Mare volcanism provides a clear window into the volcanic history of the Moon. However, it is expected that this extrusive activity would have been accompanied by intrusive activity that is hidden from view (Head & Wilson, 1992). Investigations into the nature of that intrusive activity have the potential to improve our understanding of the magmatic, tectonic, and thermal evolution of the Moon. Evidence of such intrusive activity includes floor-fractured craters (Jozwiak et al., 2012; Wilson & Head, 2018) and dike-induced graben (Head & Wilson, 1993). Gravity analyses are capable of directly detecting unseen subsurface magmatic structures, as a result of the large density contrast expected between the high-density intrusive bodies and the low-density lunar crust. The Gravity Recovery and Interior Laboratory (GRAIL) mission (Zuber et al., 2013), which orbited around
the Moon from 2011 to 2012, measured the lunar gravity field to an unprecedented precision, providing the first gravity dataset capable of resolving intrusive structures in the crust.

Bouguer gravity gradient maps, which highlight short wavelength structures, revealed a quasi-rectangular pattern of linear to arcuate anomalies on the near side (Andrews-Hanna et al., 2014, Figure 1). These anomalies approximately follow the border of the Procellarum region, a low elevation region with thin crust that encompasses the majority of the nearside maria as well as the Procellarum KREEP Terrane (PKT). This region was interpreted in early studies as a possible impact basin (e.g., Cadogan, 1974; Whitaker, 1981), but gravity data reveal it to instead be a magmatic-tectonic province (Andrews-Hanna et al., 2014). The PKT border anomalies were interpreted as volcanically flooded rift zones possibly formed in response to the cooling of the region, though their nature, origin, and significance remain uncertain. A study (Zhu et al., 2019) using hydrocode impact simulations does show that an impact origin can leave subtle crustal signatures and result in KREEP-rich terrain, but does not provide an explanation for the PKT border anomalies.

Another class of structures that are observed in the gravity data but have minimal surface expression is the randomly oriented linear gravity anomalies (LGAs), which occur in large numbers on both the near and far sides (Andrews-Hanna et al., 2013). The narrow widths (tens of km), extreme lengths (hundreds of km), and linear paths of the anomalies are consistent with expectations for dike-like bodies formed in a horizontally extensional stress regime. Both the PKT border anomalies and the LGAs are expressed as elongated, linear to arcuate positive gravity anomalies and negative gravity gradients (Andrews-Hanna et al., 2013; 2014). Despite
some superficial similarities, the PKT border anomalies and LGAs were interpreted as two distinctly different classes of structures based on their different geological settings and their appearance in different representations of the gravity gradients, but this distinction has not yet been quantitatively tested.

Previous inversions of the linear gravity anomalies found very large structures with best-fit widths of 6.7-18.0 km and best-fit heights of 35-81 km, assuming a density contrast of 550 kg/m³ (Andrews-Hanna et al., 2013). These large widths are unlike the widths of conventional dikes or even dikes within giant dike swarms (>10 m; Ernst et al., 2001), making the geological interpretation of the anomalies uncertain. The near-equant aspect ratios are also unlike those of typical dikes with aspect ratios of ~300–1500 (Gudmundsson, 1984), suggesting that the geometry of the bodies is not strongly influenced by the elastic deformation of the surrounding rock mass (e.g., Mastin and Pollard, 1988). Thus, other geometries and interpretations should also be considered.

The shape of an intrusive body has implications for its formation mechanism. Elliptical density anomalies represent the shape of a typical dike (e.g., Buck et al., 2006), in which dike dilation is resisted by the elastic deformation of the surrounding rock mass. Alternatively, the Great Dyke of Zimbabwe (Wilson, 1996), which is perhaps the best terrestrial analog to the scale of the LGAs (Andrews-Hanna et al., 2013), is a broad but elongated layered intrusion. The Great Dyke is a 2.46 Ga old, 550 km long, 1.90-3.35 km high, and 4-11 km wide linear structure with an approximately triangular cross-section that tapers downwards. Its specific mechanism of formation is debated, but most interpretations suggest that the Great Dyke formed during crustal
extension (Wilson, 1996). Alternatively, the LGAs may be explained by a dike transitioning into a sill, like the exposed intrusions within the Las Torres del Paine National Park in Chile (Kavanagh et al., 2017), providing a possible means of explaining both the widths and lengths of the anomalies while avoiding the extremely wide dike-like intrusions required by the elliptical geometry. Such a scenario could arise from a decrease in buoyancy within the intrusive network resulting from the decrease in density in the upper crust (Besserer et al., 2013), from an increase in the rigidity or fracture toughness in the rock as the dike ascends in the crust (Kavanagh et al., 2017; Gudmundsson, 2011), or perhaps from a decrease in the extensional stress in the weak impact-fractured and brecciated upper crust (Besserer et al., 2013). For a dike-sill hybrid, the path of the structure in map view may be controlled by the linear propagation of the dike, while the width and magnitude of the gravity anomaly may be strongly influenced by the overlying sill.

In this study, we use a set of spectral and spatial analyses of the gravity data to characterize the PKT border anomalies and the LGAs, and to shed light on the underlying source bodies. We first address the question of whether these two populations of gravity anomalies are clearly differentiable from each other and thus likely to represent distinctly different populations of structures. Band-passed gravity maps and localized power spectra are used to examine the extent to which they differ from the background and from each other. Next, we consider the geometry of the source bodies of the LGAs, and the large widths and low aspect ratios found in previous work. For the one LGA with a power spectrum that significantly rises above the background, we invert the power spectrum for the geometry of the underlying intrusive body using Markov Chain Monte Carlo (MCMC) analysis. This spectral-domain approach is complementary to the spatial-domain approach taken previously, and is expected to be more
sensitive to the top depths and shallow structure of the anomalies due to the strong decay of gravity anomalies with depth as a function of wavelength. We also use higher resolution data from the GRAIL extended mission (Konopliv et al., 2014), which will better constrain the nature of the sources of the anomalies. The inversions test a range of possible geometries to explore different possible analog structures and examine the sensitivity of the dimensions to the assumed geometry. The results of these analyses are then used to shed light on the nature of the underlying bodies, their formation, and the implications for lunar evolution.

2.3 Data and Methods

2.3.1 Band-Passed Gravity Gradient Maps

Both the LGAs and PKT border anomalies were first identified in Bouguer gravity gradient maps, which emphasize shorter wavelength features such as discrete structures and edges in the gravity field in comparison to the Bouguer gravity anomaly (Andrews-Hanna et al., 2013, 2014). That previous work found that the LGAs feature prominently in gravity gradient maps with a high-pass filter applied at spherical harmonic degree 30, while the PKT border anomalies only appear when the low degrees are included in the data. This difference supported the different interpretations of the LGAs and PKT border anomalies. However, beyond this simple comparison, no detailed study of the difference in scales and wavelengths between these two classes of features has been conducted. The spectral signature of the gravity anomalies is important as it provides information on the nature of the source bodies. Longer wavelength anomalies likely arise from deeper and/or smoother source bodies (e.g., features below the crust,
or shallower features that are inherently smooth at these scales). Shorter wavelength anomalies must arise from shallower and narrower source bodies (e.g., intrusions within the crust).

While localized gravity power spectra and associated quantities (e.g., admittance) are commonly used to examine the spectral nature of gravity anomalies (e.g., Besserer et al., 2014; Goossens et al., 2017), distinguishing the power spectrum of a gravity anomaly from the background can be difficult for small-scale and low-amplitude features such as the LGAs (see next section and results below) as a result of the strong small-scale variability in the gravity field (Jansen et al., 2016). The lack of topography associated with these gravity anomalies precludes the use of admittance to characterize the anomalies. Band-passed gravity maps are the simplest means of identifying the characteristic wavelengths of gravity anomalies. Thus, we generate a set of band-passed gravity gradient maps in order to determine the range in spherical harmonic degrees over which the anomalies are expressed.

Analyses were performed on the spherical harmonic degree 1200 GRAIL GRGM1200A free-air gravity model (Lemoine et al., 2014) and a spherical harmonic degree 2600 LOLA topography model (Smith et al., 2010; Wieczorek, 2015). The free-air gravity data was processed using two different Bouguer corrections (Fig. 2.1), in order to account for the difference in surface density between the feldspathic crust at 2550 kg/m$^3$ (Wieczorek et al., 2013) and the maria at 3100 kg/m$^3$ (Kiefer et al., 2012). For simplicity, global maps show the entire nearside using the maria Bouguer correction density, and the farside using the feldspathic crust Bouguer correction. Shallow density anomalies in the upper crust likely exist in a layer roughly following the surface topography. To minimize the effects of surface topography on the gravity data, we
calculated the gravity and gravity gradients on a smoothed representation of the surface topography rather than on a spherical surface (Andrews-Hanna et al., 2018). The algorithm calculated the potential at the local surface elevation at every point, using a spherical harmonic degree 30 representation of the surface topography to avoid abrupt jumps in elevation. We then generated a spherical harmonic representation of the gravity on the surface in order to enable subsequent analyses. The gravity gradients were calculated using the maximum amplitude eigenvalue of the tensor of the second horizontal derivatives of the potential (hereafter referred to as the gravity gradients; Andrews-Hanna et al., 2013).

The gravity gradients were band-passed to filter out all components above or below a designated spherical harmonic degree as high-pass and low-pass maps, respectively. The threshold degrees used in both sets of maps were varied between 10 and 260 in 10-degree increments. For the high-pass gravity gradient maps, a low-pass taper was applied at degree 500 in order to avoid the orbit-parallel striping in the data at high degrees in some locations. Using this method, the specific degree windows in which each of the anomalies is distinguishable from the background can be visually evaluated. The band-passed gravity gradient maps are useful for a qualitative spectral characterization of the gravity anomalies, particularly in areas in which the background variability in the gravity field may hamper spectral analysis, and they do not suffer from the limitations imposed by the bandwidth of spatio-spectral localizations. As will be seen below, an anomaly may be distinct in the band-passed maps even if it has a power spectrum that is not distinguishable from its surroundings, as the power spectrum does not account for the spatial organization of the anomaly visible in the maps. However, for quantitative analysis of the source of the gravity anomalies, localized power spectra are needed.
2.3.2 Power spectra

Localized power spectra of the Bouguer gravity were then used to characterize and compare the LGAs and PKT border anomalies. The Spherical Harmonic Tools (SHTOOLS) software archive (Wieczorek & Meschede, 2018) was used for localization. The localization process produces a set of spherical harmonic coefficients that optimizes both spatial and spectral concentration of an area defined by a Boolean mask and uses the coefficients to generate the localized power spectrum of that area (Wieczorek & Simons, 2005). For the LGAs and the PKT border anomalies, the localization masks were generated as the collections of points within a distance of 100 km and 160 km, respectively, from a great-circle path defined by the two endpoints of the anomalies. The greater width of the masks for the PKT border anomalies were required by the greater widths of the anomalies themselves. We defined nine localized areas in total (Fig. 2.1, outlines). The nearside areas consisted of portions of the four nearside PKT border anomalies referred to as the nearside northwest (NNW), nearside northeast (NNE), nearside southeast (NSE), and nearside south (NS) anomalies. On the farside, only the two most prominent LGAs, referred to as the farside north (FN) and farside south (FS) anomalies, were chosen for analysis, as other anomalies are both shorter and lower in amplitude. The FN anomaly occurs in a typical highlands area, and shows no clear relationship to any recognized surface features. In contrast, the FS anomaly is one of several anomalies radial to the South Pole-Aitken basin (SPA).

The power spectra of the anomalies must be compared with power spectra of background areas in similar settings that are devoid of any prominent gravity anomalies associated with
known discrete structures. For the farside northern and PKT background areas, 18.0° by 18.0° boxes within their respective regions were defined as the background masks (see Figure 2.1 for locations). The nearside background (NBG; centered at 4.31° S, 9.7° W) was chosen to avoid the variability in the gravity field in the PKT as a result of the large gravity anomalies associated with impact basins (Neumann et al., 2015), buried craters (Evans et al., 2016), and other structures buried beneath the maria. The farside north background (FNBG) was centered at 27.1° N, 202° E. The FN and FS anomalies are located in vastly different geological and geophysical settings (in the highlands and in the rim area of the South Pole-Aitken basin, respectively) and thus different background areas were needed for comparison. The farside south background (FSBG; centered at 22.3° S, 198° E) is located in a different area of the SPA rim. This background area was defined using the same method as for the FS anomaly, with the radius of the mask set at 160 km, in order to capture an area with a similar geometry relative to the SPA basin given the large variability in the gravity gradients around the basin. Additional localization masks were drawn 3° north and south of the FN anomaly and their spectra were used to quantify the local background variations in power for the Monte Carlo model (see next section).
Figure 2.1. Concatenated Bouguer gravity gradient maps calculated for surface densities of 2500 and 3100 kg/m$^3$ for the farside (left) and the nearside (right), respectively. Localization areas are outlined and labeled.

The localized power spectra were then generated in SHTOOLS with bandwidths of 40 and 80 degrees for the PKT border anomalies and LGAs, respectively, using all available well-concentrated tapers (eigenvalues $> 0.98$ for the PKT border anomalies, $> 0.99$ for the LGAs). The bandwidths were chosen to maximize the spatial concentration and to minimize the loss of signal at the start and end of the spectral range. The lower bandwidth and lower eigenvalue
threshold for the PKT border anomalies was required by the range of degrees in which they are expressed in the band-passed gravity gradient maps. From the band-passed gravity maps, we have found that much of the power in the PKT border anomalies originates from below degree 50, while power from above degree 50 dominates the LGAs (see below). Thus, the bandwidth for the PKT border anomaly localization must be less than 50. A bandwidth of 40 was chosen as so to still provide adequate spatial localization, but we acknowledge that some loss of signal at low degrees results from this choice. All background areas were localized using the same bandwidths and eigenvalue thresholds as the anomalies with which they are compared. Each taper produces a localized estimate of the power spectrum, and the final spectrum is generated as the average of the estimated spectra. Due to the differences in area and shape of the regions of interest and the differences in bandwidth, the number of tapers that satisfied the eigenvalue requirement ranged from two to four for the nearside areas, and four to twenty-two for the farside regions. Although the localization windows for the anomalies differ in size and shape from the localization windows of the background areas, these differences do not produce a bias in the resulting power (see Supplementary Material).

2.3.3 Subsurface Modeling

Next, we considered the source of the FN LGA, the one case in which the gravity power spectrum of an LGA clearly rose above the background. The nature of the density anomaly responsible for the LGA was examined by modeling the power spectra of candidate subsurface structures and comparing them with the observed spectra. As in the previous work (Andrews-Hanna et al., 2013), we use a Monte Carlo model to constrain the properties of the subsurface
structure. Our approach differs from the previous work in that we are using the power spectrum rather than spatial profiles of the gravity anomalies for inversion (Andrews-Hanna et al., 2013). Both approaches are valid, but they emphasize different aspects of the data. Power spectral approaches are expected to be more sensitive to the shallow structure that will dominate the higher degrees. Moreover, averages of gravity anomaly profiles taken orthogonal to the structures have the potential to smooth out the anomalies due to variability in the structures, which could lead to artificially deeper and broader source bodies in the inversions. In the absence of noise or spatial variability in the structure, if a unique solution exists, both methods should arrive at the same conclusion.

Modeled subsurface structures were represented as combinations of rectangular prisms, for which there is an analytic solution for the gravity anomaly (Nagy, 1966). The intrusions were modeled in a Cartesian geometry, but considering the narrow widths of the structures (10’s of km) relative to the radius of curvature of the Moon, together with the efficiency of this analytic solution, this approximation is justified. Previous work considered only a simple dike-like geometry (Andrews-Hanna et al., 2013), but the large widths and near equant aspect ratios found in that study suggest that other geometries and interpretations should also be considered. We considered three possible subsurface geometries: elliptical, triangular, and T-shaped density anomalies (Fig. 2.2). The elliptical density anomalies represent the shape of a typical dike (e.g., Buck et al., 2006) and consisted of 1000 rectangular prisms of constant lengths (parallel to the strike of the anomaly) and widths (horizontal dimension perpendicular to the strike of the anomaly), but with varying heights (vertical dimension) to provide an elliptical cross-section. The triangular density anomalies were chosen to represent the shape of the Great Dyke of
Zimbabwe (Wilson, 1996), which is perhaps the best terrestrial analog to the scale of the LGAs (Andrews-Hanna et al., 2013). The triangular anomalies narrow with increasing depth, and were created by overlaying 1000 rectangular prisms of uniform height and length, but with a linearly decreasing width. The T-shaped density anomalies were chosen to represent the possibility of a sill formed on top of a dike, and to test if such a structure can match the observed gravity with a more typical width dike coupled with a wider sill near the surface. Each T-shaped anomaly was represented as a combination of two regular rectangular prisms, with the bottom prism extending to the base of the crust at a depth of 50 km.

**Figure 2.2.** A visual representation of how the elliptical (left), triangular (center), and T-shaped (right) geometries are discretized in the model. The actual model discretization was much finer than depicted here (see text for details).
To find the best-fit geometry of the density anomalies, we used the Metropolis-Hastings Markov chain Monte Carlo algorithm (Chib & Greenberg, 1995). Both the elliptical anomalies and the triangular anomalies were characterized by the depth to the top of the anomaly, depth to the base of the anomaly, and width of the anomaly. T-shaped anomalies were characterized by the depth to the top of the anomaly, the depth to the transition between the two prisms, the width of the top prism, and the width of the bottom prism. The horizontal length of all prisms parallel to the anomaly strike was set to 350 km, such that the ends did not have any impact on the gravity profile at the midpoint. Our prior constraint set the maximum bottom depth to be 50 km, which is roughly the modeled crustal thickness in the farside highlands surrounding the FN LGA (Wieczorek et al., 2013). In addition, the maximum width considered in the a priori distribution was 68 km as constrained by the localization bandwidth of degree 80.

In order to optimize the efficiency of the MCMC runs, the Moon was rotated using spherical harmonic transformations so that the anomaly is located on and parallel to the equator. This permits us to only model a single gravity profile perpendicular to the anomaly at the midpoint in the latitudinal direction and assume that the gravity is uniform across longitude. In this way, the Cartesian form of the analytical solution of the gravity of the prism would be an appropriate representation of the gravity of the prisms, as the scale of interest is much smaller than the radius of curvature of the Moon. From the modeled gravity profile, we construct a global gravity map and localize the modeled anomaly following the same procedure used for the data. While the power spectrum of the model anomaly could be examined in the Fourier domain, the above approach ensures that the data and model are processed and localized in the same manner for purpose of comparison. In addition, the power spectrum of the respective background area
(FNBG) is added to the modeled anomaly spectrum in order to account for the pervasive random variability in the gravity field. Models were evaluated using a likelihood function, $L$:

$$L = e^{SSD/\sigma^2}$$

(1)

in which $SSD$ is defined as the summed squared difference of the modeled and observed gravity power spectra, and $\sigma$ is the root-mean-squared difference between the power spectra of the localized regions just north and south of the anomaly. Since the error in the GRAIL data is much lower than the background variability in the field (Jansen et al., 2016), $\sigma$ is used in place of the data error commonly used in such algorithms (Chib & Greenberg, 1995). For the elliptical and triangular prisms, the MCMC simulations were run using 10,000 iterations over 3 parallel chains using fixed starting points in very different parts of the parameter space. The elliptical and triangular anomalies converged on the same solution for all starting points. The T-shaped anomaly has a considerably larger phase space than the other geometries and as a result the MCMC simulation for this anomaly was run using 8 parallel chains with random initial parameters, whose values were constrained to be within an allowable range based on preliminary results (top depth $< 30$ km, bottom depth $< 50$ km, bottom prism width $< 50$ km, and bottom prism width $< $ top prism width $< 68$ km).

While our primary focus is on the spectral domain inversion of the FN LGA, we also used the MCMC model to fit average spatial profiles of the Bouguer gravity anomaly similar to those of Andrews-Hanna et al. (2013). We analyzed four LGAs (Fig. 2.7): the FN and FS anomalies, an anomaly northwest of the FS anomaly (designated as LGA 3), and an anomaly that connects the Serenitatis and Crisium basins (designated as LGA 4). Localized power spectral
analyses were not performed as the signals of these anomalies relative to the surrounding background are weaker than that of the FN anomaly. We excluded from this analysis LGAs similar in length but with greater departures from linearity, as well as a number of shorter LGAs. Test analyses of anomalies with greater departures from linearity found artificially rounded and widened average profiles, resulting in erroneously deeper and wider density anomalies. These models only considered the elliptical geometry, as preliminary results found that more complicated triangular and T-shaped geometries do not improve the fit to the data. These spatial domain inversions serve as a test of the spectral domain inversion, while providing additional constraints on the dimensions of the source bodies of a larger number of LGAs.

