SMALL SCALE GRAVITY ANOMALIES OBSERVED IN GRAIL GRAVITY DATA

by

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ABSTRACT

The Gravity Recovery and Interior Laboratory (GRAIL) data has revealed an unparalleled high-resolution gravity map of the Moon. This high-resolution data enables us to look at small-scale (10’s of km) shallow gravity anomalies and attempt to interpret them in terms of density. In chapter 2 the background variability is investigated and explained by using gravity inversions to model the small-scale three-dimensional variations in the density of the lunar crust. In an attempt to explain the density variations, we interpret them in terms of three end-member scenarios of variations in porosity, intrusions into the crust, and variations in bulk crustal composition. Though it is most likely that all three end-member scenarios contribute to the background variability, we find that the density anomalies can be caused entirely by changes in porosity.

In chapter 3 we investigate small-scale gravity anomalies radiating out from the Orientale basin, shown in GRAIL data. These radial gravity lineations are sometimes associated with secondary crater chains or catenae. Here we use gravity inversions, hydrocode modeling, and observations to investigate the radial gravity anomalies in more detail. Density inversion models show that the gravity can be matched by solutions ranging from broad low amplitude anomalies to shallow high amplitude anomalies. Hydrocode models show that the impacting material remains as a thin layer within the secondary craters and this can explain the linear gravity anomalies. However, this does not explain the majority of linear gravity features that do not have secondary crater chains associated with them. Therefore, we conclude that the majority of radial gravity anomalies represent the structure of the ejecta blanket. This density variability within the ejecta of basins likely contributes substantially to the density variability of the shallow upper crust as a whole.
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1.1 GRAIL mission

The Gravity Recovery and Interior Laboratory (GRAIL) mission was modeled after the Earth equivalent Gravity Recovery and Climate Experiment (GRACE) mission which mapped the Earth’s gravity from 2002 through 2017. GRAIL consisted of two spacecraft (Ebb and Flow) in the same orbit around the Moon. As they flew over areas of greater and lesser gravity, the relative distance between the two spacecraft varied. An instrument aboard GRAIL measured the change in their relative distances by measuring the Doppler shift, and this data was sent to Earth. Here, teams at the Goddard Spaceflight Center and the Jet Propulsion Laboratory used super computers to translate this data into a spherical harmonics model that represents the gravity field of the Moon.

At the time of the mission launch on 10 September 2011, it was promised to obtain a spherical harmonic gravity field up to degree 180. By the end of the mission on 17 December 2012, it was clear that the gravity model could be expanded beyond degree 180. Today, they are pushing the limit to obtain a degree 1800 map. In this thesis, the degree 900 gravity is used. These high-resolution models reveal features in the gravity data that were not resolved before GRAIL. The higher resolution data also brings new challenges, such as the discovery of small-scale variability in the crust that may obscure the signature of small structures (Chapter 2).

The GRAIL data has resulted in many interesting discoveries, as it can be used in new and interesting ways. For example, Andrews-Hanna et al. (2013) used the high resolution of the gravity data to apply gravity gradiometry to the data. They used the second spatial derivatives of the gravitational potential to calculate the maximum gradients, resulting in a map that highlights
the locations where the Bouguer gravity had the highest gradient. From this work, they found a number of linear gravity anomalies interpreted to be ancient vertical tabular intrusions or dikes. These features are hundreds of kilometers long and of pre-Nectarian to Nectarian age (~4.5-3.8 Ga), indicating extensional tectonism occurring at that time. Another example of research done with GRAIL is a better estimate of the Love numbers (Lemoine et al., 2013; Williams et al., 2013; Lemoine et al, 2014). The Love numbers, h, k, and l are dimensionless parameters measuring the rigidity of a planetary body such as the Moon, and the susceptibility of its shape to change in response to tidal potential. An accurate knowledge of the Love numbers of the Moon, together with the low degree gravity field coefficients, can be used to constrain models of the deep lunar interior. Other authors have used GRAIL to obtain global density and porosity maps (Wieczorek et al., 2013). They used the correlation coefficient of the Bouguer gravity and surface topography between degrees 150 and 310, as everything of lower degrees is mainly associated with variations in crustal thickness. Wieczorek et al. (2013) minimized the correlation coefficient using data in a span of 12° of latitude, and found an average density of 2550 kg/m$^3$, with an uncertainty of ±18 kg/m$^3$ with variations on the scale of ±250 kg/m$^3$.

Density on the Moon does not only vary laterally, but also with depth. Besserer et al. (2014) used the gravity field up to degree 550 to determine the density as a function of degree, and found that density tends to increases with degree, implying that density increases with depth. Another thing GRAIL has allowed us to do, is to find things that were previously buried, such as craters (Evans et al., 2016; Sood et al., 2017). Sood et al. (2017) used gravity gradients (as discussed before) of the high degree gravity model in order to highlight the rims of potential buried craters and then cross correlated potential craters with free-air gravity and Bouguer gravity to positively identify buried craters. On the other hand, Evans et al. (2016) looked for
potential buried craters in topography, crustal thickness models, free-air gravity, Bouguer gravity, and the gravity gradient maps. These potentials he then compared to maps of “anti-eigenvalues”, which