2.4 Results

2.4.1 Band-passed gravity gradient maps

Band-passed gravity gradient maps provide visual insight into the degree ranges over which the anomalies are distinguishable from the background. We note that the Bouguer correction density of 3100 kg/m³ chosen for the nearside is appropriate for the maria, but results in large anomalies in the nearside feldspathic terrain, including the Imbrium rings. Our results show that the gravity signal of the nearside PKT border anomalies (NW, NNE, NSE, NS) is concentrated in the low degrees. The pentagonal (rectangular in a properly projected map) shape of the border anomalies can be seen in the degree 20 low-pass map (Fig. 2.3, deg 20) but not in the degree 10 low-pass map, which indicates the anomalies begin to distinguish themselves from the background between degrees 10 and 20. This is supported by the fading of portions of the NE, NSE, and NS anomalies in the degree 20 high-pass map (Fig. 2.4, deg 20) compared with the
degree 10 high-pass map. From degrees 20 to 40, the low pass maps show increasing detail at higher degrees, while the high pass maps show increasing detail at lower degrees, indicating that the anomalies have appreciable signal from degrees 20 to 40 as well. At degree 40, only the NNE and NW anomalies are visible in the high-pass maps, and these largely disappear by degree 60, with only small segments visible up through degree 90. Beyond degree 90, high-pass maps of the nearside PKT region do not show any evidence for the border anomalies. These results indicate that the majority of the signal from these anomalies lies between degrees 10 and 40-60, corresponding to half-wavelengths of roughly 545 km to 90-135 km. Isolated sections contain signal extending to degree 90, corresponding to structures at scales as small as 60 km.

In contrast to the PKT border anomalies, our results show that the expression of the farside linear gravity anomalies is concentrated in the middle to high degrees (50-250). In the low-pass maps, the FS anomaly first appears at degree 30 (Fig. 2.3), and largely disappears in the high-pass maps at around degree 120 (Fig. 2.4) though a short segment remains visible out to higher degrees. The FN anomaly first appears in the low-pass maps at the slightly higher degree of 50-60, and disappears from the high-pass maps at degree 250. The eastern half of FN fades at a lower degree range of 130-140 in the high-passed maps, suggesting a possible lateral change in the properties (e.g., depth or shape) of the source of the anomaly. These results show that the FS anomaly exhibits considerable gravitational signal between degrees 30 and 120 (half-wavelengths of 181 and 45 km, respectively), while the FN anomaly covers a wider range of degrees of 50 to 250 (half-wavelengths of 109 and 22 km, respectively). A large number of smaller LGAs exist, which typically fade into the background for high-pass filter cutoffs of 100-120.
The clear difference in the spatial scales over which the LGAs (22-110 km) and the PKT border anomalies (90-545 km) are expressed implies a fundamental difference in the sources of the anomalies. Given that longer wavelengths often arise at greater depths, while short wavelengths must arise at shallow depths, the difference in degree range may suggest a difference in the depth of origin. Adopting an approximate relationship between spherical harmonic degree and the maximum depth of the source anomaly as a guideline (Bowin, 1983), the continued power of the LGAs at degrees up to 120 and 250 indicates that the upper portions of the source bodies likely reach depths shallower than 7–15 km. While shallower source depths are possible, we do not see any signal at higher degrees, even in the high-resolution extended mission data. Given the crustal thickness of ~50 km in this region, this simple relationship indicates that the sources of these gravity anomalies are embedded within the crust. This will be tested more quantitatively in the inversions below. In contrast, the upper bound of the range in degrees over which the PKT border anomalies are expressed would indicate source depths of 30-45 km, which is well below the base of the crust in those areas. However, the fact that the longer wavelength PKT border anomalies are associated with mare at the surface and are found in areas of thinner crust argues against such a deep source, and Bowin’s relationship cannot rule out smoother source bodies confined to shallower depths. If the gravity anomalies arose solely from density anomalies at the base of the crust at ~30 km depth (Wieczorek et al., 2013), then gravity anomalies at degree 60 would be attenuated by 50% relative to anomalies at degree 20, and thus the source depth cannot explain the lack of signal beyond degree 60. For the PKT border anomalies, the lack of high degree structure is more likely due to the smooth nature of the source bodies, consistent with the previous models of thickened mare and thinned crust along the border anomalies (Andrews-Hanna et al., 2014).
Figure 2.3. Bouguer gravity gradient maps of the Moon with low-pass filters applied at degrees 10 to 120. The nearside hemisphere (right half of each map) and farside hemisphere (left half) were processed with Bouguer correction densities of 3100 kg/m$^3$ and 2500 kg/m$^3$, representative of the mare and highlands, respectively. Anomalies of interest are highlighted in the “Degree 10” panel.
Figure 2.4. Bouguer gravity gradient maps of the Moon with high-pass filters applied at degrees 10 to 120. Anomalies of interest are highlighted in the “Degree 10” panel.

2.4.2 Power Spectra

The band-passed gravity gradient results discussed in the previous section indicate that the power spectra of the PKT border anomalies may be expected to rise above the background between degrees 40 and 60 for the localization bandwidth of 40, since lower degrees are affected by the localization. Within that range, the NSE spectrum shows the most deviation from the background spectrum, followed by the NNE spectrum (Fig. 2.5b,c). The NSE spectrum differs from the background spectrum by about 1 order of magnitude at degree 40 before decreasing to 0.7 orders at degree 60. However, the edges of the Serenitatis and Nectaris basins may contaminate the power spectrum of this anomaly, and the band-passed maps show that the majority of power from the NSE anomaly is below degree 40. The maximum power spectrum excess relative to the background for the NNE anomaly is 0.7 orders of magnitude. However, the excess in the low degree power spectra is comparable to the variability in the power spectra across all degrees for both of these anomalies. In contrast, the NS, and NNW (Fig. 2.5a,d) show minimal power excess at low degrees, with a greater excess at high degrees (>70) at which they are not observed in the bandpassed gravity maps. This comparison is complicated by the variability in the Bouguer gravity data over the maria arising at least in part from buried craters and impact basins (Neumann et al., 2015; Evans et al., 2016). The lack of a consistent excess in power at low degrees in comparison with high degrees renders spectral approaches to probing the
nature and origin of the PKT border anomalies ineffective, while spatial approaches (e.g., Andrews-Hanna et al., 2014) can be more useful.
**Figure 2.5.** Bouguer gravity power spectra calculated for the regions of interests on the nearside (a-c) and farside (d-f). The error is marked as the shading around the plotted spectra.

In contrast to the nearside PKT border anomalies, the power spectrum of the FN LGA anomaly is consistently greater than that of the background over the degree range indicated by the bandpass maps (Fig. 2.5e). The power of the FN anomaly is twice that of the background at degree 80, and decreases gradually until equaling the power of the background at degree 250. There is also a slight change in slope of the spectrum at around degree 140. This observation is consistent with the high-pass filtered gravity gradient maps, in which the FN anomaly becomes indistinct from the background beyond degree 250.

In contrast to the FN anomaly, the FS anomaly along the rim of the South Pole-Aitken basin has a power spectrum indistinguishable from the background area on the SPA rim (Fig. 2.5f). The spectral behavior can be explained by the overall greater magnitude and higher variability of the gravity anomalies at all degrees along the SPA rim, including other linear gravity anomalies both radial and circumferential to the basin. Thus, the abundance of large magnitude gravity anomalies across all degrees along the rim of the SPA basin precludes the use of localized power spectra to study the nature of this LGA.

In summary, the PKT border anomalies are expressed only in the low degrees (20-60), and have power spectra that do not clearly rise above the background in the expected degree ranges due to both the variability in the field and the limitations of the bandwidth localization. In contrast, the LGAs are expressed primarily in the high degree gravity data (40-250) in the band-passed maps, though only the power spectrum of the FN anomaly on the northern farside exceeds
the background this range of degrees. While the PKT border anomalies and the FS anomaly are clearly prominent features in map view, their power spectra do not rise above the background in the localized analysis. This lack of distinction is a result of gravity anomalies of comparable scale and magnitude surrounding the features. This result shows the importance of considering both spatial and spectral approaches to gravity analysis.

2.4.3 MCMC Results

Based on the distinctive power spectrum of the FN anomaly, we next used the Markov chain Monte Carlo (MCMC) model to constrain the dimensions of the source of the gravity anomaly. The increase in power for the FN anomaly relative to the background was observed from roughly degree 80 to 250 (Fig. 2.5e). However, the power spectrum from degree 80 to 92 was omitted from the calculations of model misfit as its spectral slope is a sharp departure in behavior from the rest of the spectrum and the background power spectrum also exhibited both a greater slope and considerable variability over this degree range. The exclusion of degrees <92 in the inversion will result in some loss of the longer wavelength component of the anomaly, and may result in an underestimate of the width and/or bottom depth. For the three chosen geometries (elliptical, triangular, and T-shaped cross-sections), the model was used to explore the depth of the top, the height, and width of the intrusion (and height of the top prism and width of the bottom prism for the T-shaped anomaly). The Monte Carlo model was used to identify the best-fit geometry (based on the lowest RMS misfit), and the average and one standard deviation (1-σ) geometry (based on the a posteriori distribution of model parameters). Results are presented below in terms of the best-fit solution (mean ± standard deviation). However, the
aspect ratio fits were asymmetrically skewed due to the large variations in height and width, and thus the results are shown in terms of the median and asymmetric $\pm 1\sigma$ range.

The results (Table 2.1) show a consistent best-fit top depth of 9-13 km across all three shapes (ellipse, triangle, T). The mean top-depth follows a similar distribution, with a common range of 15-19 km and a standard deviation of $\pm 4-6$ km. The corresponding 95% confidence interval is 6-31 km, which is an improvement over the previous spatial approach that constrained the top depth to a wider range of 8 to 41 km (95% confidence interval for a density contrast of 550 kg/m$^3$; Andrews-Hanna et al., 2013). The increased constraint in the confidence interval compared with the previous spatial domain analysis is an expected result of the spectral method used here, which gives greater sensitivity to the top depth.

On the other hand, the fits for width and height were less well constrained, with a strong tradeoff between the two. The elliptical prism runs returned a height of 37.4 km (18.1±8.4 km, mean ± standard deviation) and width of 11.3 km (25.1±10.0 km). The triangular prism, designed to emulate a structure such as the Great Dyke of Zimbabwe, has large uncertainties in its height and width fits as well, with a height of 35.0 km (17.7±8.0 km) and a width 12.2 km (28.1±13.5 km). Note that the best-fit values may sometimes lie outside of the $1\sigma$ range about the mean, depending on the shape of the \textit{a posteriori} distribution in the parameter space. The T-shape model was intended to test a hybrid structure in which a sill overlies a dike, but also allows for end-member solutions such as fully dike-like or sill-like structures. In multiple restarts, the models converged on a simple geometry in which the top and bottom prisms had comparable widths of 7.0 km (21.0±11.2 km) and 6.2 km (19.0±9.7 km), respectively, and a true T-shape
solution was not found. The combined heights for the T-shape model were similar to those of the other models.

Table 2.1: MCMC Results for modeling the FN anomaly (all values in units of km).

<table>
<thead>
<tr>
<th>Model</th>
<th>Top Depth$^a$</th>
<th>Height</th>
<th>Width</th>
<th>Bottom Height</th>
<th>Bottom Width</th>
<th>Aspect Ratio$^{b,c}$</th>
<th>RMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Elliptical</td>
<td>9.11(14.7±3.53)</td>
<td>37.4(18.1±8.36)</td>
<td>11.3(25.1±9.95)</td>
<td>N/A$^d$</td>
<td>N/A</td>
<td>3.31(0.72±0.33)</td>
<td>0.100</td>
</tr>
<tr>
<td>Triangular</td>
<td>13.4(15.6±4.73)</td>
<td>35.0(17.7±7.98)</td>
<td>12.2(28.1±13.5)</td>
<td>N/A</td>
<td>N/A</td>
<td>2.87(0.65±0.40)</td>
<td>0.106</td>
</tr>
<tr>
<td>T-shape</td>
<td>10.2(18.7±5.63)</td>
<td>1.16(12.3±8.40)</td>
<td>7.01(21.0±11.2)</td>
<td>38.7(20.2±9.62)</td>
<td>6.20(19.0±9.71)</td>
<td>5.69(1.65±1.62)</td>
<td>0.099</td>
</tr>
</tbody>
</table>

$^a$ Top depth, height, and width results presented as best-fit (mean ±1-$\sigma$ range)

$^b$ Aspect ratio results presented as best-fit (median±1-$\sigma$ ) due to strongly asymmetric posterior distribution

$^c$ The T-shape aspect ratios are calculated using the widths and heights of the bottom structure only.

$^d$ N/A - not applicable

Generally, the solutions for the three prism types (elliptical, triangular, T) returned results of similar RMS misfit for the best-fit models (~0.10, in normalized power units). The RMS misfits are around 4 times lower than the background variability (see methods section). The T-shape geometry has the lowest misfit (0.099), followed by the elliptical cross-section (0.100), and then the triangular cross-section (0.106). The T-shape model has a larger number of free parameters and thus was expected to provide a better fit to the multiple slopes within the power spectrum, though the difference in fit across the three geometries is not significant.
Across all geometries tested, the best-fit models were elongated in the vertical dimension (height/width aspect ratios of 2.9–5.9), but the distribution of acceptable solutions included sill-like source bodies elongated in the cross-strike horizontal direction. The similarity in fit of the three geometries is clearly seen in the comparison of the synthetic and observed power spectra (Fig. 2.6a). The predicted gravity profiles for all models provide reasonable but imperfect fits to the observed spatial profile, which is expected since the model fitting was based on the power spectrum rather than the spatial profile (Fig. 2.6b). Also, as mentioned above, some of the amplitude of the anomaly is lost because only degrees >92 in the power spectrum were used in the inversion.

**Figure 2.6.** a Bouguer gravity power spectra comparison between the observed FN anomaly and the best-fit models. b Bouguer gravity profile comparison between the observed FN anomaly and the best-fit models.
The spatial domain inversions for the FN anomaly yielded similar results to the spectral domain inversion (Table 2.2), and fit the observed profile well (Fig. 2.7). The top depth of 17.6 km (20.4±4.9 km) was deeper than the best-fit value from the spectral approach, while the 1-σ range was somewhat broader than but in agreement with the previous result. We consider the shallower top depth from the spectral approach to be more reliable, as the spatial approach relies on an average profile that has likely experienced some smoothing. The height of 21.7 km (17.5±7.6 km) and width of 23.5 km (36.5±17.4 km) were also consistent with the spectral domain models. Both sets of models are in agreement with the previous analysis based on the GRAIL primary mission data, with a best-fit height of 15.1 km and width of 29.1 km. The resulting best-fit top depths from the spatial inversions of the remaining anomalies (Table 2.2) were in the range of 21.5-28.1 km, and the best-fit widths and heights were in the range of 18.3-39.7 km, and 20.6-28.3 km, respectively. This corresponding best-fit aspect ratios (height/width) range from 0.6-1.6, including both vertically and horizontally elongated cross-sections. The 1-σ range in widths and heights allow both dike-like and sill-like solutions for each of the anomalies, though all solutions are much wider and more equant in cross-section than typical dikes. The range in cross-sectional area (volume per unit length) for LGAs FS, 3, and 4 of 406–732 km² is similar to the FN cross-section of 403 km². The combined modeled volume of the four anomalies is $1.76 \times 10^6$ km³.
Figure 2.7. Bouguer gravity profile comparison between the real LGAs and the modeled elliptical prisms.
Table 2.2: Spatial-domain MCMC results for elliptical cross-section models of the four LGA anomalies.

<table>
<thead>
<tr>
<th>Anomaly</th>
<th>Top Depth</th>
<th>Height</th>
<th>Width</th>
<th>Aspect Ratio</th>
<th>RMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>FN</td>
<td>17.6 (20.4±4.94)</td>
<td>21.7 (17.5±7.61)</td>
<td>23.5 (36.5±17.4)</td>
<td>0.923 (0.86±0.075)</td>
<td>0.345</td>
</tr>
<tr>
<td>FS</td>
<td>26.1 (26.0±4.61)</td>
<td>23.4 (16.9±5.28)</td>
<td>39.7 (57.1±17.9)</td>
<td>0.591 (0.43±0.24)</td>
<td>0.233</td>
</tr>
<tr>
<td>LGA3</td>
<td>21.5 (26.9±5.58)</td>
<td>28.3 (14.4±6.15)</td>
<td>18.3 (42.1±18.5)</td>
<td>1.55 (0.58±0.24)</td>
<td>0.322</td>
</tr>
<tr>
<td>LGA4</td>
<td>28.1 (26.8±5.05)</td>
<td>20.6 (13.6±4.93)</td>
<td>32.2 (51.8±17.4)</td>
<td>0.640 (0.44±0.24)</td>
<td>0.349</td>
</tr>
</tbody>
</table>

2.5 Discussion

2.5.1 Comparison of the PKT border anomalies and the LGAs

Both the band-passed gravity gradient maps and the localized power spectra confirm that the PKT border anomalies are dominantly long-wavelength low-degree (L~20-60) features of the gravity field, in contrast to the LGAs which show continued signal out to degree 120-250. The PKT border anomalies are clearly wider in map view than the LGAs (Andrews-Hanna et al., 2013; 2014), but these results go further by showing that the PKT border anomalies lack structure at the smaller scales of the LGAs. A relative lack of high degree structure in the gravity could be a result of either a larger depth of origin or a physical structure that is inherently smooth. Both the association of the PKT border anomalies with mare volcanism at the surface and the thin crust in the region argue against a deeper source for the anomalies. Thus, the PKT border anomaly source bodies are fundamentally smoother and longer wavelength structures than
the sources of the LGAs. This inference is consistent with previous models of the PKT border anomalies as arising from broad zones of crustal thinning beneath the maria associated with some manner of extension or rifting prior to the volcanism. In contrast, the LGAs lack a surface signature but must arise from sources at depths much shallower than the crust-mantle interface, and therefore the source bodies must be embedded within the crust.

### 2.5.2 Geometry of the underlying intrusive bodies and possible analogs

For the LGAs, the dike-like elliptical prisms and the triangular prisms for our MCMC models were both able to fit the power spectra of the FN anomaly to comparable misfit values. The modeled best-fit widths for the dike-like structures are roughly 10–40 km, which are much larger than the meter to tens of meter scale widths of typical dikes and dike swarms (Wilson, 1990; Ernst et al., 2001). One of the goals of the T-shape modeling was to attempt to fit the gravity data with a more typical width dike in combination with a sill, rather than a single very wide intrusion. However, such a solution was not found within the allowed parameter space. Thus, our analyses indicate that a dike-sill hybrid structure does not provide a solution that matches the gravity for a more typical width dike. The failure of the T-shaped models has compelled us to evaluate possible analog features of similar width scales elsewhere in the Solar System in order to constrain the origin of the FN anomaly and other LGAs.

As discussed above, the Great Dyke of Zimbabwe, located in southeastern Africa, is a well characterized Earth analog of similar length and width to the modeled FN anomaly. The triangular shape of the Great Dyke (Wilson, 1996) can match the observed gravity anomalies on
the Moon as well as the elliptical models to within the observed variability in the data. The width of the Great Dyke from surface exposures (4–11 km) and gravity inversions (5–8 km) fall well below the 1-\(\sigma\) ranges for the FN LGA from our inversions (15.6–41.6 km). Similarly, the height of the Great Dyke from surface observations and drill cores (1.9–3.35 km) and gravity models (1–4 km) are less than our inversions of the FN LGA (9.7–25.7 km, respectively). The gravitational expression of the Great Dyke is much narrower than that of the LGAs (full-widths at half-maximum of 5 km and 50 km, respectively), reflecting the much greater depth extent of the latter and resultant widening of the gravity anomaly. The aspect ratio of the Great Dyke of approximately 0.1–1 is consistent with the more equant aspect ratios within the 1-\(\sigma\) range from our triangular models of 0.3–1.7 though the best-fit models prefer a taller and narrower structure.

Models based on gravity traverses support the presence of an underlying feeder dike and/or deep-seated magma chamber extending as deep as 12 km beneath some sections of the Great Dyke (Wilson, 1996). The Great Dyke is presently exposed at the surface but is interpreted as an intrusive laccolith. The depth of the Great Dyke at the time of its formation is not known.

Thermochronometer data and thermal history models support uplift and denudation of the Zimbabwe craton during two events in the Phanerozoic by \(\sim 2.5–10.6\) km and \(0.5–2.5\) km (Mackintosh et al., 2017), suggesting that initial top depths of the intrusion of up to \(\sim 3–13\) km are possible. Our modeled top depths of \(10.9–20.3\) km (1-\(\sigma\) range) are consistent with the exposure of the Great Dyke by this erosion. Given the many differences in the settings of the Great Dyke and the LGAs and the uncertainty in the inversion results, the differences in geometry do not exclude it as a possible analog, with the FN LGA being a factor of \(\sim 2–10\) larger, extending to greater depths, and with a higher aspect ratio preferred.
Both horizontal (pure shear, wrench tectonics) and vertical (lithospheric flexure) tectonics models have been proposed to explain the formation of the Great Dyke (Wilson & Prendergast, 1989), though some form of crustal extension must have been present during the intrusion (Wilson et al., 1987). Wilson (1996) hypothesizes that the Great Dyke formed from first the collision of the Zimbabwe and Kaapvaal Cratons and the overthrusting of the local province, and then the crustal extension as a result of subsequent horizontal rotations in the local maximum stress field. While models based on plate tectonics are not applicable for the Moon, a global crustal extensional regime may have been possible on the early Moon (see discussion below). However, strain rates from such global processes on the Moon are likely to be much lower than the plate tectonic strain rates associated with the formation of the Great Dyke, further limiting its usefulness as an analog.

The elliptical models provide equally good fits to the data for a geometry consistent with a single dike-like intrusion. Possible analogs to the FN anomaly that may be similar in geometry are the inferred intrusions beneath the collapse pits near Valles Marineris on Mars (Mège et al., 2003). The observed collapse pits are inferred to be the result of under-pressurization of the magma in elongated dike-like intrusions, causing collapse of the overlying crustal material. This process likely occurred in an extensional regime as well, during the formation of Valles Marineris. The lower-bound dimensions of the precursor dike-like structures can be inferred from the volumes of the collapse pits. Assuming bodies that are 25-75 km in height, the minimum widths would be 0.4-1.2 km (Andrews-Hanna, 2012a). Since the under-pressurization did not likely entail complete drainage of the magma from the intrusions, much greater widths are likely required, perhaps similar to the dimensions of the modeled lunar structures. The top
depths of the martian intrusions are poorly constrained, though, given their relationship to the Valles Marineris canyons, top depths beneath the ~8 km-thick accumulation of lava flows making up this part of Tharsis (McEwen et al., 1999) seem likely. No collapse pits, nor any visible structures associated with the anomaly, have been observed in the terrain above the FN anomaly. However, the lack of collapse pits could be explained by a lack of magma underpressurization, while a lack of graben (e.g., Schultz et al., 2004) may be a result of the large top depths of the intrusions or subsequent modification of the surface by impact cratering. The martian intrusions around Valles Marineris are not an ideal geodynamic analog to the lunar intrusions, as the former are likely related in some way to stresses from volcanic loading (Andrews-Hanna, 2012b), while the lunar LGAs are not clearly associated with any known surface or subsurface loads. However, similar gravity anomalies inscribing circular arcs around lunar basins interpreted as ring dikes intruded into the basin rings (Andrews-Hanna, 2018) are associated with flexural stresses from the central mascons. All things considered, the intrusions underlying the martian collapse pits, while poorly understood themselves, are presently the best analog to the sources of the LGAs.

Linear chains of collapse pits and elongated troughs on Mars have been compared to terrestrial volcanic rifting at the East Pacific Rise and at Iceland (Mège et al., 2003). The East Pacific Rise is underlain by a series of elongated magma reservoirs or sills up to 4 km wide and <100 m thick (Singh et al., 2006). The upper magma reservoir is underlain by an additional sill or partially molten zone (Dunn et al., 2000; Marjanović et al., 2014) which, if present on the Moon, would also contribute to a positive gravity anomaly and may help to reconcile this analog with the scale of the density anomalies beneath the LGAs. However, axial magma chambers
such as this are typically associated with fast-spreading ridges, and thus this analog would require a surprisingly high rate of extension which is difficult to reconcile with proposed models to explain the lunar extension (see below). Elongated magma chambers are also inferred beneath dike swarms in Iceland along the crust-mantle boundary at 8-12 km depth (Gudmundsson, 1995), and would contribute to the gravity anomaly from the dikes themselves (see below).

Alternatively, the LGAs may be interpreted as representing swarms of dikes of more typical widths of 1 to >10 m (Ernst et al., 2001). The lack of distinct structure at higher degrees (L > 120-250) does not support the presence of a small number of larger intrusions and would require instead a large number of closely and uniformly spaced narrower dikes. Previous models in which density was kept as a free parameter (Andrews-Hanna et al., 2013) can be used to test the scenario of numerous dikes below the resolution of the data, as the individual dikes would not be separately resolved. For the 1-σ lower bound on the density contrast of 90 kg/m$^3$ from that study and an assumed true density contrast of 550 kg/m$^3$, 10-m-wide dikes would require a spacing of <60 m over a zone ~10 km wide, making up >15% of the crust. By contrast, some of the most closely spaced dike swarms on Mars and Venus inferred from the observed surface fractures and graben have typical spacings of ~5 km (e.g., in Tantalus Fossae on Mars and a graben swarm centered at 16°S 352°E on Venus; Ernst et al., 2001). However, a rift-parallel swarm of dikes associated with a volcanic center in Iceland has a cumulative width of 1.1 km over a section 12.8 km wide, with dikes comprising 8.6% of the crust (Klausen, 2006). In other similar dike swarms in Iceland, dikes typically make up 1-5% of the volume of the crust, but can reach values as high as 15-28% (Gudmundsson, 1995). Thus, we cannot rule out the possibility that the LGAs are the manifestations of closely spaced swarms of narrow dikes, possibly
underlain by a deeper magma chamber. As with other analogs, such tightly clustered dike swarms are associated with crustal extension and rifting.

2.5.3 Age and geological relationships

There are no visible crustal effects such as faults or collapse pits above the FN anomaly to constrain its timing through superposition relationships, though there may be a subtle surface depression associated with it (Sawada et al., 2016). Instead, the timing of the anomaly’s formation may be constrained by looking into how the anomaly interacts with two superimposed craters. The FN anomaly stretches between the 62-km-diameter Upper Imbrian crater Schjellerup in the east and the 160-km-diameter Nectarian crater Rowland in the west (labeled as A and B, respectively, in Fig. 2.8a) and crosses a range of Nectarian and pre-Nectarian surfaces between these two craters. In the gravity gradient map (Fig. 2.8a), the FN anomaly becomes more diffuse near the western end, coinciding with the location of Schjellerup crater as shown in Fig. 2.8b. At the other end, the eastern tail of the FN anomaly ends abruptly at Rowland crater, but is continuous with a similar anomaly that curves around the crater without any decrease in gravity gradient signal. Negative gravity gradient rings interpreted as dikes intruded into basin ring faults are common around multiring impact basins elsewhere on the Moon such as around Orientale (Andrews-Hanna et al., 2018), though no tectonic ring is expected around a crater of this size. Negative and positive gravity gradient rings are also often associated with crater rims as a result of local densities that differ from that used in the Bouguer correction, and this provides the most likely explanation for the anomaly at the rim of Rowland. The simplest interpretation is that the FN gravity anomaly is weakened by the smaller Schjellerup crater and erased by the larger Rowland crater as a result of the impact modification and excavation of the underlying
intrusion. The predicted excavation depths for Rowland and Schjellerup are ~16 km and ~6 km, respectively. Inward and upward flow of crustal material below the excavation cavity during the collapse of the transient cavity would disrupt the intrusions to greater depths (O’Keefe & Ahrens, 1999). Beneath the excavation cavity, the shockwaves generated by the formation of the craters would have crushed out the porosity in the porous megaregolith, while generating porosity within a low porosity intrusion (Collins, 2014; Milbury et al., 2015), thereby reducing both the density contrast between the intrusion and the surrounding crust and the resulting gravity anomaly. GRAIL gravity data shows a robust trend of gravity anomaly with crater diameter across all diameters (Soderblom et al., 2015) consistent with substantial impact-generated porosity. Thus, the predicted depth extent of the intrusions, ranging from a top depth of 14.7±3.5 km down to the base of the crust, is consistent with a modest reduction in the density contrast from impact generation and crushing of porosity beneath Schjellerup, and the excavation and disruption of the intrusion to greater depths beneath Rowland. Based on this inference, we infer that the FN anomaly must be older than the Nectarian-aged Rowland crater.
Figure 2.8. Zoomed-in gravity gradient map a and LRO mosaic b in Mercator projection of the FN anomaly with the anomaly outlined in white, and the Schjellerup and Rowland craters labeled as A, B.

2.5.4 Implications for lunar volcanism

Neutral buoyancy of magma relative to the surrounding crustal and mantle material is thought to be a major factor in the vertical extent and eruptibility of dikes and intrusions (e.g., Wieczorek et al. 2001). For an intrusion ascending from the mantle into the crust, the magma would be positively buoyant in the mantle and negatively buoyant in the crust. The best-fit density structure for the farside highlands outside of SPA has a surface density of 2308 kg/m$^3$, surface porosity of 20.5%, and a porosity e-folding depth scale of 28.4 km (Besserer et al., 2014). For a crustal thickness of 50 km, a grain density of 2904 kg/m$^3$, a mantle density of 3220 kg/m$^3$ (Wieczorek et al., 2013) and a magma density of 2900-3010 kg/m$^3$ (Wilson and Head, 2017), a neutrally buoyant intrusion with a top depth of 15 km would extend down into the mantle to a depth of 22-52 km below the base of the crust (ignoring the likely possibility of vertical variations in the stress state). Given the smaller density contrast between the magma and the mantle and the attenuation of gravity anomalies with depth of the source body, the observed gravity anomaly would be dominated by the crustal portion of the intrusion. In this scenario, the vertical extent of the intrusion below the crust may exceed that within the crust, and the aspect ratio of the intrusions would be a factor of ~2 greater than predicted by the gravity anomaly.
Furthermore, the volumes of the associated intrusions would be a factor of 2 or more greater than inferred from the gravity data.

An intrusion sourced from the same depth in the mantle would erupt for crust thinner than 30–32 km for the above magma densities. These crustal thicknesses are generally limited to the nearside mare region and the floors of impact basins. Thus, similar and contemporaneous magmatic activity to that responsible for the FN anomaly would have been eruptive on the nearside, and may have been responsible for the earliest phases of mare volcanism and the volcanic resurfacing of the oldest basin floors. While the oldest crater retention ages for the exposed mare surfaces are typically <4.0 Ga, with one surface in Mare Tranquilitatis dated to 4.2 Ga, both mare units below the surface and cryptomare in the highlands may be older (Hiesinger et al., 2011). Fragments of mare basalt in the Apollo samples provide evidence for mare volcanism as early as 4.2-4.3 Ga (Taylor et al., 1983; Dasch et al., 1987). We thus interpret that the FN anomaly intrusion may have been the farside equivalent of the earliest phases of nearside mare flooding, but which failed to reach the surface.

The total magma volume within the FN intrusion using the simplest elliptical shape model from the spectral inversion is $1.7 \times 10^5$ km$^3$, while the calculated volume for all four LGAs in the spatial inversion is $7.8 \times 10^5$ km$^3$. Allowing for an equal volume below the base of the crust would double this number. If we assume that the cross-sectional area of these anomalies of ~400 km$^2$ is typical, and scale by the total combined length of all probable LGAs identified by Andrews-Hanna et al. (2013), we arrive at a volume of $2.1 \times 10^6$ km$^3$. This intrusive volume is a factor of five lower than the estimated volume of the nearside maria of $\sim 1 \times 10^7$ km$^3$ (Head 1975;
Evans et al., 2016; Gong et al., 2016; Head & Wilson, 1992). This difference is, to first order, consistent with previous arguments that the overall farside magmatic production is a factor of \(~10\) lower than that of the nearside based on both the observed abundance of maria within SPA (Wieczorek et al. 2001) and on the predictions of thermochemical evolution models (Laneuville et al., 2013).

2.5.5 Implications for lunar geodynamics

The original observation of evidence for a global set of randomly oriented vertical tabular intrusions was interpreted as consistent with some thermal evolution models that predict an early epoch of global expansion as a result of the thermal equilibration of a shallow-magma-ocean Moon (Andrews-Hanna et al., 2013; Solomon, 1977; Solomon and Chaiken, 1976). However, a shallow magma ocean is difficult to reconcile with the expected thermal state of a rapidly accreted Moon of giant impact origin (Pritchard and Stevenson, 2000), particularly for recent high angular momentum lunar formation models involving symmetric giant impacts (Canup, 2012) or impacts into a fast-spinning proto-Earth (Cuk and Stewart, 2012), or for lunar formation from a super-critical disk (Lock et al., 2018). Rapid accretion of the Moon (Salmon and Canup, 2014, 2012) would likely lead to a hot initial state, though early accretion of cooler Roche-interior material (Canup et al., 2015) could result in a somewhat cooler and more hydrated deep interior of the Moon. While a shallow magma ocean is difficult to reconcile with current lunar formation models, geophysical evidence for the base of the magma ocean remains equivocal (see review in Andrews-Hanna et al., New Views of the Moon 2, in press).
An early period of expansion could also be an outcome of volume changes during magma ocean crystallization (Elkins-Tanton & Bercovici, 2014). Such an early episode of expansion may not lead to dike propagation in the conventional sense, given the likely hot state and weak rheology of the crust at that time. On the other hand, the reduced viscosity contrast between the warm early crust and the magma may be more conducive to the formation of large-scale dike-like intrusive diapirs (Rubin, 1993). However, at least some of the farside linear gravity anomalies post-date the South Pole-Aitken basin, with a possible formation age of 4.25 Ga (Garrick-Bethell et al., 2020), after solidification of the magma ocean was complete (Elkins-Tanton et al., 2011). Later overturn of the buoyantly unstable magma ocean cumulates would lead to partial melting of the rising orthopyroxene and olivine cumulates from the lower mantle, causing both expansion and magmatism (Elkins-Tanton & Bercovici, 2014). This overturn may have occurred after the SPA-forming impact (Melosh et al., 2017).

Later concentration of KREEP-rich material on the lunar nearside (e.g., Parmentier et al., 2002), possibly triggered by the SPA-forming impact (Zhang et al., 2022; Jones et al., 2000), would have a dominant effect on the subsequent thermal evolution (Grimm, 2013; Laneuville et al., 2013; Wieczorek and Phillips, 2000). The enhanced heat flow and convection within the PKT can cause net expansion of the nearside, but the farside in this scenario would always be contracting (Laneuville et al., 2013). Given the abundance of linear gravity anomalies on the farside, some other source of extensional stress in the crust is needed. Subsequent sinking of KREEP-rich ilmenite-bearing cumulates (Zhang et al., 2013a, 2013b) may have concentrated heat producing elements above the core, which would have caused an early period of deep
warming and expansion. Similar models have previously been invoked to explain an early strong lunar dynamo (Stegman et al., 2003).

Thus, while our results are consistent with the previous analysis in interpreting the linear gravity anomalies as elongated vertical tabular intrusive bodies of some sort that likely formed in a global horizontally extensional stress state, we cannot at this time connect this stress state to a single geodynamic process. However, proposed mechanisms for generating global extension involve slow processes and low strain rates. In contrast, the analog structures on Earth (Great Dyke of Zimbabwe, axial magma chamber along the East Pacific Rise, dike swarms in Iceland) are associated with high strain rates driven by plate tectonics, while the one planetary analog (chains of collapse pits and troughs along Valles Marineris) remains poorly understood but is associated with flexural strain.

2.6 Summary and conclusions

In this study we used band-passed gravity gradient maps and localized power spectra analyses to study the nature of the PKT border anomalies, which were hypothesized to be volcanically flooded rift valleys (Andrews-Hanna et al., 2014), and the Moon-wide linear gravity anomalies, which were interpreted as subsurface giant dike-like intrusions (Andrews-Hanna et al., 2013). The band-passed maps revealed that while the PKT border anomalies are primarily long wavelength features whose spectral power was concentrated between degrees 10 and 60, the linear gravity anomalies are short wavelength features whose spectral power ranged from degree 50 to 250. Power spectra analyses revealed that the spectra of the PKT border anomalies do not
rise significantly above that of the background area, while the power spectrum of the FN LGA does. The clear contrast between the spectral characteristics of the two types of anomalies implies significant and fundamental differences in the origin and nature of the anomalies, as differences in depth cannot reconcile the difference in power. The high degree component of the LGA signal indicates that the source bodies reach depths as shallow as \(~7–15\) km, while the longer wavelengths characterizing the PKT border anomalies are more likely attributable to the smoothness of the source bodies rather than their depth,

We then used a Metropolis-Hastings MCMC algorithm to examine the nature of the density anomaly responsible for the FN LGA by modeling the power spectra of candidate subsurface structures and comparing them to the observed spectra. We investigated three possible shapes of the subsurface structure: an elliptical dike-like intrusion, a Great Dyke-like triangular prism, and a T-shape structure used to test a dike-sill hybrid. The MCMC models showed that all three geometries provided reasonable matches to the power spectrum and produced gravity profiles that were similar, though a true T-shaped solution with a more geologically reasonable dike width was not found. This power spectrum approach is more sensitive to the top depth of the anomalies than the previous spatial analyses, confirming the top depths of the intrusions of \(~10\) km. This top depth is supported by the weakening and obliteration of the FN gravity anomaly by two craters as a result of their excavation and disruption of the upper portions of the underlying intrusive bodies. This superposition relationship also supports an age for the FN intrusion of early Nectarian or pre-Nectarian.
Dikes are tectonic-magmatic structures that occur on the Earth, and are thought to be present on other major solid bodies as well. The widths (∼9–23 km) of the lunar LGAs are much larger than the widths of typical dikes (∼1–100 m). On the Earth, the Great Dyke of Zimbabwe, the axial magma chamber in the East Pacific Rise, and linear dike swarms underlain by a magma reservoir are analogs of comparable scale, though all formed in high-strain plate tectonic settings. The inferred intrusions beneath the large collapse pits around Valles Marineris on Mars are intrusive bodies of comparable scale and elongation to the sources of the LGAs, and are perhaps the most relevant analog. The presence of populations of giant linear intrusions with widths of 1–10 km on the Moon and Mars attests to the importance of a poorly understood form of magmatism.

While a well-understood analog to the LGA intrusive bodies is lacking, all possible analogs form in extensional environments, confirming the interpretation of widespread extensional strain recording an epoch of global expansion on the early Moon. However, the implications of that expansion remain unclear, with possible explanations including thermal equilibration after crystallization of a shallow magma ocean, interior warming due to the sinking of KREEP, expansion during the final stages of magma ocean crystallization, and expansion due to the overturn and partial melting of the magma ocean cumulates. Regardless of the explanation, this early period of expansion exerted a dominant control on the form of early intrusive activity, as well as the limited compressional tectonics found later in lunar history. The LGA intrusions also featured prominently in the volcanic-magmatic evolution of the Moon, with the estimated total volume of all intrusions adding up to approximately 20% of the volume of the nearside maria. Similar intrusions in the thinnest crust on the nearside would have been eruptive,
and may have contributed to the earliest phases of mare volcanism. While much remains to be done to elucidate the nature and cause of these ancient intrusions, they clearly feature prominently in the magmatic, tectonic, and geodynamic history of the Moon.

2.7 Acknowledgements

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2.8 Supplementary Material

2.8.1 Spherical harmonic localization window testing

To compare the spectral signal between the LGA and the PKT border anomalies and their respective background regions (Fig. 1), localized power spectral analysis (Wieczorek & Simons, 2005) was conducted to extract spectral information from designated regions on the Moon. The bandwidths used to generate the localized power spectra were chosen to maximize the spatial concentration and to minimize the loss of signal at the start and end of the spectral range. All background areas were localized using the same bandwidths and eigenvalue thresholds as the anomalies with which they are compared. However, due to the differences in the area and shape of the regions of interest, the number of tapers that satisfied the eigenvalue requirement ranged from two to four for the nearside areas, and four to twenty-two for the farside regions. As a
result, the difference in the localization window shapes as well as the number of tapers in the power spectral analyses of the anomalies compared to those of the background areas may produce a bias in the results. We test for the existence of the bias in the following text.

The presence of the bias was tested by comparing the independently-generated Bouguer gravity power spectra of the Moon between two sample windows, the FNBG (square-shaped) region and the FN (elongated) region (see Fig. 1 in main paper). Each spectrum in Fig.1 above is the localized power spectrum using the taper of a particular region and a synthetic randomized background gravity field. The synthetic fields were generated using the Bouguer gravity power spectrum of the entire Moon, from which random spherical harmonic coefficients were generated under the condition that they can recreate the original power. The resulting comparison (see Figure 2.S1) shows no systematic bias in power from using localization windows of different shapes, but do show the inherent variance associated with sampling a field using localized power spectral analysis.
Figure 2.S1. A composite of independently-generated synthetic Bouguer gravity power spectra using the FNBG or the FN regions of interest as the localization window. Each line of power is extracted from a unique synthetic gravity field which is generated from the Bouguer gravity power spectrum of the entire Moon. The synthetic gravity fields are unique in that while their spherical harmonic coefficients together will always return the original power it is generated from, the individual coefficients vary in sign and amplitude.
2.9 References


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Chapter 3: The missing craters and basin rings beneath the lunar maria

This chapter has been submitted to *JGR: Planets* and is currently in revision.

3.1 Abstract

Evidence for a population of craters buried beneath the nearside maria has been found in the gravity data returned from the lunar GRAIL mission. Although the total population of buried and visible craters within maria is comparable to the crater population in non-mare regions at large diameters, a deficit was observed for craters less than ~90 km in diameter. This deficit is surprising because the data can resolve craters down to 10 km in diameter. Similarly, the Imbrium basin only has a partially exposed ring system, with individual ring widths of up to ~100 km, but where those rings should be buried beneath the mare surface, we find the gravitational signature mostly non-existent. Consistent with both of these observations, gravity power spectra over the nearside maria show a marked deficit at high degrees ($l \geq 120$) relative to the power spectra of a simulated volcanically flooded cratered surface. In this study, we test a series of mechanisms and scenarios that may have resulted in the observed deficits in the buried crater populations, comparing our models to the data using localized power spectra and recovered crater size-frequency distributions from the modeled surfaces. Our results indicate that the observed crater deficit, power spectrum deficit, and missing rings of Imbrium are best explained by a smoothing of the pre-mare surface. We represent this smoothing as a diffusional process, which might occur as thermal erosion during the earliest stages of the mare eruptions. The removal of the missing craters and Imbrium rings was a massive and unprecedented event.
that sheds light on the early evolution of the mare region, possibly supporting high temperature voluminous floods of lava early during mare formation. These results may also have relevance to similar events elsewhere in the Solar System, such as the formation of the volcanic plains on Mercury, Venus, and Mars.

3.2 Introduction

The nearside lunar mare region represents the most volcanically active (Head, 1975; Hiesinger et al., 2011), and perhaps tectonically active (Andrews-Hanna et al., 2014), region on the Moon. Past studies have constrained the time history of emplacement of the topmost flows (Hiesinger et al., 2011, 2003, 2000), with a peak emplacement age of ~3.6 billion years ago. The average depth of the maria outside of the major impact basins is ~1 km (e.g., de Hon, 1979; Evans et al., 2016; Gong et al., 2016; Head, 1975), with these thick deposits presumably overprinting an older cratered surface resembling the farside highlands. However, the nature and history of the early mare flooding is unknown, as only the top-most surfaces representing the end stages of that flooding history are visible.

A difference in the surface crater size frequency distributions (CSFDs) has been observed between the farside highlands and the nearside mare, showing the comparatively youthful age of the mare surface (e.g., Neukum et al., 2001). GRAIL Bouguer gravity data led to the discovery of quasi-circular mass anomalies (QCMAs) in the nearside maria, which were hypothesized to be the signal from buried craters formed in the pre-mare surface (Evans et al., 2016; Sood et al., 2017). Adding the QCMA population to the exposed mare crater distribution results in a distribution that matches the non-mare CSFD at the large diameters. However, a dearth of buried craters less than
~90 km in diameter in the exposed and buried mare CSFD in comparison to the non-mare CSFD was observed (Evans et al., 2018), despite gravity data capable of resolving craters down to ~10 km in diameter. In contrast, observations of abundant buried craters in both topography data and their tectonic signatures on Mars and Mercury show that buried craters in this size range are preserved on those bodies (e.g., Byrne et al., 2016; Frey, 2006; Ostrach et al., 2015). As such, a dearth of the small, buried craters in the maria may be in some way related to the unique details of early mare emplacement and/or the early evolution of the mare region.

The rings of the Imbrium basin within the nearside mare pose a similar problem, in that the expected gravitational signatures of the ring segments buried beneath the maria are missing. Imbrium is a well-studied multi-ring basin (e.g., Head III, 1976; Spudis et al., 1988; Wilhelms and McCauley, 1971) in part due to its relatively young age (~3.87 Ga, Orgel et al., 2018) and large diameter (~1145 km). However, only portions of Imbrium’s rings are exposed in topography and gravity data. The remnants of the outer ring of Imbrium are the Montes Carpatus, Appeninus, and Caucasuses, which span 40, 58, and 23 degrees of arc around Imbrium, respectively, reaching maximum lengths of 600 km and heights of 6 km. The only unambiguous remnant of Imbrium’s middle ring is Montes Alpes, which spans 32 degrees of arc around Imbrium and is 280 km in length and 2.4 km in height. Remnants of the inner ring of Imbrium are exposed only as small isolated massifs such as Montes Rectii, Teneriffe, and Pico, which span up to 90 km in length and rise up to 1.8 km in height. The innermost ring of Imbrium is not exposed and is visible only as a ring of wrinkle ridges interpreted as overlying thrust faults following the edge of an uplifted mantle plug (Byrne et al., 2015). The remaining ring segments,
comprising the majority of Imbrium’s ring system, are presumably buried beneath the mare surface and experienced a similar mare flooding history as that of the pre-mare craters.

However, as we show below, there is no gravitational signature from the rings expected to be buried beneath the mare surface. Notably, the widths and heights of the Imbrium rings are comparable to the diameters and depths of the largest missing craters. The absence of the buried rings in the gravity data thus may be related to the deficits in small, buried craters and perhaps to the early evolution of the lunar mare region. The apparent erasure of features of this magnitude from the gravity field is very surprising.

In this study, we compare the nearside mare region to models of a volcanically flooded cratered surface from the lunar farside. We quantify the deficit of buried craters in both the spatial domain using crater size-frequency distributions (CSFDs) and in the spectral domain using Bouguer gravity power spectra, and investigate different candidate scenarios that may have produced the crater deficit. First, we test whether the attenuation of the gravity signal due to the thickness of the mare layer above the buried craters can explain the deficit. Next, we test whether a lower density contrast due to a higher density substrate can explain the deficits. Then, we consider a scenario in which the lunar surface at the time of cratering was already covered by a dense layer such as mare from much earlier eruptions, such that only large impacts penetrate into the lower density crust beneath. Finally, we test whether a simple diffusive process, representing physical processes such as impact erosion or thermal erosion, can produce the deficit. We also consider whether these mechanisms can explain the missing rings of Imbrium, and apply the
models to the topography of Orientale to compare the resulting gravity maps to the area around Imbrium.

3.3 Data and Methods

3.3.1 Simulation of a volcanically flooded surface

In order to model and test the effects of the proposed scenarios on the gravity of the nearside mare, an analog surface to the pre-mare nearside surface is needed. The current farside highlands region was selected as the analog surface, as the total population of exposed and buried craters with diameters >100 km in the maria approximately match the farside CSFD (Evans et al., 2018), and the majority of craters with diameters of 20-100 km formed in the Pre-Nectarian and Nectarian eras (e.g., Yang et al., 2020), before the final nearside mare flooding at ~3.6 Ga. While the same population of impactors in terms of size distribution is likely to have impacted the nearside mare surface and farside surface, it is not known whether the pre-mare basement has the same age as the farside comparison region, which may result in the deficiency of certain impact populations in a region’s CSFD. However, variations in the small crater (<~50 km) density within the farside surface are much lower (Head et al., 2010) than the difference between the buried small-crater population in the maria and the farside small crater population, which implies a roughly uniform age in which the global crustal basement formed. We focus our comparisons using a region between 152°E and 218°E and 5.52°S and 46.6°N (Fig 3.1) as the analog pre-mare surface.
Next, we simulate a lava-flooding event on the analog pre-mare surface, the current farside highlands, to represent the expected gravity for the nearside maria, and serve as the basis of comparison between the observed nearside gravity and modeled farside gravity for our different modification scenarios. Analyses were performed on the spherical harmonic degree 1200 GRAIL GRGM1200A free-air gravity model (Lemoine et al., 2014; Zuber et al., 2013) and a spherical harmonic degree 2600 LOLA topography model (Smith et al., 2010; Wieczorek, 2015). To simulate the flooding event (Fig. 2a), we used free-air gravity data processed using a Bouguer correction initially at the density of the lunar feldspathic crust at ~2500 kg/m$^3$ (Goossens et al., 2020; Wieczorek et al., 2013), and then the gravity from the relief along the crust-mare interface was added assuming a density contrast of -600 kg/m$^3$ (based on a mare bulk density of 3100 kg/m$^3$; Kiefer et al., 2012). This combination is identical to simply assuming a density of 3100 kg/m$^3$ when calculating the Bouguer gravity for the farside, although we also consider other density assumptions in testing mechanisms responsible for the missing craters below. For the nearside mare region, a Bouguer correction density of 3100 kg/m$^3$ is also appropriate as a representation of the density of the mare layer. The gravity from relief used in the Bouguer corrections was modeled using a finite-amplitude algorithm (Wieczorek and Phillips, 1998) in the Spherical Harmonic Tools (SHTOOLS) software archive (Wieczorek and Meschede, 2018). To minimize the effects of long-wavelength surface topography on the gravity data, including the nearside-farside asymmetry, we calculated the gravity and gravity gradients on a smoothed representation of the surface topography rather than on a spherical surface (Andrews-Hanna et al., 2018). The smoothing algorithm calculated the potential at the local surface elevation at every point, using a spherical harmonic degree 30 representation of the
surface topography to avoid abrupt jumps in elevation, and the gravity anomaly and gravity gradients were calculated from this potential model. The spherical harmonic degree of 30 represents a half-wavelength of 182 km, and thus the smoothing is not applied to and does not affect the features of interest, which have half-wavelengths of <90 km.

The flooded farside Bouguer gravity was then calculated at an elevation of 1 km above the reference surface to account for the thickness of the lava, based on the mean mare thickness in the non-basin regions of the nearside maria (e.g., de Hon, 1979; Evans et al., 2016; Gong et al., 2016; Head, 1975), so that the modeled gravity of the flooded farside is equivalent to the nearside gravity which was calculated at the mare surface. The Bouguer gravity gradients are used to highlight small-scale anomalies in the gravity data, and were calculated using the maximum amplitude eigenvalue of the tensor of the second horizontal derivatives of the potential (hereafter referred to as the gravity gradients; Andrews-Hanna et al., 2013). The quantitative comparisons between the flooded farside and the nearside were performed in both the spectral domain in the form of localized power spectra, as well as in the spatial domain in the form of crater size frequency distributions, as described in sections 3.3.2 and 3.3.3 below. Identical analyses were performed for the observed gravity of the mare region and the models of the volcanically flooded farside representing different mechanisms for accounting for the crater deficit, as described in section 3.3.4. As discussed below, these analyses consider a wider range of mare density contrasts and thicknesses than the nominal values stated above as possible explanations for the missing crater gravity anomalies.
3.3.2 Power spectra

Localized power spectra were then used to characterize and compare the volcanically flooded farside highlands and the nearside mare regions. The localization process produces a set of spherical harmonic coefficients that optimizes both spatial and spectral concentration of an area defined by a Boolean mask and uses the coefficients to generate the localized power spectrum of that area (Wieczorek and Simons, 2005). For the farside highlands, the localization mask was generated by incorporating a large portion of the farside highlands excluding the South Pole-Aitken basin (SPA) as well as prominent gravity anomalies associated with known discrete structures such as the Moscoviense, Fitzgerald-Jackson, and Dirichlet-Jackson basins (Neumann et al., 2015). The nearside mask (Fig. 1a) was drawn within the mare region, excluding the prominent anomalies associated with the major basins such as Imbrium and Serenitatis (Neumann et al., 2015), as well as the prominent PKT border structures (Andrews-Hanna et al., 2014). We note that the Bouguer correction density of 3100 kg/m$^3$ chosen for the nearside is appropriate for the maria, but results in large anomalies in the nearside feldspathic terrain, including the Imbrium rings, but these areas were excluded in our localization masks.

The crust-mantle interface is the dominant source of the Bouguer gravity signal at low degrees ($l \approx 80$). Since there is a significant difference in the mean crust-mantle interface depth between the nearside ($h_{c,n} \approx 25$ km) and farside ($h_{c,f} \approx 55$ km), there would be a significant difference in the gravity arising from the crust-mantle interface, due to the attenuation of gravity as a function of depth (i.e. distance). As a result, all other things being equal, the Bouguer signal at long wavelengths for the farside is expected to be weaker than the nearside. We thus apply a
degree-dependent correction \((f_l)\) to the low degree Bouguer gravity of the farside to scale the signal up to what it would be if arising from the shallower depth of the nearside crust-mantle interface:

\[
f_l = \left( \frac{R_p - h_{EN}}{R_p - h_{cf}} \right)^l
\]

where \(l\) is the spherical harmonic degree and \(R_p\) is the mean planetary radius. A cosine taper is used to reduce this correction to zero between degrees 30 and 110, since the higher degrees are dominated by shallow density anomalies rather than relief along the crust-mantle interface (Jansen et al., 2017).
**Figure 3.1.** Topography (a), Concatenated Bouguer gravity (b) and gravity gradient maps (c) calculated for surface densities of 3100 kg/m$^3$ for the farside (left) calculated at a 1 km altitude above the reference radius as well as including the effect of increased crust-mantle interface depth as compared to the nearside, and the nearside (right) calculated at the surface with no crust-mantle interface-related tapers applied. The non-rectangular masks in 1a correspond to the areas used for the power spectra localization analyses (Section 2.2) for the lunar farside (F1) and nearside (N1). The rectangular mask (F2) corresponds to the area of focus for analyses in Section 3. The mare-covered regions are highlighted by the black outline (Hiesinger et al., 2011) in all three subplots.

The localized power spectra were then generated in SHTOOLS with a bandwidth of 30 for both the farside highlands and the nearside mare, using all available well-concentrated tapers (eigenvalues > 0.99). It has been verified that the high concentration factor and the difference in localization masks have no significant effect on the resulting spectra (see Supplementary Material). The bandwidth was chosen to maximize the spatial concentration and to minimize the loss of signal at the start and end of the spectral range. Since the crater diameter (~90 km) at which the mare CSFD decreases in comparison to the non-mare CSFD corresponds to the half-wavelength at a spherical harmonic degree of ~60, the bandwidth for the localizations must be less than 60. A bandwidth of 30 was chosen to ensure that the wavelengths of interest are not affected by the localization.
3.3.3 Crater-Detection Algorithm

In addition to using comparisons in the spectral domain to test the results of the models, the ability of the models to reduce crater density of the relevant sizes was evaluated by both visual inspection of the resulting gravity maps, and quantitative comparisons of the crater size-frequency distribution derived from the resulting gravity maps. Comparisons were made between the observed nearside mare gravity and the modeled flooded farside gravity maps.

To reduce subjective bias in determining whether a crater’s signal has been rendered undetectable from the background in the model gravity maps, we created an algorithm to determine whether craters are still preserved or have been erased from the gravity. The algorithm is designed to mirror how we would distinguish craters from the background by checking whether it exhibits the expected gravity or gravity gradient behavior. Buried craters can be distinguished in spatial views in Bouguer gravity as circular zones that are distinct from the background (typically a positive anomaly within a filled crater, though negative anomalies are also observed), or in Bouguer gravity gradients as a continuous rim of positive gravity gradients surrounding an area of negative gravity gradients. Thus, each crater candidate must satisfy one of these two conditions in the algorithm, as the class of features interpreted as buried craters which exhibit strong negative Bouguer gravity signals do not exhibit strong signals in the gravity gradients.

The gravity condition of the algorithm is that the absolute difference between the mean Bouguer gravity within the crater candidate (out to 70% of the crater radius) and the mean of the surrounding areas (out to 130% of the crater radius) exceeds a set threshold (25 mGal), and that
the standard deviation of the Bouguer gravity within the crater candidate rim must be less than a set threshold (15 mGal). The gravity gradient condition of the algorithm is that at least one rim segment spanning >30% of the rim perimeter located along any part of the rim in the Bouguer gravity gradients must be greater than a set value (12 E), and that the median of the gravity gradient values interior to the crater candidate rim must be less than a set value (0 E). The specific values used in the conditions are subjective estimates chosen such that the algorithm provided a reasonable approximation to visual identifications of craters. The algorithm uses the known locations of the craters in the farside test region (Head et al., 2010), and, for comparison, the locations of identified buried craters from (Evans et al., 2018, 2016). While limitations exist (Robbins, 2019) for craters < 30 km in diameter listed in Head et al., 2010, the effect on our results would be insignificant, as it would further highlight the near-farside contrast in CSFD, and the diameters of focus are at around 80 km.

Running the algorithm on the farside highlands with the mask in Fig. 3.1 applied yields a crater detection rate of 78.2%. The missing craters are mostly sourced from smaller-sized craters, whose signals have been heavily modified by the signals of superposed older and larger craters, as well as any subsequent impacts, resulting in craters which may be detectable in topography, but whose signals have been heavily degraded in gravity. The imperfect recovery rate does not affect our results, as the same algorithm is applied to the nearside and flooded farside and the results are used in a comparative sense. Subjecting the algorithm to a false positive test of detecting craters on a longitudinal mirror flip of the lunar gravity map using the same farside population yields a false positive rate of 17.3%. Changing the thresholds used in the algorithm in either direction increases the number of either false positives or of false negatives.
The recovered crater distributions in the algorithms are compared to one another using area-normalized CSFDs between the nearside mare and the flooded farside in their respective regions of interest. We note that the power spectra and recovered crater CSFD are different and independent ways of evaluating the models against the data. The use of both methods is important, as they capture different aspects of the observed gravity field. A particular model may satisfy one of these constraints without satisfying the other, as will be seen below.

3.3.4 Proposed scenarios to explain the observed crater erasure

3.3.4.1 Lava thickness effect

The first hypothesis for explaining the apparent crater erasure is that the layers of mare on the nearside surface are thicker than what is currently estimated (~1 km, e.g., de Hon, 1979; Evans et al., 2016; Gong et al., 2016; Head, 1975). The increased thickness of the mare would result in an attenuation of gravity as a function of depth/distance due to the increased distance between the spacecraft and the mare-substrate boundary (Fig. 3.2b). This attenuation would decrease the amplitude of gravity anomalies arising at the base of the maria, with a wavelength dependence that would preferentially reduce the signatures of smaller craters. This effect can be modeled by changing the reference radius when calculating the gravity on the surface for the flooded farside models. The hypothesis was tested for assumed mare thicknesses of 1–4 km.
3.3.4.2 Reduced density contrast

The second hypothesis is that the density contrast between the surface mare layer and the cratered substrate is less than the estimated value of 600 kg/m$^3$ (Kiefer et al., 2012). This reduced density contrast would diminish the gravity anomaly of all craters, and may make smaller craters undetectable as their signal fails to rise above the background variability (Fig. 3.2c). Such a condition could be possible if, for example, the crust immediately beneath the maria is of intermediate density between mare and pure anorthosite. Candidate materials include low-Ca pyroxene, which has been identified in the PKT region in remote sensing data (Nakamura et al., 2012). Low-Ca pyroxenes include both the magnesium-rich enstatite ($\rho \sim$3200 kg/m$^3$) and the iron-rich ferrosilite ($\rho \sim$3800 kg/m$^3$), but high magnesium content in the nearside mare region (Klima et al., 2011) implies that enstatite is the dominant low-Ca pyroxene species. The 3200 kg/m$^3$ grain density of pure enstatite is in between the noritic anorthosite grain density of $\sim$2900 kg/m$^3$, and nearside mare basalt grain density of 3400 kg/m$^3$ (Kiefer et al., 2012). For comparison, remote sensing constraints on the grain density of the crust indicate a typical density of $\sim$2900 kg/m$^3$ for the farside highlands or $\sim$2950 kg/m$^3$ for nearside highlands, in comparison to $\sim$3050 kg/m$^3$ for the more pyroxene-rich floor of the South Pole-Aitken basin (Wieczorek et al., 2013). Constraints on bulk density of the crust from high-degree gravity and topography yield average surface densities of 2400 kg/m$^3$ within the South Pole-Aitken, compared with $\sim$2300 kg/m$^3$ for the farside highlands and 2800 kg/m$^3$ for the maria (Goossens et al., 2020). Thus, a pyroxene-rich crust beneath the maria may be expected to reduce the density contrast to 400 kg/m$^3$. 
A reduction of the porosity in the crust below the maria by compaction or thermal annealing could decrease this density contrast further, though the porosity extends to substantial depth and we find that thermal annealing is limited to the material immediately underlying the maria (see Supplementary Material). In addition, the mare volcanism could have induced melting and incorporation of the plagioclase substrate into the mare to result in a reduction of the density of the lowermost mare layers by \( \approx 100 \text{ kg/m}^3 \) (Scoates, 2000). Similarly, the mare could have induced melting during ascent through the plagioclase crust (Finnila et al., 1994), incorporating the plagioclase into the mare. The density contrast variations were implemented by changing the density used when calculating the gravity from the mare-flooded surface. Two additional density contrasts were tested as the density contrast between the surface mare and the substrate layer: 400 kg/m\(^3\) and 200 kg/m\(^3\), the latter being a more extreme value than expected requiring a low-porosity and pyroxene-rich substrate and incorporation of substantial re-melted plagioclase in the mare.
Figure 3.2. Illustrations demonstrating each scenario discussed in Section 2.4, including: (a) volcanic flooding by mare, (b) volcanic flooding by a thicker layer of mare, (c) volcanic flooding over higher density crustal material, (d) existence of dense pre-existing material reduces signal from craters, (e) existence of dense pre-existing material that is layered above higher density crustal material, and (f) surface smoothing.

3.3.4.3 Pre-existing dense layer

Next, we consider the possibility that there was a discrete layer of dense material already present on the surface at the time of the cratering, prior to the mare flooding at ~3.6 Ga. The dense layer could be earlier volcanic eruptions that had occurred in the nearside prior to the formation of the sub-mare impact craters (Fig. 3.2d). In addition, the dense layer could also be a result of other vertical variations in crustal composition (Arai et al., 2008). If the dense layer is of similar density to the subsequent mare covering, the boundary between the two layers would be indistinguishable in gravity. If a small impactor formed a crater prior to the final mare flooding, that crater would have failed to completely penetrate through the dense layer to the lower density plagioclase underneath, and as a result the gravitational signal of small craters would be significantly decreased. As the impactor size increases, the resulting crater will eventually penetrate through to the plagioclase and the gravitational signal will approach that of craters formed in the plagioclase crust itself. In this sense, the discrete dense layer acts as a filter preventing smaller craters from penetrating into the underlying low density crust, and reducing the gravity anomaly of intermediate-sized craters. This dense layer can also explain the
existence of some circular negative gravity anomalies interpreted as impact craters that uplifted underlying low density crust into a layer of mare or similarly dense material (Fig 3b of Evans et al., 2016). Fresh complex craters identified in the maria with diameters of 20–80 km have depths ranging from 2.5–4 km (Kalynn et al., 2013), and so a 2-km thick layer of pre-impact dense material should screen out the majority of the signal from 20-km diameter craters, while reducing the gravity anomaly of 80-km diameter craters by half.

This scenario was modeled by creating a density interface at depth that parallels the surface topography outside of the craters. For areas within craters that do not penetrate through the subsurface density interface, the density interface follows a synthetic surface level with the surroundings, generated using the topography power spectrum. For craters penetrating through the subsurface density interface, the interface then follows the crater floor such that the craters extend into the underlying low density feldspathic crust. Craters were identified using the crater database of Head et al., 2010. The density contrast between the dense layer and underlying crust was set equal to the assumed density contrast between the mare and plagioclase of 600 kg/m³. Subsequent mare flooding would then bury both the cratered surface and underlying density interface, but with an assumed density equal to the upper dense layer.

As the pre-mare dense layer scenario was found not to reproduce the buried crater deficit (see results below), we also implemented a hybrid scenario in which pre-mare dense layer overlies a crust of intermediate density between the plagioclase and mare (Fig. 3.2e). As discussed above, this reduced density contrast could be brought about by a crust rich in low-Ca pyroxene-rich or a layer of mare which has incorporated plagioclase from the underlying crust.
3.3.4.4 Surface smoothing

The next candidate mechanism to explain the apparent erasure of craters is some process that physically removed the smaller craters and perhaps modified larger craters as well. Given the selective loss of smaller features, we choose to model this as a diffusional process (Fig. 3.2f), which could include impact-induced modifications to the topography from impact gardening and seismic shaking, as well as thermal and mechanical erosion from high-temperature, long duration lava flows or ponded lava (Byrne et al., 2013; Greeley et al., 1998; Hurwitz et al., 2013b). Impact gardening can include processes such as ejecta deposition, micrometeorite bombardment, and secondary cratering by distal ejecta fragments (Gault et al., 1974). Seismic shaking from basin-forming impacts may be responsible for smoothing the lunar surface prior to the youngest basin-forming impact at Orientale (Kreslavsky and Head, 2012). While impact-induced surface smoothing has been well documented on the Moon (e.g., Fassett and Thomson, 2014), it acts on smaller scales than likely needed to explain the crater deficit. Alternatively, thermal erosion by flowing lava is responsible for significant surface erosion in the formation of lunar rilles (Head and Wilson, 1981) and larger lava channels on Mercury (Hurwitz et al., 2013a), and thermal erosion beneath ponded lava may also have played a role during the early mare eruptions. We defer a more detailed discussion of the causes and implications of the possible erosional smoothing of the pre-mare surface to Section 4.2 below.

While impact-induced surface modification is dominantly diffusional (Fassett and Thomson, 2014; Soderblom, 1970), thermal and thermo-mechanical erosion are not exclusively
diffusive in nature, and likely include both diffusive and advective components. However, heat flow is a diffusional process and mechanical erosion in general entails a diffusive component (Harrison, 2000). For simplicity, we do not explicitly model thermo-mechanical erosion of the pre-mare substrate, which would require a large number of assumptions regarding the nature of the eruptions (e.g., their flow paths and discharge history) and the pre-mare surface, and has typically been modeled as a one-dimensional process (Williams et al., 2000). Rather, we use diffusion as a simple proxy of an erosional process capable of preferentially removing small-scale features, and we highlight thermo-mechanical erosion by lava flows as a candidate mechanism.

We represent this erosion by modeling diffusion of the topography of the cratered highlands surface, using the resulting smoothed relief for calculating the gravity at the interface between lava and underlying crust in the flooded farside analog model. As the purpose of this model is not to explicitly test one specific diffusional process, but to ascertain whether diffusion or physical smoothing of the surface more generally is able to reproduce the observed crater deficits, neither the diffusion constant nor the timestep are physically meaningful, and we set the diffusion constant to 1 m$^2$/s. The diffusion equation in a spherical coordinate system is solved by a 2-D explicit finite difference model, with the timestep set to half of the required Courant-Friedrichs-Lewy stability condition (Courant et al., 1928). The boundary conditions are set as reflective in the latitudinal boundaries. The model is run to $10^7$ iterations, but the resulting maps are checked at intermediate timesteps for comparison to the nearside.
3.3.5 *Comparisons between the Imbrium basin and the Orientale basin*

In addition to probing the cause of the crater deficit, we also seek to understand the reason behind the missing rings of Imbrium in the gravity data where they are expected to lie buried beneath the mare surface. A comparison between Imbrium basin and models of a volcanically flooded Orientale basin can provide insight into how mare flooding can modify a large impact basin. However, there are Bouguer gravitational signals that are present at Orientale that may not be present or as prominent in Imbrium. The negative gravity gradient rings interpreted as dikes intruded into basin ring faults at Orientale (Andrews-Hanna et al., 2018) are absent around Imbrium, possibly due to the prominence of effusive volcanism in comparison to intrusive volcanism as a result of the thinner nearside crust (Wilson and Head, 2017). In addition, Orientale is surrounded by a prominent annulus of thickened but sub-isostatic crust (Andrews-Hanna, 2013; Neumann et al., 1996; Wieczorek et al., 2013), but the effect of a thickened crust from the Imbrium ejecta may have been reduced by crustal relaxation or during the collapse of the transient cavity due to the high heat flow of the region (Ding and Zhu, 2022; Miljković et al., 2016; Mohit and Phillips, 2006).

To be conservative and reduce the contribution to the gravity from the additional sources in Orientale, the flooded gravity map of Orientale was produced by first calculating only the gravity from topography using the assumed density contrast between mare and crust at 600 kg/m$^3$. Then, in order to simulate the contributions to the gravity from small-scale density anomalies or variations in crust-mantle interface relief without including Orientale-specific sources such as the basin ring dikes, a randomized Bouguer map was added to the gravity from topography. The
randomized map was generated using the localized Bouguer power spectrum of a background region of the farside, based on the farside background mask in Fig. 3.1 but modified to exclude additional weaker gravity anomalies within the mask (e.g., Korolev, anomaly at 60°N 120°E to 30°N 150°E). This approach allows us to compare the flooded Imbrium rings to the Outer Rook and Cordillera of the flooded Orientale basin, but the innermost ring of Imbrium and Inner Rook of Orientale surrounding the uplifted mantle plug cannot be compared.

To explain the missing Imbrium basin rings, we consider the same scenarios used to explain the small crater deficit. However, some of the scenarios tested to explain the missing craters cannot be invoked to explain the missing basin rings of Imbrium because of the different nature of these features. A particularly thick mare layer is not considered, as burying the tops of the ring scarps beneath several kilometers of mare is unlikely given their continuity with exposed scarps and isolated massifs, and this scenario was found to be ineffective at explaining the missing craters. The reduced density contrast between the mare and crust as a whole is not considered as the strong gravity gradient signature where the Imbrium rings are exposed and visibly embayed by maria requires a large density contrast between the exposed rings and the embaying mare. Furthermore, we find that the expression of the exposed rings in the Bouguer gravity is minimized for an assumed crustal density of ~2550 kg/m³, indicating that the crust making up the rings is similar to the lunar highlands (see Supplementary Material). Similarly, the pre-mare dense layer would not reduce the gravity signature because of the tectonic nature of the rings (Andrews-Hanna et al., 2018; Johnson et al., 2016; Nahm et al., 2013). This discrete dense layer can act as a filter preventing smaller craters from excavating into the underlying low density crust to explain the missing craters beneath the maria. However, in the formation of the basin
rings, the top and bottom surfaces of this dense layer would be offset equally across the ring faults, and there would be only minor reduction in the gravity signal based on the thickness of the layer and attenuation of gravity with distance. Thus, the only model tested above with relevance to the missing Imbrium rings is the scenario of erosional smoothing of the pre-mare surface by some manner of diffusive mechanism. We model the diffusive smoothing of the topography of Orientale prior to volcanic burial as before, and compare the results with the real signature of the Imbrium Basin.

3.4 Results

3.4.1 Flooded farside vs. present nearside

We first compare the model of the volcanic flooding of the unmodified farside with the observed nearside gravity to quantitatively constrain the differences in recovered craters and power spectra. The initial comparison between the Bouguer gravity gradient of the nearside mare and the flooded farside model (Fig. 3.3b,d) in spatial view reveals a significant difference in the recovered crater density, where the flooded farside has a much higher crater density than the nearside. The relative lack of visible craters in the nearside gravity is consistent with the previously observed crater deficit of the region (Evans et al., 2016).
Figure 3.3. Bouguer gravity and gravity gradient map of a portion of the (a,b) nearside, (c,d) flooded farside highlands, along with corresponding cumulative CSFD (e) and power spectral comparisons (f) of the high mare thickness models. In panels a and c, the craters recovered by
the algorithm are highlighted by white circles, the center of the recovered craters is marked by green triangles, and the craters not identified by the algorithm are marked by white triangles. In addition, in panel c the black outlines represent areas excluded from F1 (Fig. 3.1a) due to the prominent impact features in the gravity gradient.

The cumulative CSFD plots of the craters recovered by the algorithm from the flooded farside models and the observed nearside (Fig. 3.3e) provide a quantitative comparison of the crater populations. The algorithm recovers 78.2% of the known crater population in the flooded farside, and 66.7% of the craters identified by Evans et al., 2018, 2016. Given that there are a number of competing gravity anomalies making the identification of more degraded and overlapping craters in the gravity data challenging in both areas, and that there is some degree of subjectivity in the identification of the buried craters in the maria, these rates are considered to be acceptable.

The original crater size frequency distribution in Evans et al., 2018 shows a small deficit in the mare and non-mare populations starting at the 80-100 km-diameter bins, with a greater deficit at smaller diameters. Our cumulative CSFDs of the nearside and flooded farside show overlapping distributions between the two areas at >80 km diameter bin, but shows a clear deficit beginning in the >60 km diameter bin (Fig. 3.3e), indicating that the deficit begins at diameters <80 km. Thus, despite using very different methods (automated vs manual identification of craters, different representations of the gravity gradients, and our comparison to a model of the volcanically flooded farside), and despite the subjective nature of identifying poorly preserved
craters in gravity data, we arrive at a similar conclusion regarding the deficit of small craters. The CSFD of recovered large craters (>140 km-diameter) also show a departure from the similarly-sized craters in the nearside as well, which may be due to either the difficulty of recovering ancient eroded structures on the nearside, the differences in the crater populations themselves (e.g., Miljković et al., 2013), the effects of small number statistics, or crater retention age.

Power spectral comparisons provide insight into the degree ranges over which the flooded farside model power spectrum deviates from the nearside power spectrum, capturing effects not only of the missing craters, but surface roughness on the buried interface in general as a function of scale. The comparisons of the nearside and the flooded farside power spectra (Fig. 3.2f) show that the power spectra are roughly equivalent from degrees 55 to 110, but the nearside power drops to a factor of 4 below that of the flooded farside model from degree 150 to the end of our spectral degree range. The point at which the two spectra begin to diverge (degree 110) corresponds to a half-wavelength of ~50 km, and thus is to first order consistent with the crater deficit we find beginning at diameters <80 km.

3.4.2 Gravity attenuation due to increased mare thickness scenario

We first consider whether the apparent deficits in the nearside small crater population and gravity power spectra could be an effect of the thickness of the overlying maria due to the wavelength-dependent attenuation of gravity with distance. We consider mare thicknesses of up to 4 kilometers, which is four times higher than the estimated average thickness of the mare plains (e.g., de Hon, 1979; Evans et al., 2016; Gong et al., 2016; Head, 1975). A comparison of
the cumulative CSFD of the models for a range of mare thicknesses to the cumulative CSFD of
the nearside (Fig. 3.4e) shows that the loss of recovered craters due to the reduction of gravity
from greater mare thicknesses is negligible (roughly 10%) compared to the observed crater
deficit of ~80%. Similar results are seen in the power spectral comparisons, as even the 4-km
maria spectrum cannot fully account for the power spectral deficit (Fig. 3.4f). The spectrum
modeled using an assumed thickness of 4 km of lava in the flooded farside model matches the
nearside spectrum for L>350, but overestimates the spectrum between 130–350, which
corresponds to half wavelengths of 16–42 km, and is consistent with the cumulative CSFD result.
Additionally, although a mare thickness of greater than 4 km is likely attained in impact basins
(Solomon and Head, 1980), such deposits are inconsistent with mare thickness estimates of the
broader region outside of basins (DeHon, 1979; DeHon and Waskom, 1976; Evans et al., 2016;
Gong et al., 2016; Hoerz, 1978). Thus, even this overly large mare thickness cannot explain the
deficit in both the mare region crater population and power spectrum.
**Figure 3.4.** Bouguer gravity and gravity gradient maps of a portion of the nearside (a,b), and flooded farside highlands using topography from models calculated assuming a 4-km thick mare (c,d), along with corresponding cumulative CSFD plots (e) and power spectral comparisons (f) for mare thickness of 2 and 4 km.

### 3.4.3 Reduced density contrast scenario

We next consider whether the apparent deficits in the nearside small crater population and gravity power spectra could be an effect of a reduced density contrast between the mare and the underlying material, thus reducing the gravity anomalies of the buried craters. In comparison to the lava thickness effect, the cumulative CSFDs of the reduced density contrast models come much closer to matching the cumulative CSFD of the nearside (Fig. 3.5e). However, the shapes of the CSFDs are significantly different. The 400 kg/m$^3$ model overpredicts the CSFD from 20–60 km, covering most of the diameter range of interest. The 200 kg/m$^3$ density contrast model overpredicts the crater population from 20–40 km, then underpredicts from 60–120 km. Comparing the shapes of the CSFDs, the reduced density contrast model preserves the shape of the farside CSFD while shifting it slightly downwards, reflecting a uniform decrease in the recovery rate across all diameters. Similarly, setting the density contrast between the mare and the substrate layer to 400 kg/m$^3$ failed to account for the power spectrum deficit (Fig. 3.5f). However, a 200 kg/m$^3$ density contrast in the flooded farside model is able to match the power spectrum of the nearside at the high degrees (>150). This low density contrast would require both a less dense mare layer and a more dense substrate layer, as could arise from emplacement of
mare that had melted and entrained a substantial amount of plagioclase above a low-porosity and pyroxene-rich crust, as discussed above.
Figure 3.5. Bouguer gravity and gravity gradient map of a portion of the (a,b) nearside, flooded farside highlands using topography from models calculated using (c,d) 200 kg/m$^3$ density contrast between mare and substrate, along with corresponding cumulative CSFD plots (e) and power spectral comparisons (f) of the density contrast models.

3.4.4 Pre-existing dense layer scenario

The pre-existing dense layer scenario assumes that a discrete layer of material with similar density to the mare was already present on the surface before the sub-mare crater population formed. This layer would have prevented smaller craters that formed before the final mare flooding from penetrating through to the less dense plagioclase-rich crust layer, thus reducing the gravity anomalies of those buried craters. The thickness of this dense layer is varied between 2–3 km. The 2-km pre-mare dense layer model provides the best fit to the cumulative crater density at 20 km (Fig. 3.6e) and it reduces the number of smaller craters recovered while preserving the number of larger craters. However, this model also predicts substantial loss of craters for the 100 km diameter bin and projects a lower than observed cumulative crater density for crater diameters of 40–90 km. Similarly, a 2-km layer pre-mare dense layer also reduces the power spectrum of the flooded farside by a factor of two, but not enough to match the observed power spectrum (Fig. 3.6f). The 3-km pre-mare dense layer generates similar results to the 2-km models, providing a slightly better fit to the cumulative CSFD at the smaller diameters but a poor fit at the larger diameters, while more closely matching the power spectrum.
We also implemented a combination scenario, in which the pre-mare dense layer and overlying maria have a density contrast relative to the substrate layer lower than expected (200 kg/m$^3$). The modeled gravity gradient map and recovered CSFD of this model better resemble those of the nearside mare in comparison to the models considering only increased mare thickness or reduced density contrast (Fig. 3.6c). The cumulative CSFD and power spectra (Fig. 3.6e, f) of this model provides reasonable fits to the nearside at the small diameter bins (20-60 km), and at the larger spherical harmonic degrees (> 150), respectively, while still preserving the signatures of the large craters.
Figure 3.6. Bouguer gravity and gravity gradient map of a portion of the (a, b) nearside, flooded farside highlands using topography from models calculated using the pre-mare dense layer model with (c, d) 2-km of material and 200 kg/m³ density contrast, along with corresponding cumulative plots (e) and power spectral comparisons (f) of the pre-mare dense layer and density contrast models.

3.4.5 Diffusive surface smoothing scenario

We then consider whether the apparent deficits in the nearside small crater population and gravity power spectra could be an effect of some manner of surface smoothing and degradation of crater topography resulting in a smoothed density interface at the base of the maria. This smoothing is represented by a simple diffusional process, but actual mechanisms may differ. In order to obscure the small crater population, it is not necessary to fully remove the craters, but rather to mute the expression of their rims in the gravity gradients, which is the primary means of identifying their signature (length scales of order 10 km for a 90 km-diameter crater). The diffusion model results are presented for dimensionless times ($\tau_D$) normalized relative to the diffusion timescale of $D^2/(\Delta t \cdot \kappa)$ for a length scale of $D=10$ km and a model timestep of $\Delta t=100$ s, making the results independent of the particular diffusivity or absolute timescale.

Running the diffusion model on the lunar topography to both 0.5 $\tau_D$ and 1 $\tau_D$ prior to mare flooding predicts crater deficits at diameters <80 km, with the 1 $\tau_D$ model coming closer to matching the small crater density (Fig. 3.7g). Both models also preserve the recovery of the large diameter craters. In contrast, the 3$\tau_D$ model results in a crater deficit at diameters <140 km, in
conflict with the observations of the maria. We also note that in the nearside maria (Fig. 3.7a,b), the gravitational signatures of those craters that are observed commonly lack short wavelength rim signatures in the gravity gradients, as also found in the results of the diffusion models (Fig. 3.7c-f).

Diffusion to $0.5–1 \tau_D$ provides a reasonable match to the power spectrum over different degree ranges, while $3 \tau_D$ predicts an overly low power spectrum throughout the region of interest (Fig. 3.7h). The slightly different slope of the modeled power spectrum compared to the observed spectrum may indicate that the observed surface smoothing was not purely diffusional in nature (e.g., an advection-diffusion process that may both create and destroy relief) and may include effects such as differences between the assumed density contrast used in the Bouguer corrections and the true density contrast between the mare and the substrate.
**Figure 3.7.** Bouguer gravity and gravity gradient map of a portion of the (a,b) nearside, flooded farside highlands using topography from models diffusing to (c,d) $\tau_D$ and (e,f) $3\tau_D$. Cumulative plots (g) and power spectral comparisons (h) of the nearside mare, flooded farside model, and the flooded farside model calculated using topography after modifications by the diffusion pipeline.

The topography map diffused to $\tau_D$ shows significant visible deviations from the original topography (Fig. 3.8). The topographical signatures of small craters (<60 km) have all but disappeared, and the signatures of medium sized craters (~60-120 km) have been significantly reduced. Only the topographical features of the largest craters remain largely unaffected. This diffused topography may represent the relief of the mare-substrate contact if surface smoothing is responsible for the deficit of craters in the gravity data.

**Figure 3.8.** Original and diffused to $\tau_D$ (a,b) topography of a selected area in the farside region.
3.4.6 Comparisons between the Imbrium basin and the Orientale basin

As discussed above, only portions of the Imbrium basin’s rings are exposed and detectable in topography and gravity data (Fig. 3.9c,e). Where the rings do not rise above the mare surface and are presumed to be buried beneath the mare, we found no gravity signature of the buried rings (Fig 3.9c,f).

The Orientale basin may serve as an analog of the Imbrium basin before it was flooded by the maria (Head, 1982), as it is of roughly similar dimensions (Orientale at 930 km diameter, Imbrium at ~1150 km diameter). Orientale basin has three prominent ring structures, the Inner Rook, the Outer Rook, and the Cordillera rings (see discussions of ring morphology in Hartmann and Kuiper, 1962; Head, 1974; Wilhelms et al., 1987). The outer and middle ring remnants of Imbrium are analogous to the Cordillera and Outer Rook rings, which are thought to have formed as tectonic ring scarps (Andrews-Hanna et al., 2018; Johnson et al., 2016; Nahm et al., 2013). The inner ring remnants of Imbrium are thought to be analogous to Inner Rook ring, which is interpreted as having formed as an uplifted peak ring around the central basin cavity (Johnson et al., 2016). Signatures of the rings of Orientale are visible both in topography, where they fully encircle the basin and extend for up to 100 km in width and 4 km in height, and in the free air gravity as the short wavelength structure of the rings is uncompensated. In the Bouguer gravity gradients, they are expressed as either paired positive and negative rings of amplitude ±30 E, representing tectonic offsets across a shallow density interface, or as negative rings of amplitude -30 E, representing ring dikes intruded into the underlying faults (Andrews-Hanna et al., 2018).
Each of the three rings is continuous around the basin, with only modest variations in relief (Nahm et al., 2013).

As are rings around most lunar basins where they have not been modified by subsequent impacts or volcanism, the rings of Imbrium were likely continuous at the time of formation as well. Based on comparisons to Orientale and other multi-ring basins, multiple continuous rings fully encircling Imbrium should have been present after its formation, and a continuation of the ring system buried beneath the maria is expected (Head, 1974). Isolated massifs in westernmost Mare Imbrium may be expressions of buried rings, supporting their continuity. However, the gravitational signature of the buried rings is not apparent in the data.

A comparison between the Imbrium basin and models of a volcanically flooded Orientale basin can provide insight into the expected signature of a mare flooded impact basin. As described in the methods, we calculated the gravity of the volcanically flooded surface topography of Orientale and added a Bouguer model with randomized spherical harmonic coefficients generated from the farside highlands power spectra to represent subsurface variations in density not associated with the basin structure. The rings are clearly visible and continuous around most of the basin in the gravity gradients, with amplitudes of up to ±50 E (Fig. 3.9d). The negative and positive anomalies are paired because they are generated by a step in the density interface at the ring scarps. There are areas where rings of Orientale are not clearly visible in this flooded farside gravity model, but they are limited in extent (e.g., the southwest quadrant of Cordillera) and it is unlikely that all of the flooded Imbrium rings resemble that area. In contrast, there is no gravitational signature of the rings around Imbrium where buried by mare,
including the region of the outer ring continuing westward from the Montes Carpathus (Fig 3.9c,e). In addition, clear gravity gradient signatures of rings are absent in western Mare Imbrium and in Mare Frigoris. Thus, we conclude that the rings of Imbrium where they are buried by the maria, with expected buried relief of up to 4 km and widths up to 100 km, should be clearly visible in the gravity data and yet are undetectable.

As discussed in the section 2.5 above, the only model tested above with relevance to the missing Imbrium rings is the scenario of erosional smoothing of the pre-mare surface by some manner of diffusive mechanism, as the gravitational signature of the exposed rings of Imbrium are not compatible with any of the reduced density contrast scenarios. Here we test whether the modeled gravity from a diffused model of Orientale can reproduce the non-detection of the volcanically buried Imbrium rings in the gravity. The results (Fig. 3.9d,g,h) from applying the diffusion model on the Orientale topography show that diffusion to a time of $\tau_D$ erases most of the ring signature of Orientale, as diffusion to $\tau_D$ leaves behind discontinuous arcs of rings with gravity gradients reaching values of $\sim 30$ E, but these discontinuous anomalies are obscured by the background variability in the gravity field in most places. The gravity gradient signatures of the rings in this model are comparable to what is currently observed for the buried rings of Imbrium basin. Taken together with the results of the previous sections, we find that some manner of surface smoothing process, of which we take diffusion as a proxy, during or prior to the early mare eruptions can explain both the missing craters and the missing rings of Imbrium. Although we cannot rule out the possibility that Imbrium never had a complete ring system, in contrast to expectations based on other lunar basins, the similarity in scale between the missing basin rings and the missing craters supports a similar process being responsible for both.
Figure 3.9. Maps of the topography for Imbrium (a) and Orientale (b). Bouguer gravity gradient map for Imbrium (c) compared with the lava-flooded model of Orientale (d). Topography of Imbrium with proposed ring locations outlined (e). Rectangle indicates location of panels outlining the southern portions of Imbrium where rings are expected to be present but are only partially visible in both topography and gravity gradient (f), and Bouguer gravity gradient map calculated using topography modified by the diffusion model at $0.1\tau_D$ (g), and at $\tau_D$ iterations (h), on the lava-flooded model of Orientale Basin.

3.5 Summary and Discussion

3.5.1 Missing craters and basin rings

In summary, a dearth of buried craters <90 km in diameter as revealed by gravity data was found in mare-covered regions in comparison to craters in the non-mare covered regions. A similar deficit is observed when comparing the CSFD recovered from the gravity for the nearside mare region with the CSFD recovered from the gravity of the modeled mare-covered farside. The comparison between the localized power spectra of the nearside mare region and the model of a mare-flooded farside similarly yields a power deficit in the mare region of roughly a factor of 4. In addition, there is no gravitational signature of the rings of Imbrium where they are inferred to be buried beneath maria, in contrast to models of a mare-flooded Orientale basin.

Models assuming a thicker mare layer or a reduced density contrast between 200 and 400 kg/m$^3$ are unable to match the nearside gravity data in both the spectral and spatial domains. A thicker mare layer cannot explain the crater deficit, and can only explain the power spectrum
deficit for an unrealistic thickness of 4 km. Although the model assuming a reduced density contrast of 200 kg/m$^3$ between the mare and underlying crust was able to match the power spectrum, it did not match the CSFD. The model assuming a 2-km pre-mare dense layer was able to fit the spectral data as well as the recovered CSFD. Although this model did not provide a perfect fit to the CSFD, it reduced the recovered fraction of small craters relative to large craters, consistent with the observations.

However, none of the models above can explain the missing basin rings of Imbrium in the Bouguer gravity, since the gravity signatures of the rings where exposed and embayed by mare require low density materials making up the rings and a large density contrast relative to the surrounding maria. Models simulating diffusive smoothing of the surface beneath the maria provide the best fit to both the crater deficit and power spectrum of the nearside maria, as well as the missing basin rings of Imbrium, and thus is the preferred hypothesis behind the observations of the nearside mare region. The models of the gravitational signatures of the diffused cratered surface also matched the character of the craters found in the nearside mare region, many of which lack short-wavelength rim signatures. Since both the discrete layer of intermediate density and the diffusional smoothing models can explain the crater deficit, the implications of both are discussed below.

3.5.2 Implications of the pre-mare dense layer and the lowered density contrast scenario

The model representing a 2-km thick pre-mare dense layer fit the nearside data in both the spectral and spatial domains, though it fails to explain the missing rings of Imbrium. In this
model, the layer of dense material must predate the crater population, and thus must date to the pre-Nectarian and perhaps to the stabilization of the crust itself. The simplest means to bring about such a dense layer is through ancient volcanic eruptions. A very early period of volcanism, perhaps distinct from the later mare volcanism, is predicted by some thermochemical models as a result of the combination of a warm starting condition and the concentration of heat producing elements in the PKT (Laneuville et al., 2018, 2013). Fragments of mare basalt in the Apollo samples provide evidence for mare volcanism as early as 4.2-4.3 Ga (Dasch et al., 1987; Lawrence et al., 2008; Taylor et al., 1983). There is evidence as well for dikes stalled beneath the farside highlands during farside magmatic activity in the pre-Nectarian era that may have been concurrent with pre-Nectarian nearside volcanic activities (Liang and Andrews-Hanna, 2022).

Alternatively, the models with a reduced density contrast of 200 kg/m$^3$ between the mare and the underlying crust, with or without a pre-impact dense layer, can also match the data. The reduced density contrast of 200 kg/m$^3$ between the mare and the substrate is consistent with the existence of an intermediate layer enriched in low-Ca-pyroxene (e.g., enstatite, grain density $\sim$3200 kg/m$^3$), which has been detected by the Kaguya/Selene mission (Nakamura et al., 2012). This pyroxene-rich material could be a result of the differentiation of the melt sheet of a putative Procellarum impact basin (Nakamura et al., 2012), by analogy with the pyroxene rich floor and presumed impact melt sheet of the South Pole-Aitken basin (Hurwitz and Kring, 2014; Nakamura et al., 2009). Alternatively, this unique composition may reflect the unique thermal and magmatic evolution of the PKT region (Laneuville et al., 2013; Wieczorek and Phillips, 2000). Thus, there is some support for the existence of an ancient dense layer between the mare and the underlying feldspathic crust and/or a reduced density contrast between mare and crust,
whether from ancient volcanism, the putative Procellarum impact basin, or the early evolution of the PKT.

However, this scenario cannot explain the lack of a gravitational signature of the buried Imbrium rings. For a thin (~2 km) layer of dense pyroxene-rich material, formation of the rings by tectonic offsets across ring faults would offset the interface between this dense layer and the underlying crust, resulting in only slightly reduced gravity signatures of the rings. Moreover, the gravitational signature of the exposed Imbrium rings indicates a low density consistent with typical highlands materials (see Supplementary Material).

3.5.3 Implications of the surface smoothing scenario

3.5.3.1 Impact-induced diffusion

The model of surface smoothing by diffusional processes was also able to produce results that fit the nearside power spectrum as well as the nearside cumulative crater size-frequency distribution. Importantly, the diffusion model can also explain the missing rings of Imbrium where buried beneath the maria. However, the diffusion model as implemented was not specific to a particular process, but rather represents some manner of erosional process that preferentially removes short wavelength features. The observed topographic degradation of the lunar surface is strongly controlled by diffusive processes (Fassett and Thomson, 2014), which can include seismic shaking, ejecta deposition, micrometeorite bombardment, and secondary cratering by distal ejecta fragments.

Seismic shaking from basin-forming impacts can have a strong diffusional effect on surface topography (Kreslavsky and Head, 2012). In addition, the ejecta blanket as a result of the
Imbrium impact could have filled the small craters and greatly lowered the gravitational signal of larger ones, as observed on the farside around Orientale (Fassett et al., 2011). The nearside mare region encompasses several major basins, including Nectaris, Serenitatis, and, most importantly, Imbrium, which would have blanketed their surroundings with ejecta. If a basin were to have significantly altered the topography of the nearside pre-mare craters before the final mare flooding, the most likely candidate would be the largest basin most central to the region and youngest in age, Imbrium basin. However, a qualitative consideration of the radial distance from Imbrium basin and either the observed gravity signal or recovered crater population (Figs. 1b-c, 2a) reveals no obvious correlation. In terms of ejecta burial, the area in the nearside mare within our region of interest (Fig. 3.2a) extends to roughly 3× the radius of the outermost ring, which may only be buried by ~100 m of ejecta if the ejecta thickness profile is similar to that around Orientale (Fassett et al., 2011). The smallest crater of interest in our studies is 20 km in diameter, which corresponds to a typical depth of >2.5 km (Kalynn et al., 2013). More importantly, ejecta and seismic shaking at Orientale have not obscured its rings, and thus would not be expected to do so for Imbrium. Thus, Imbrium effects, including both seismic shaking and ejecta burial, can be ruled out as the main mechanism of crater and basin ring erasure.

Diffusive modification of the lunar surface continues to be a pervasive process modifying craters and other small-scale features (Kalynn et al., 2013). It has been shown (Minton et al., 2019) that secondary cratering by small distal ejecta fragments is likely the main source of diffusive topographic degradation of the lunar surface for meter-scale or larger features, as this mechanism controls the crater saturation equilibrium. While present-day rates of modification are low and limited to small-scale features (Fassett and Thomson, 2014), the rate and scale of
this modification would have been greater earlier in lunar evolution when the impact flux was higher (e.g., Hartmann, 1966). However, the nearside pre-mare craters and exposed farside craters would have experienced similar modification by secondary cratering. Thus, it is unlikely that those processes would have resulted in diffusional modification of the pre-mare nearside topography beyond that already expressed in the lunar farside. The main driving process that erased the pre-mare craters must have been unique to the mare-region. With that in consideration, the prime candidate process that would be unique to the nearside mare region is thermo-mechanical erosion from the mare flooding itself.

3.5.3.2 Thermal and thermo-mechanical erosion

Erosion by flowing lava is well documented on Earth (Greeley et al., 1998), the Moon (Hurwitz et al., 2013b), and Mercury (Byrne et al., 2013; Hurwitz et al., 2013a). For Mercury, this process has eroded valleys up to 25 km wide and 1 km deep (Hurwitz et al., 2013a). Thermal or thermo-mechanical erosion by the maria could cause smoothing of the mare substrate without affecting areas not covered by mare, and thus could explain the striking difference between the gravity of the mare region and the modeled gravity of mare-covered highlands. Thermal erosion would selectively affect smaller scale features, consistent with the observations. Erosion by the mare flows may also help to explain the step-like transitions at the highland-mare contact in some places, previously interpreted as rings of a Procellarum impact basin (Whitaker, 1981). However, thermal and thermo-mechanical erosion are not strictly diffusional processes, and likely include both diffusive and advective components, depending on the nature of both the flows and the substrate. As discussed above, detailed modeling of the potential for erosion by the mare flows is beyond the scope of this study. Instead, we simply note that the gravity of the
mare region requires preferential smoothing of the small-scale relief, with total elevation changes on the order of hundreds of meters required.

A number of previous studies have considered simple 1-D idealized models of erosion beneath lava flows on both Earth and the Moon (e.g., Kerr, 2009, Williams et al., 2000). Erosion by the lavas may have taken one of four forms: thermal erosion by laminar flows, thermal erosion by turbulent flows, thermal erosion by ponded lava, and thermo-mechanical erosion by turbulent flows. Of these, thermal erosion by laminar flows is the least efficient mechanism (Greeley et al., 1998), as the bottom boundary of the flows rapidly cools by conduction, and we do not consider it further. Thermal erosion by turbulent flows results in higher erosion rates, as they are well mixed and maintain higher temperatures at the ground surface (Huppert et al., 1984; Huppert and Sparks, 1985; Williams et al., 2001, 2000). The intense heat and long duration of the mare flows could have melted and swept away large portions of the crater topography. Thermal erosion has been known to produce some sinuous rilles on the surface of the Moon (Huppert et al., 1984; Huppert and Sparks, 1985; Williams et al., 2001, 2000), though many rilles likely formed as collapsed lava tubes and did not involve thermal erosion. The typical rille depth is of order 100 m, in comparison to the mean (and 95% range) in topography reduction for the 0.5 and 1 $\tau_D$ models of 324 (7.5-1040) m and 406 (11.1-1280) m, indicating the ability for even these small flows to generate erosion at the same order of magnitude as that needed to explain the gravity observations. However, sinuous rilles are inferred to have formed from flow volumes of $<$100 km$^3$ (Hurwitz et al., 2013b), which is five orders of magnitude smaller than the estimated volume of the nearside maria at $\sim$1×10$^7$ km$^3$ (de Hon, 1979; Evans et al., 2016; Gong et al., 2016; Head, 1975), and thus the flooding of the nearside maria has
potential for much greater erosion. The potential for thermal erosion of lava flows much more voluminous than those that produced sinuous rilles is difficult to constrain, particularly given the lack of constraints on the nature of the earliest mare eruptions. Thermal erosion rates by lunar basalts have been estimated to be in the range of 0.1–1 m/day (Hurwitz et al., 2012; Williams et al., 2000), while thermal erosion for flows forming channels on Mercury have been estimated to be in the range of 1.8-17 m/day (Hurwitz et al., 2013a). For the 95% range in topography reduction from the $1\tau_D$ model of up to 1280 m of topography reduction, these rates would require 0.2–35 years to accomplish. For comparison, flood basalt eruptions on Earth can last up to 14 years with volumes up to 1300 km$^3$ (Self et al., 2006).

Thermo-mechanical erosion is able to produce a higher erosion rate of the substrate (up to 20 m/day; Williams et al., 1998). Estimated thermomechanical erosion rates for channels on Mercury are up to 130 m/day (Hurwitz et al., 2013a). On Earth, thermo-mechanical erosion of ~100 m requires a turbulent, high volume (>100 km$^3$), long flow (>100 km) of a high temperature lava such as a komatiite kept at its liquidus, as well as an unconsolidated, high water content substrate allowing for the vaporization of water to dislodge small-sized particles from the rock surface (Williams et al., 2001). The high volume and long flow conditions are easily met in the nearside mare. While the lack of water makes thermo-mechanical erosion less likely, the high porosity of the lunar highland surface (Besserer et al., 2014; Wieczorek et al., 2013) could aid in the dislodging of surface material by mechanical means. However, mechanical incorporation of substrate material in lunar flows is estimated to only raise the erosion rate a small amount (~12%, Williams et al., 2000).
Given that lavas erupt more readily in areas of thinner crust (Head and Wilson, 2017), at least some of the mare eruptions may have occurred in topographic lows and ponded rather than flowing (Greeley and Womer, 1982). In such a lava pond, the compositional density contrast between the lavas and the melt generated from the underlying substrate (Finnila et al., 1994) can drive compositional convection, as happens with basaltic intrusions into felsic crust (Kerr, 1994) and laminar flow of basalt over felsic crust (Kerr, 2009), which allows for the estimation of the rate of thermal erosion. The melting rate for thermal erosion at the base of a lava layer driven by compositional convection due to the buoyancy of the melting substrate is given by (Kerr, 1994):

\[
V = 0.35 \left( \frac{g(\rho_l-\rho_m)k_l^2}{6.22\mu_mS^4} \right)^{\frac{1}{2}} \left( 1 + \frac{k_l}{k_mS} \right)^{-\frac{1}{2}}, \quad \text{where} \quad S = \frac{\rho_sL_s + \rho_sc_s(T_m-T_g)}{\rho_lT_l - \rho_mT_m}.
\]

(2)

where \(\rho_l\) and \(\rho_m\) are the densities of the lava and melted substrate, \(\mu_m\) is the viscosity of the melt \((10^6 \text{ Pa}s; \text{Hummel and Arndt, 1985})\), \(k_s\) and \(k_m\) are the thermal conductivities of the lava and melt (taken to be equal at 3 W/m K), \(\kappa_l\) is the thermal diffusivity \((10^{-6} \text{ m}^2/\text{s})\), \(T_m\) is the melting temperature of the substrate \((1050^\circ\text{C})\), \(T_l\) and \(T_g\) are the lava and substrate temperatures \((1300^\circ\text{C} \text{ and} 0^\circ\text{C})\), \(L_s\) is the latent heat of fusion of the substrate \((8\times10^5 \text{ J/kg})\), and \(c_l\) and \(c_s\) are the specific heats of the lava and substrate (here both taken to be 1340 J/kg \(^\circ\text{C}\)). For basalt over plagioclase on the Moon, the density contrast between the basaltic lava and plagioclase melt is \(\rho_l-\rho_m=200\) kg/m\(^3\) (Scoates, 2000), leading to a melting rate of \(~10\) m/year. However, this melting rate may be short-lived, as settling of dense crystals to the base of the magma layer will halt thermal erosion (Kerr, 1994).
Whether thermal or thermomechanical erosion are capable of smoothing the pre-mare substrate depends on the nature of the earliest mare eruptions, which is poorly understood. Observations are limited to the topmost flows in a sequence of mare basalts averaging ~1 km in thickness (e.g., de Hon, 1979; Evans et al., 2016; Gong et al., 2016; Head, 1975). Thermal or thermomechanical erosion by flowing lava would require sheet-like flows from distributed sources affecting the entire mare region, rather than channelized flows as observed in exposed terrains on the Moon and Mercury. Given the unconstrained nature of the earliest mare eruptions, such flows cannot be ruled out. The latest formed flows at the surface involved large volumes and long runout distances by terrestrial standards, indicating high discharge rates (Head and Wilson, 1992; Yingst and Head III, 1998). Large areas of the mare surface are not divided into distinct flow units and may have involved more sheet-like flow of lava. The earliest flows may have involved even higher discharges and greater capacity for thermal erosion, given the unique nature of these flows arising from the strongly heated PKT region (Laneuville et al., 2013) and possibly caused by mantle overturn events (Zhong et al., 2000). Indirect evidence for an early phase of voluminous outpourings of lava is found in the topography of the mare region and distribution of mare thickness. The surface throughout the mare region is remarkably uniform in elevation, despite variations in mare thickness (and thus pre-mare relief) of up to 9 km (DeHon, 1979; Evans et al., 2016; Solomon and Head, 1980). This distribution could be brought about by either locally sourced eruptions tuned to match the pre-mare relief to fill all areas to the same level, or long-distance transport of lava from source to sink so as to fill in impact basins and other depressions while resulting in a relatively uniform surface elevation. In the latter scenario, long-duration sheet-like flows of lava and lava ponding may result (Greeley and Womer, 1982),
maximizing the potential for erosion by the flows. If the prominent gravity anomalies bordering the nearside mare region are the magma plumbing system responsible for the eruptions (Andrews-Hanna et al., 2014), their distribution would imply long-distance transport from sources distributed around the margins of the maria toward the central mare region, consistent with distributed volcanic erosion throughout the region.

3.6 Conclusions

A dearth of craters <90 km in diameter was found in the lunar crater populations in mare-covered regions in comparison to craters in the non-mare covered regions, consistent with previous analyses (Evans et al., 2018). Similarly, a comparison between the localized power spectra of the nearside mare region and a model of the volcanically flooded mare proxy region in the farside highlands shows a power deficit of the mare region of roughly a factor of 4. In contrast to the continuous rings of Orientale and other basins on the Moon, the majority of the rings of Imbrium are largely absent from surrounding topography and have no recognizable gravity signal where they are presumed to be buried beneath the maria.

To put these observations in perspective, given a deficit in the N(>20) of ~800 craters per $10^6$ km² (Fig. 3.3e), and a mare surface area of ~6×$10^6$ km² (Head, 1975), the observed crater deficit resolved by the GRAIL data corresponds to over 4500 missing craters greater than 20 km in diameter. Similarly, assuming initially continuous rings around the Imbrium basin, only ~10% of the middle (Alpes) ring and only ~33% of the outermost (Carpathus-Appeninus-Caucuses) ring are exposed, leaving arcuate chains of mountains stretching 5000 km in length, ~100 km in width, and several km in height that should be buried beneath the mare surface but are
undetected in the gravity data. In contrast, observations of abundant buried craters in both topography data and their tectonic signatures on Mars and Mercury show that buried craters in this size range are preserved on those bodies (e.g., Byrne et al., 2016; Frey, 2006; Ostrach et al., 2015). These dramatic deficits on the Moon require an explanation.

We proposed and tested a series of scenarios that could have resulted in the observed deficit in buried craters in the mare region in comparison to craters of similar sizes in non-mare regions. The scenarios include a much thicker nearside mare, a lower density contrast between mare and substrate, the presence of a discrete layer of dense material before the pre-mare crater population formed such that smaller craters were unable to excavate to the plagioclase underneath, and finally, some manner of erosional process that significantly smoothed the substrate topography before being flooded by the mare.

The scenarios were tested by comparing both the Bouguer localized power spectra as well as the crater density of the mare proxy region, modified to simulate the scenarios, to those of the nearside mare. However, while both the discrete layer of dense material and the diffusional smoothing models can explain the crater deficit, the discrete layer model cannot explain the missing basin rings of Imbrium in the Bouguer gravity, which leaves erosional processes as our preferred hypothesis. While erosional processes that modify the lunar surface include micrometeorite bombardment, secondary cratering, seismic shaking, ejecta burial, and thermal erosion, the processes related to impacts would also have occurred in the farside region, and comparisons of the buried crater density near and far from Imbrium exclude basin effects. Thus, we favor some manner of thermal or thermo-mechanical erosion of the substrate by lava as
the main mechanism, which is the only mechanism capable of explaining the deficit of small nearside craters, the deficit in the gravity power spectrum, as well as the missing rings of Imbrium. At the very least, some amount of thermal erosion is certain to have occurred during the initial mare floodings that flowed over the pre-mare crustal material.

We emphasize that thermal erosion of the scale necessary to explain the gravity observations would be unprecedented and vastly exceeds the scale of documented thermal erosion on Earth and other bodies. However, the deficit in small craters buried beneath the maria, the anomalous Bouguer power spectrum over the maria, and the missing Imbrium rings clearly require some process to have smoothed out the pre-mare surface on scales less than ~80 km over an area exceeding $5 \times 10^6$ km$^2$. Given the similarly unprecedented scale of the mare flows and the poorly understood nature of the early mare eruptions, the possibility of high discharge sheet-like flows or ponded mare capable of sustained thermal erosion smoothing the pre-mare surface cannot be ruled out. While much remains to be done to elucidate the nature and cause of the sequence of mare flooding on the nearside, particularly the earliest outpourings that are hidden from view, they clearly feature prominently in the magmatic, tectonic, and geodynamic history of the Moon.

3.7 Supplementary Material

3.7.1 Spherical harmonic localization window testing

To compare the spectral signal between the nearside mare region and the model of a mare-flooded farside, localized power spectral analysis (Wieczorek & Simons, 2005) was conducted to extract spectral information from designated regions on the Moon. The bandwidths
used to generate the localized power spectra were chosen to maximize the spatial concentration and to minimize the loss of signal at the start and end of the spectral range. All background areas were localized using the same bandwidths and eigenvalue thresholds as the anomalies with which they are compared. However, due to the differences in the area and shape of the regions of interest, the number of tapers that satisfied the eigenvalue requirement ranged from sixteen to twenty-two. As a result, the difference in the localization window shapes as well as the number of tapers in the power spectral analyses of the anomalies compared to those of the background areas may produce a bias in the results. We test for the existence of the bias in the following text.

The presence of the bias was tested by comparing the Bouguer gravity power spectra of the Moon between two windows of the same shape, the N1 mask (see Fig. 3.1 in main paper) and the N1 mask translated to be approximately within the boundaries of the F1 mask (Fig. 3.S1a), which is defined as the F3 mask. While the spatial area of the N1 mask is not perfectly preserved due to latitudinal translation, pure longitude translations of the N1 mask would result in either the significant overlap of the mask with a major lunar basin, or the majority of the translated mask would be outside of the boundaries of the F1 mask. The localized power spectrum of the F3 mask utilized the same gravity map as that of the F1 mask, which is the Bouguer gravity from a model of the mare-flooded farside (Section 2.1). The resulting comparison (see Fig 3.S1b) shows no systematic bias in power from using localization windows of different shapes, but do show the inherent variance associated with sampling a field using localized power spectral analysis.
Figure 3.S1. (a) Topography of the Moon. The non-rectangular masks correspond to the areas used for the power spectra localization analyses (Section 2.2) for the lunar farside (F3) and nearside (N1). The mare-covered regions are highlighted by the black outline (Hiesinger et al. 2011). (b) Localized Bouguer gravity power spectral comparisons of the nearside mare region and the flooded farside models localized using the N1 mask for the nearside, and the F1 and F3 masks for the farside.

3.7.2 Thermal Annealing

3.7.2.1 Methods

In this study, we quantify the deficit in craters in both the spatial domain using crater size-frequency distributions (CFSDs) and in the spectral domain using Bouguer gravity power spectra, and investigate different candidate processes that may have produced the crater deficit. The fifth of the possible mechanisms behind the apparent crater erasure of the nearside mare is viscous closure of the pore space in the crust beneath the maria, in which the overlaying lava has heated the crustal material beneath past the thermal annealing, reducing the porosity of
lunar regolith present (Besserer et al. 2014) from ~20% to 0%. This would have the effect of increasing the density of the substrate material and reducing the density contrast between the mare and the substrate, and reduce the gravitational signal of buried craters. Heat flow is diffusional, and thus the annealing front is expected to follow a diffused representation of surface topography in our models. In this scenario, the flooded farside gravity would have a contribution directly from the flooded topography at the mare-crust interface with a density contrast of ~300 kg/m$^3$ (the density of the mare, less the grain density of lunar highland materials at ~2800 kg/m$^3$; Kiefer et al. 2012), and a contribution from the more diffused annealing front from depths beneath the mare-crust interface with a density contrast of about 340 kg/m$^3$ (assuming a 12% porosity and a grain density of 2800 kg/m$^3$).

Previous studies showed that thermal annealing of the pore space can be represented as occurring at a distinct temperature that depends on the crustal rheology (~650°C for the lunar plagioclase-rich crust; Wieczorek et al., 2013). Although previous studies showed thermal annealing of pore space around intrusions is expected to be minimal (Head & Wilson 2019), we include this in the scenarios tested here for completeness’s sake. The thermal diffusion equation was used to track the depth of the annealing front with time:

$$\frac{\partial T}{\partial z} = \frac{k}{\rho c_p} \nabla^2 T(r, z)$$

(1)

Applying this model to the lunar farside would be computationally expensive, so as a small-scale test we apply the model to the initial crater topography of Tycho (Smith et al., 2010; Wieczorek, 2015). With a diameter of 80 km, Tycho is at the threshold at which crater gravity
anomalies are beginning to be removed, and is thus a useful test case. Diffusion is modeled in an axisymmetric geometry, due to the inherent approximate axisymmetric nature of craters. The specific heat and density used were 1.1 kJ/(kg K) (Robertson, 1988) and 2800 kg/m$^3$ (Kiefer et al., 2012), but the thermal conductivity was assumed to be porosity- (Warren & Rasmussen 1987, Besserer et al. 2014) and temperature-dependent (Warren & Rasmussen 1987). The porosity is set to zero once the temperature for thermal annealing is crossed. The feedback between thermal annealing and thermal conductivity could accentuate the downward propagation of the annealing front by trapping heat at shallow depths. The initial conditions are that the entire crater is covered by low-TiO$_2$ basaltic lava at its solidus temperature (1150°C, Williams et al., 2000) extending up to 1 km above the crater rim. In addition, the boundary conditions are that the surface of the lava is set to the ambient temperature of 250 K, and that there is a constant basal heat flux of 55 mW/m$^2$, which is an estimate of early lunar heat flux at ~3.6 Gyr (Laneuville & Wieczorek 2013). The model is run with a horizontal and vertical resolution of 50 m and is run until the temperature front required for thermal annealing, 650°C, reaches its maximum depth. From the shape of the annealing front, we assess whether sufficient smoothing occurs relative to the surface topography to reduce the gravity signatures.

3.7.2.2 Results

In the thermal annealing model (Fig. 3.S2), the model was run for 700,000 years in total, with the cooling lava heat gradually dissipating throughout the crust and the maximum temperature of the interior portion of the lava is reduced to 1009 K (from 1432 K). While the temperature of the crust beneath the crater was raised to upwards of 800 K, the maximum depth of thermal annealing (corresponding to a temperature of 923 K) was only 200 m. At 200 m penetrating depth,
the diffusive effects of the annealing front relative to the flooded topography are minor, and the annealing front closely follows the surface topography. A 200 m layer of intermediate density between mare and substrate would be insufficient to account for the reduction in gravitational signal. The thermal annealing front failed to reach greater depths because the average of the lava solidus temperature (1432 K) and the ambient temperature (250 K) of 836 K is less than the required annealing temperature (923 K), meaning the annealing front can propagate no deeper than a depth equal to the mare thickness. Cooling at the surface reduces this further. Given that this effect is negligible for this test case, we do not consider it further and do not model the effect on the flooded farside gravity.

**Figure 3.S2.** Temperature vs depth results of the thermal annealing model on the topography of modern-day Tycho Crater for (left) 0 years (right) 350,000 years. The total maximum annealing front depth is shown as the red outline.
3.7.3 Gravity of the exposed Imbrium ring

To explain the missing Imbrium basin rings, we consider the same scenarios used to explain the small crater deficit, such as a reduced density contrast between the mare and the rings. However, the reduced density contrast between the mare and crust as a whole is not considered as the strong gravity gradient signature where the Imbrium rings are exposed and visibly embayed by maria requires a large density contrast between the exposed rings and the embaying mare. Furthermore, the expression of the exposed rings in the Bouguer gravity is minimized for an assumed crustal density of \( \sim 2550 \, \text{kg/m}^3 \), as the absolute Bouguer gravity gradient of the line profile across the exposed Imbrium ring is minimized at a Bouguer correction density of \( \sim 2550 \, \text{kg/m}^3 \) (Fig. 3.3). This result indicates that the crust making up the rings is similar in density to the lunar highlands.
Figure 3.S3. (a) Topography of the southeastern region of Imbrium, with the line corresponding to the topography (b) and gravity (c) profiles highlighted. (b) Topography profile of the line spanning across the exposed Imbrium ring. (c) Absolute values of the Bouguer gravity gradient of the line profile across the exposed Imbrium ring using four densities for Bouguer correction.
3.8 References


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Chapter 4: Vestiges of a lunar ilmenite layer revealed by GRAIL gravity data

This chapter has been submitted to Nature Geoscience and is currently under review. That manuscript has been reformatted here, including the placement of the Methods within the main text rather than at the end of the manuscript.

4.1 Abstract

The crust and mantle of the Moon are thought to have formed through the crystallization of a magma ocean following the accretion of the Moon. The final stage of crystallization results in a solid cumulate mantle which is gravitationally unstable, with a layer of dense ilmenite-bearing cumulates (IBCs) containing incompatible-element-rich materials such as KREEP (potassium, rare earth elements and phosphorus) forming above the less dense olivine orthopyroxene layers. This configuration is thought to have resulted in a planetary-scale mantle overturn and the downwelling of IBC material into the core-mantle boundary that may have led to the observed nearside-farside asymmetry in titanium and KREEP abundance, though there is a lack of identified physical evidence for the overturn. Here, we propose that the observed quasi-rectangular configuration of gravity anomalies that surround the Procellarum KREEP Terrane (PKT) are the IBC material within past zones of downwelling that had remained after last stages of global mantle overturn. We show that the observed pattern, magnitude, and modeled dimensions of the PKT anomalies are consistent with those predicted for the IBC downwelling remnants. These gravity anomalies provide the first physical evidence for the nature of the global
magma overturn during the final stages, revealing the present-day distribution of ilmenite-rich material in the subsurface and providing constraints on the timing of mantle overturn.

4.2 Introduction

The Procellarum KREEP Terrane (PKT) is a region in the lunar nearside characterized by high abundances of potassium (K), rare-earth elements (REE), which include the heat producing elements such as thorium and uranium, and phosphorus (P). The PKT region strongly overlaps with the nearside lunar mare region and represents the most volcanically active (Head, 1975; Hiesinger et al., 2011), and perhaps tectonically active (Andrews-Hanna et al., 2014), region on the Moon (Fig. 4.1). It has been proposed that the voluminous volcanic activity is a natural result of the high abundances of heat producing elements present in the region (Wieczorek and Phillips, 2000). This region also hosts an abundance of ilmenite-rich basalts (Lucey et al., 1994; Sato et al., 2017). The origin of both the KREEP-rich and ilmenite-rich materials has been linked to the evolution of the Moon following the initial global magma ocean state (Elkins-Tanton et al., 2011; Hess and Parmentier, 1995) after accretion. Models of the crystallization of the magma ocean predict late-stage crystallization of increasingly dense oxides, culminating in a global layer of dense ilmenite (FeTiO3) -bearing cumulates (IBCs), with the incompatible elements (Elkins-Tanton et al., 2011), including the major heat producing elements, concentrated in the final liquids to create the reservoir from which KREEP-rich materials are sourced (sometimes referred to as urKREEP). The global IBC layer is predicted to sink into the interior due high density contrast (~300 kg/m³; ref. 6) relative to the underlying mantle cumulates. This material may have settled at the core-mantle boundary (CMB), prior to warming and buoyantly rising (Zhong et al., 2000). Such mantle overturn is consistent with the source depths of the Ti-rich glasses (Elkins Tanton et
al., 2002), where IBC material would initially sink and then rise back through the mantle to produce the Ti-rich mare basalts (Zhang et al., 2013; Zhong et al., 2000).

In their simplest form, magma ocean crystallization models predict a globally symmetric layer of IBCs and KREEP-rich material, which is in clear contrast to the currently asymmetric distribution (Prettyman et al., 2006; Sato et al., 2017) with a high concentrations of IBCs and KREEP in the PKT and nearside mare region. A number of models have been proposed to explain this observed asymmetric distribution, including the effects of giant impacts on the nearside (Zhu et al., 2019) or farside (Jutzi and Asphaug, 2011), and migration via a degree-one Rayleigh-Taylor instability (Parmentier et al., 2002), possibly aided by the spreading thermal anomaly of the SPA impact (Jones et al., 2022; Zhang et al., 2022). However, direct physical evidence for the nature of this process has been lacking. Today, the origin of the compositional asymmetry in both titanium and KREEP-rich material remains one of the most significant unresolved questions in lunar science.

One proposed scenario to form the asymmetry is the migration of the initially global KREEP- and Ti-rich IBC layer in response to the spreading thermal anomaly from the South Pole-Aitken impact (Jones et al., 2022; Zhang et al., 2022). Following migration of this material to the nearside, this model predicts the ensuing Rayleigh-Taylor instability to take the form of a network of sheet-like downwellings of IBC material (Zhang et al., 2022). After the overturn, remnants of the IBC material are predicted to remain at the crust-mantle boundary at the sites of downwelling, resulting in intersecting, narrow, linear zones of IBC-rich material in a polygonal configuration (Zhang et al., 2022).
Previous analyses of data from the Gravity Recovery and Interior Laboratory (GRAIL) mission (Zuber et al., 2013) to the Moon revealed a polygonal pattern of linear to arcuate anomalies on the near side (Andrews-Hanna et al., 2014) surrounding the PKT region (Fig. 4.1c). Interestingly, the gravity signatures of the PKT border anomalies differ from those of linear anomalies observed elsewhere (Liang & Andrews-Hanna 2020), in that the former are dominated by longer wavelengths and the latter by shorter wavelengths, implying a different origin and formation history. While the anomalies were initially interpreted as volcanically flooded rift zones possibly formed in response to the cooling of the region (Andrews-Hanna et al., 2014), their nature, origin, and significance remain highly uncertain. This observed polygonal pattern of gravity anomalies is now seen to closely resemble the predicted polygonal structure of the remnants of the IBC-rich layer following the overturn of the mantle.
Figure 4.1. (a) Global lunar topography (Wieczorek, 2015) in kilometers, with the mare outlined, (b) surface TiO$_2$ weight percent abundance (Sato et al., 2017), (c) GRAIL Bouguer gravity gradient (in units of Eötvös; 1 E = 10$^{-9}$ s$^{-2}$) calculated using a surface density of 2550 kg/m$^3$. The anomalies used in the Monte Carlo analysis are highlighted in black and labeled with their designated identifiers in c.

In this study, we combine GRAIL gravity data together with models of the overturn of the ilmenite-rich materials to reveal the vestiges of the nearside ilmenite layer as a result of the early lunar mantle evolution. First, we show that the forward-modeled gravity of the predicted IBC distribution from the mantle overturn models is consistent with the gravity of the PKT anomalies as observed by the GRAIL mission. Next, we demonstrate that the cross-sectional geometry of the structures beneath the PKT anomalies are consistent with that of the predicted ilmenite downwelling residuals. Finally, we invert the gravity and topography data for the thickness and distribution of the IBC layer. The results of these analyses shed light on the nature and timing of the lunar global mantle overturn.

4.3 Data and Methods

4.3.1 Thermochemical models of the mantle overturn event

In previous work, the overturn of the post-magma ocean cumulate mantle was modeled by first using the iSALE-3D shock physics code (Elbeshausen et al., 2013) to simulate a variety
of impact diameters and angles, and then the evolution of the resulting thermal anomaly, as well as the initial IBC layer, is recorded in CitcomS (Zhong et al., 2008). The resulting interfaces between the IBC, crust, and the mantle after the termination of mantle overturn are returned as the modeled Ti-abundances. Full details of this modeling can be found in that work.

4.3.1.1 Comparison of Modeled Ti Abundance with GRAIL Gravity

The modeled IBC thickness is converted to gravity originating from the density interface between the IBC layer and the mantle. First, the IBC thickness is converted into the interface between IBC and mantle by subtracting the crustal thickness and the IBC thickness from the surface topography. The gravity from the interface was then modeled using a finite-amplitude algorithm (Wieczorek and Phillips, 1998) in the Spherical Harmonic Tools (SHTOOLS) software archive (Wieczorek and Meschede, 2018), using the mean planetary radius as the reference radius. The assumed density contrast between the IBC layer and mantle was 300 kg/m³. A point of note is that while the mantle overturn models (Zhang et al., 2022) assumed a lower density contrast of ~90 kg/m³ between the IBCs and surrounding mantle, a density of 300 kg/m³ would be a more appropriate value for the density difference between mantle and remnant IBC layers located at the crust-mantle interface. During the migration to the nearside and initial sinking of the IBCs, they would be underlain by the dense, late-stage, low Mg# (defined as the molar ratio of Mg/(Mg+Fe)) cumulates, with the specific density contrast depending on the mixing ratio between ilmenite and other minerals in the IBCs (Parmentier et al., 2002; Zhang et al., 2017). However, after overturn, less dense, high Mg# cumulates would lie at the top of the mantle, with densities ~200 kg/m³ less than the original upper mantle cumulates. Thus, the present-day density contrast between the IBCs and the underlying mantle should be substantially
greater than the density contrast prior to and during the overturn. The gravity is then converted into gravity gradients using SHTOOLS. Bouguer gravity gradients are used to highlight anomalous regions in the gravity data, and are calculated using the maximum amplitude eigenvalue of the tensor of the second horizontal derivatives of the potential (referred to as the gravity gradients (Andrews-Hanna et al., 2013)).

4.3.2 Monte Carlo Markov Chain Algorithm

Next, having established the correlation between the modeled gravity after the mantle overturn and what is observed in GRAIL gravity, we use the gravity data directly to constrain the geometry of the remnant IBC material after mantle overturn. As in the previous work (Liang and Andrews-Hanna, 2022), we use a Monte Carlo Markov Chain (MCMC) model to constrain the properties of the subsurface structure. Our approach differs from the previous work in that we are using the spatial profiles of the gravity anomalies for inversion rather than the power spectrum. Both approaches are valid, but they emphasize different aspects of the data. Power spectral approaches are expected to be more sensitive to the shallow structure that will dominate the higher degrees, whereas the PKT border anomalies are expected to be below the crust-mantle interface and have dominantly low degree power (Liang and Andrews-Hanna, 2022). In addition, the tradeoff between bandwidth and concentration as part of the localization process would leave very small spectral windows for localized power spectral analysis. However, in the absence of noise or spatial variability in the structure, if a unique solution exists, both methods should arrive at the same conclusion.

The spatial profile used for the MCMC modeling was sourced from the spherical harmonic degree 1200 GRAIL GRGM1200A free-air gravity model (Lemoine et al., 2014; Zuber et al.,
2013) and a spherical harmonic degree 2600 LOLA topography model (Smith et al., 2010; Wieczorek, 2015). A Bouguer correction density of 3100 kg/m³ is also appropriate as a representation of the density of the mare layer. To minimize the effects of long-wavelength surface topography on the gravity data, including the nearside-farside asymmetry, we calculated the gravity and gravity gradients on a smoothed representation of the surface topography rather than on a spherical surface (Andrews-Hanna et al., 2018). The ‘gravity on the surface’ algorithm calculated the potential at the local surface elevation at every point, using a spherical harmonic degree 30 representation of the surface topography to avoid abrupt jumps in elevation, and the gravity anomaly and gravity gradients were calculated from this potential model. As the features of interest do not have expressions in topography and occur beneath the smooth maria, the resulting gravity is not strongly affected by the details of the Bouguer correction. In order to generate the profiles, the Moon was rotated using spherical harmonic transformations so that the corresponding anomaly is parallel to the equator. This permits us to use gravity profiles perpendicular to the anomaly in the latitudinal direction. The singular gravity profile used as the observed data is generated by the averaging of 60 and 68 profiles for the NE and S anomalies, respectively.

Modeled subsurface structures were represented as combinations of rectangular prisms, for which there is an analytic solution for the gravity anomaly (Nagy, 1966). The subsurface geometry we considered is the triangular prism (Fig. 4.2), which is consistent with the geometry of the remnant IBC layers as predicted by the mantle overturn models. The triangular density anomalies consisted of 1000 rectangular prisms of constant lengths (parallel to the strike of the
anomaly) and heights (vertical dimension), but with varying widths (horizontal dimension perpendicular to the strike of the anomaly) to provide a triangular cross-section.

![Diagram of triangular prism](image)

**Figure 4.2.** A visual representation of how the triangular geometry is discretized in the model. The actual model discretization was much finer than depicted here (see text for details).

To find the best-fit geometry of the density anomalies, we used the Metropolis-Hastings Markov chain Monte Carlo algorithm (Chib and Greenberg, 1995). The triangular prism was characterized by the top depth, the bottom depth, and the width of the anomaly, from which the thickness can be derived using the difference between the top and bottom depths. The *a priori* widths were allowed to vary on a uniform distribution between 50 and 200 km, the top depths varied from 20 to 50 km, and the bottom depths varied from the top depth 100 km below the top depth. We set a constraint on the maximum bottom depth to 150 km, which corresponds to the depth in which gravity from a spherical harmonic degree $l=30$ feature loses 90% of its signal.
Models were evaluated using a likelihood function, $L$:

$$\quad L = e^{SSD/\sigma^2}$$

(2)

in which $SSD$ is defined as the summed squared difference of the modeled and observed gravity profile, and $\sigma$ is the standard deviation of a background area in a relatively featureless part of the nearside (57 km radius circle centered at 11.4 N, -27.1 E). Since the error in the GRAIL data is much lower than the background variability in the field (Jansen et al., 2017), the background variability $\sigma$ is used in place of the data error commonly used in such algorithms (Chib and Greenberg, 1995). The MCMC simulations were run using 10000 iterations over 5 parallel chains, and the resulting posterior distributions and root-mean squared error (RMS) comparisons are shown in Supplementary Figures 4.S1 and 4.S2.

### 4.3.3 Gravity and Topography Inversion

The MCMC models provide a constraint on the top depths of the anomalies, placing them near the base of the crust. This justifies the use of full spherical harmonic inversions of the gravity and topography data for the thickness of the IBC layer under the assumption that it is below the crust. Our algorithm of gravity and topography inversion is based on previous work (Broquet and Andrews-Hanna, 2023) which uses a two-step two-layer loading model to constrain lunar mare thickness, and adds a third step to constrain the thickness of the IBC layers.

The two-step method uses two separate inversions of a systems of 5 equations, with 8 unknowns as updated from ref. (Banerdt, 1986). This approach makes use of a thin-shell model
that partitions the crust into a mare top load, a feldspathic top and bottom load, and computes lithospheric displacement as a function of the elastic thickness of the lithosphere. The first inversion uses the observed topography ($H$), the observed geoid ($G$), and the assumption of isostasy in the pre-mare crust at long wavelengths (< spherical harmonic degree 90) Isostasy between pre-mare topography and the pre-mare crust is a reasonable assumption as the pre-mare crust likely solidified from a global magma ocean (Elkins-Tanton et al., 2011), and thus the pre-mare surface would be the isostatic ratio with the bottom loads in the crust.

The resulting mare thickness map from the first step is then thresholded to a minimum value of zero and then clipped to the observed mare and cryptomare distribution (Nelson et al., 2014; Whitten and Head, 2015). The updated mare thickness map is then input as a known load into a second inversion that exactly fit gravity and topography to obtain self-consistent crust and mare thickness maps matching the expectation of an approximately isostatic pre-mare crust.

The PKT anomalies show stronger Bouguer gravity signal in comparison to the surrounding nearside mare. The additional, third inversion considers the additional gravity signal of the PKT border anomalies to originate from IBC materials in the mantle. First, a mask corresponding to the spatial locations of the PKT anomalies is created. Next, the crustal and mare thickness values generated by the second step within the mask are removed and replaced by a surface interpolated from data in the surroundings. The interpolation used the minimum curvature approach of ref. (Smith and Wessel, 1990). The interpolated crustal and mare thicknesses are then input into the inversion in order to solve for the hypothetical gravity map without the PKT anomalies. Then, the difference in gravity signal between the observed and
modeled is inverted to generate the thickness of the IBC layers. The result is a model of the thicknesses of the crust, mare, and IBC layer that reproduced gravity and topography, follows the expectation of an isostatic pre-mare crust, and attributes the gravity excesses in the PKT border anomalies to variations in the thickness of an underlying IBC layer.

4.4 Results

4.4.1. Predicted gravity following the mantle overturn

The nearside Ti-abundances as well as the IBC thickness predicted by the mantle overturn model of Zhang et al. 2022 are used to forward model the gravitational signal of the IBC layers (Methods). The distribution of IBCs below the crust after global mantle overturn in that model are stable and may be present below the crust today. That study assumed a density contrast of ~88 kg/m$^3$ between the IBCs and the dense upper mantle cumulates prior to mantle overturn. After the overturn, lower density cumulates would have risen to the uppermost mantle, increasing the density contrast with respect to the residual IBCs. We forward model the resulting Bouguer gravity anomalies using the modeled Ti-abundances assuming that the residual IBC material has a density of 3700 kg/m$^3$, relative to the surrounding mantle density of 3400 kg/m$^3$. The Bouguer gravity gradients are used to highlight small-scale anomalies in the gravity data, and were calculated using the maximum amplitude eigenvalue of the tensor of the second horizontal derivatives of the potential (hereafter referred to as the gravity gradients (Andrews-Hanna et al., 2013)).
The forward modeling reveals a polygonal configuration of Bouguer gravity gradient anomalies in the lunar nearside (Fig. 4.3a-c), which resembles the currently observed gravity anomalies surrounding the PKT region. The maximum amplitudes of the PKT anomalies and the forward modeled IBC layers are consistent to the first order, with the PKT anomalies have a peak magnitude of roughly -30 E, and the forward modeled IBC anomalies have peak magnitudes of roughly -15 E. The widths of the PKT anomalies and the modeled IBC layers are consistent to the first order as well, with the PKT anomalies having widths of 125 ± 66 km, and the modeled IBC layers having widths of 70 ± 23 km. The overturn models predict an inverse dependence of the lengths of the individual linear anomalies as well as the sizes of the polygonal shapes on the mantle-IBC viscosity contrast. A viscosity contrast of $10^{-3}$ between the IBCs and the mantle results in polygonal patterns of anomalies with a mean and standard deviation of 1253 ± 258 km edge lengths, while viscosity contrasts of $10^{-2}$ and $10^{-1}$ results in polygonal patterns of anomalies with 789 ± 154 km and 589 ± 172 km edge lengths, respectively (Fig 2). In comparison, the PKT border anomalies, when measured as individual linear anomalies, have lengths of 1246 ± 435 km, favoring lower viscosities for the IBCs. However, the models predict a more uniform polygonal network of anomalies traversing through the interior of the region, while the observed gravity field is dominated by a single quasi-rectangular structure of linear anomalies around the edges of the region. Nevertheless, the interior of the PKT region does exhibit a number of less prominent quasi-linear gravity anomalies, and it is also possible that some of the original structure has been modified by subsequent impact basins (see discussion below).
Figure 4.3. Forward modeled Bouguer gravity gradient maps from Ti-abundance in the mantle overturn models assuming viscosity contrasts of $10^{-3}$ (a), $10^{-2}$ (b), and $10^{-1}$ (c) between the mantle and the IBC layers.
4.4.2. Dimensions of the residual IBC bodies constrained by gravity

The physical geometry of the residual IBC bodies is investigated by comparing the observed Bouguer gravity profiles to a forward model in a Monte Carlo Markov Chain (MCMC) framework (Methods). The MCMC analysis represents the density anomalies in an inverted triangular prism geometry, for which the top depth, width, and thickness are the free parameters. Observed average Bouguer gravity profiles of the two largest segments of the PKT border anomalies, designated as “NE” and “S” (Fig. 4.1a), are inverted using the MCMC algorithm. The resulting top depths (Figure 4.4, Table 4.1) for the anomalies (41 ± 13 km and 39 ± 16 km, respectively) are consistent with the IBC remnants being located just below the base of the crust (~35 km, ref. (Wieczorek et al., 2013)), and providing justification for our subsequent topography and gravity inversion analysis. The modeled thicknesses of the ‘NE’ and ‘S’ PKT border anomalies (69 ± 20 km and 59 ± 22 km, respectively) are to first order consistent with the thicknesses of the IBC layers in the mantle overturn models (27 ± 12 km), but the thicknesses returned by the MCMC models may support larger and deeper IBC remnants or higher viscosity contrasts than predicted. However, the models allow a moderate range in widths and thicknesses at the 1-σ level, which reflects the tradeoffs between width and height, as well as the lower sensitivity of gravity for anomalies arising at significant depths. Overall, the modeled top depths and thicknesses are consistent with a high-density source located beneath the crust, which is also supported by a lack of high degree signature (Liang and Andrews-Hanna, 2022) from the anomalies.
Figure 4.4. Bouguer gravity profile comparisons of MCMC model runs for the NE and S PKT anomalies.

<table>
<thead>
<tr>
<th>Anomaly</th>
<th>Top Depth (km)</th>
<th>Thickness (km)</th>
<th>Width (km)</th>
<th>RMS (mGal)</th>
</tr>
</thead>
<tbody>
<tr>
<td>NE</td>
<td>48.0 (41+13,-12)</td>
<td>84.5 (69+20,-21)</td>
<td>169 (201+49,-51)</td>
<td>8.26</td>
</tr>
<tr>
<td>S</td>
<td>33.0 (39+15,-16)</td>
<td>44.9 (59+22,-23)</td>
<td>212 (194+62,-59)</td>
<td>6.78</td>
</tr>
</tbody>
</table>

Table 4.1. Results from the MCMC model runs for the PKT border anomalies.

Heights and widths are given as the best-fit (top number), and mean with 1-σ range (bottom numbers).
4.4.3 Distribution of the residual IBC bodies constrained by gravity

The MCMC models support density anomalies at the base of the crust as the source of the observed gravity anomalies. To provide an additional constraint on the distribution and thicknesses of the IBC layers, we perform a series of inversions on observed lunar topography and free-air gravity. We invert the gravity and topography for the thickness of the crust, mare, and IBC-rich layer using a three-step approach. The first two steps solve for the thickness of the mare and crust following our previous work (Broquet and Andrews-Hanna, 2023), based on the assumption of long-wavelength pre-mare isostasy in the nearside crust. In the third step, we assume that the narrow gravity anomalies of the PKT border anomalies arise from variations in the thickness of an IBC layer at the base of the crust. We interpolate the crust and mare thickness to be uniform across the border anomalies, and use these together with the gravity and topography to solve for the thickness distribution of the IBC layer (Methods).

As with the comparison of the observed gravity to the mantle overturn models, the overall pattern of residual IBC bodies from the gravity inversion is broadly consistent with the models. The observed topography and gravity inversions return a range of IBC thicknesses with a mean and 1-σ range of 29 ± 14 km, and for the NE and S border anomalies in particular, these inversions yielded IBC thickness values of ~35 km, which is consistent with those predicted by the mantle overturn models and our MCMC calculations (Fig. 4.5). In addition, our topography and gravity inversions show that the regions of highest IBC layer thickness are located in the corners of the polygonal features, which is consistent with those predicted in the models, but the layer thicknesses are higher than those predicted by the mantle overturn models. The areas of
highest IBC thicknesses located in the northwestern and northeastern corners of the polygonal anomaly structure, as well as in the southeastern anomaly, whose gravitational signal may have included that of the Serenitatis Basin. We also note that the predicted thicknesses from the overturn models for the higher viscosity contrast \((10^3)\) better match the thicknesses constrained from gravity, consistent with the above comparisons of anomaly length.

There is not a direct correlation between either the PKT border anomalies as revealed by gravity or the predicted thickness of the IBC layer and the distribution of ilmenite-rich maria on the surface. However, such a correlation is not expected. The ilmenite-rich basalts are thought to have formed from the sinking of the IBCs into the deep lunar interior, followed by their warming and buoyant rise leading to decompression melting and mare volcanism (Zhang et al., 2013; Zhong et al., 2000). Thus, only a regional correspondence between the two is expected.

Figure 4.5. (left) IBC thickness of the PKT anomalies as calculated from inversions of observed gravity and topography. (right) Averaged profiles of IBC thickness across the NE and S
anomalies in the areas used in the MCMC model, as well as the averaged relative profiles of IBC thickness in comparison to the background area in the overturn model using $10^{-3}$, $10^{-2}$, $10^{-1}$ viscosity contrasts between mantle and IBC.

4.5 Implications for lunar mantle evolution and basin chronology

The PKT anomalies and their origins as IBC remnants provide critical insight into the chronology of early lunar evolution and impact cratering in comparison to the timing of the global magma overturn. The Humorium, Serenitatis, and Asperitatis basins are seen to closely interact with the PKT anomalies. The S anomaly overwhelms the signal ejecta on the eastern edge of Humorium in the gravity gradients. However, the anomaly ends abruptly at the edge of the thinned crust in the center of the basin with no signal indicative of its presence there, which implies that the Humorium forming impact postdated the overturn and either excavated a significant portion of the anomaly or significantly disrupted it through the intense deformation during the excavation and collapse of the transient cavity. This inference is consistent with a predicted Humorium excavation depth of ~45 km (ref. Miljković et al., 2013). In contrast, models of the overturn of this dense layer do not predict the pattern of the residual IBC material to be at all affected by pre-existing basins.

On the other hand, while the gravitational signal of the PKT border anomalies significantly modify the signals of the ejecta of the Serenitatis and Asperitatis basins, the basin centers do not as obviously intersect with the projected trajectories of the PKT anomalies. Nevertheless, our reconstruction of the path of the eastern border anomaly does cross the edge of
the uplifted mantle plug of Serenitatis, and the further south would continue across the edge of
the Asperitatis basin (Fig. 4.S3). As with Humorum, neither basin shows any sign of the
gravitational signature of the PKT border anomalies within the thin crust in the basin centers.
Thus, we confidently conclude that the Humorum basin postdates the border anomalies, and our
preferred interpretation is that both the Serenitatis and Asperitatis basins also postdate the PKT
anomalies. Based on recent estimates of the ages of these basins (Orgel et al., 2018), this would
constrain the formation of the PKT border anomalies and the global IBC overturn to be before
the Humorum impact (4.09 Ga), and likely before the the Serenitatis (4.22 Ga) and Asperitatis
(>4.17 Ga) impacts, and may favor an age for SPA older than the suggested age of 4.25 Ga
(Garrick-Bethell et al., 2020).

The most direct constraint on chronology comes from the ilmenite-rich basalts. The
migration of the IBC-rich material to the nearside is predicted to take ~100-300 Myr (Jones et
al., 2022; Parmentier et al., 2002; Zhang et al., 2022). Following the sinking of the IBCs and any
entrained KREEP-rich material, models predict a delay of 300-400 Myr before the dense
cumulates are sufficiently warmed to become positively buoyant to rise back to the upper mantle
to generate the KREEP-rich basalts (Zhang et al., 2017; Zhong et al., 2000), consistent with the
ages of the oldest Ti-rich basalts in Mare Tranquilitatis of 3.7-3.8 Ga (Hiesinger et al., 2011).

The PKT border gravity anomalies may provide the most direct physical evidence for the
nature of the overturn of the post-magma ocean cumulate mantle and sinking of ilmenite into the
deep interior. The observed gravity anomalies reveal the present-day locations of the vestiges of
the final stages of lunar mantle overturn, which is dominated by a large polygonal structure, with
additional interior anomalies that may be partially obscured by the complicated signal within the PKT. The location of these anomalies only within the central nearside confirms previous suppositions that the dense late-stage cumulates migrated to the nearside prior to sinking into the interior (Parmentier et al., 2002; Zhang et al., 2017). The gravity anomalies confirm the prediction that the final sinking of these dense materials took the form of sheet-like downwellings (Zhang et al., 2022). The results of the mantle overturn models are dependent on the physical properties of the mantle and IBC layer, and thus the PKT border anomalies can be used to constrain the viscosity contrast between the ancient mantle and the IBCs (with a preferred value of $10^3$), as well as the density of the IBCs (consistent with an assumed value of 3700 kg/m$^3$). Thus, the lunar gravity field has preserved a critical record of the overturn of the lunar mantle that has been widely postulated to be one of the defining events in early lunar history, but whose details have until now remained unknown.
4.6 Supplementary Material

**Figure 4.S1.** Root-mean squared error (RMS) plotted in comparison to top depth, thickness, and width of the NE (a-c) and S (d-f) anomalies in the MCMC runs.
Figure 4.S2. Posterior distributions of top depth, thickness, and width of the NE (a-c) and S (d-f) anomalies in the MCMC runs.
**Figure 4.S3.** Projected paths of the PKT anomalies across the Serenitatis and Asperitatis basins shown on the (a) Bouguer gravity, (b) Bouguer gravity contour, and (c) Bouguer gravity gradient. These maps are generated using a low-pass at spherical harmonic degree 150, and a Bouguer correction density of 3100 kg/m³.

### 4.7 References

https://doi.org/10.1126/science.1231753

https://doi.org/10.1038/nature13697


https://doi.org/10.1029/JB091iB01p00403

https://doi.org/10.1038/s41550-022-01836-3


https://doi.org/10.1002/2013JE004477


Zuber, M.T., Smith, D.E., Watkins, M.M., Asmar, S.W., Konopliv, A.S., Lemoine, F.G.,
Field of the Moon from the Gravity Recovery and Interior Laboratory (GRAIL) Mission.
Chapter 5: Conclusions

The Gravity Recovery and Interior Laboratory (GRAIL) mission has uncovered a series of scattered linear and arcuate gravity anomalies in the lunar near and far sides, as well as anomalies in a quasi-rectangular structure that border the PKT (hereafter referred to as PKT border anomalies). In Chapter 2, my band-passed maps and power spectral comparisons showed that the linear anomalies have significant power at the higher spherical harmonic degrees (from 40 to 150), while in comparison the power of the PKT anomalies is predominantly at the lower degrees (below 60). These results imply that the linear anomalies are likely shallower and less wide features in comparison to the PKT anomalies.

The interpretation that is most consistent with the linear anomalies is dike intrusions into the farside crust. Dike intrusions are common on the Earth, but are of scales in ones to hundreds of meters in width (Ernst et al., 2001), but the MCMC results of the linear anomalies indicate features of kilometers in width. These features are unprecedented in terms of scale, and provide additional evidence for past farside magmatism, but the calculated volume of the intrusive bodies is orders of magnitude lower than that of the mare plains on the nearside, indicating a fundamental difference in volcanic activity between the two regions of the Moon.

Chapter 2 also shows that there is a fundamental difference in origin between the linear anomalies and the PKT anomalies, and thus an origin other than intruded dikes must be considered for the PKT anomalies. As discussed in the introduction in Chapter 1, lunar global mantle overturn has been proposed to address numerous fundamental questions in lunar science, including the asymmetry in radioactive element distribution between the near and far sides of the Moon, as well as the abundant high-Ti basalts in the nearside mare. The global mantle overturn...
would involve the overturning of ilmenite-bearing cumulates (IBCs), which are thought to have formed a global layer at the top of the mantle at the end of magma ocean crystallization that would have been gravitationally unstable in comparison to the mantle. The global mantle overturn is proposed to take the form of a migration of IBCs to the nearside followed by their sinking to the core-mantle boundary, resulting in the asymmetry of radioactive element abundance and high-Ti flows.

Recent studies (Jones et al., 2022; Zhang et al., 2022) have modeled in detail the global migration and sinking of IBC material. These models show that overturn took the form of sheet-like downwellings in a polygonal patterns in map view, which bears a striking resemblance to the quasi-rectangular anomaly that surrounds the PKT region. In Chapter 4, I investigated the hypothesis that the PKT anomalies are remnants of the IBC downwellings that remained after the last stages of global mantle overturn. First, I forward modelled the gravity of the remnant polygonal configurations of IBC layers predicted in the mantle overturn model, and the resulting magnitude, configurations, and dimensions of the polygonal edges are consistent with those of the PKT anomalies. Next, I used Monte Carlo methods and topography and gravity inversions to calculate the top depths and dimensions of the IBC structures. The resulting top depths are consistent with structures beneath the crust-mantle boundary, and the resulting dimensions are consistent those predicted in the mantle overturn models. With this new interpretation, the PKT anomalies serve as the first physical evidence of the global mantle overturn, which currently is used to address multiple fundamental questions about the Moon.

We observe that there are multiple basins that have cross cutting relationships with the PKT anomalies. Specifically, the southern edge of the PKT anomaly is projected to intersect with
the center of the Humorum mascon, and overwrites the signal of the eastern edge of the ejecta in the gravity gradients. The projected paths of the PKT anomalies are in contact with the edges of the Serenitatis and Asperitatis basins as well, but do not as clearly intersect with the basin centers. However, gravitational signal resulting from the remnant IBCs is not apparent in the center of any basin. This observation implies that Humorum post dates the PKT anomalies, and the same is likely true for Serenitatis and Asperitatis as well, constraining the time of the sinking of the IBC rich material to be >4.22 Ga.

In Chapter 3, I investigated the different mechanisms that could have resulted in the missing gravitational signatures of craters <90 km in diameter where buried beneath the maria, as well as the missing rings of Imbrium. While the pre-mare dense layer model was able to explain the missing craters, it could not have explained the missing rings of Imbrium as the exposed Imbrium have densities close to the highland material. Thus, the explanation most consistent with observations is some manner of surface smoothing during or prior to the mare eruptions which we represent as diffusion, presumably through thermochemical erosion of the substrate due to hot overflowing lava. Some thermal erosion is likely to have occurred due to clear evidence of past mare volcanism and flooding. This result has implications for the volume, temperature, and duration of the mare flooding as mare flows of unprecedented high volume, temperature, and duration are needed in order to erode such features.

In all, my work was able to elucidate several outstanding areas in lunar evolution. First, my investigations in Chapter 4 provide the first physical evidence for the global mantle overturn, which has been used to address numerous fundamental observations of the Moon. Next, my work in Chapter 3 proposes that thermal erosion is the mechanism that was responsible for the missing
Imbrium rings, which are estimated to be of 100s of km in scale, as well as the deficit in nearside buried craters, and as a result places a constraint on the potential of thermal erosion of the nearside mare floodings. Finally, my work in Chapter 2 has presented new evidence on significant early farside magmatic activity and constraints on farside volcanism in comparison to those on the nearside. I have hopes that the methods and interpretations in this study may be used in the future when gravity data of similar precision is achieved in other solid bodies in the Solar System, and I sincerely look forward to the time when I would be able to view those gravity maps. In conclusion, the work presented in this thesis has revealed new aspects of numerous areas in early lunar history, such as early farside volcanism, the nearside mare floodings, as well as the global mantle overturn.

References


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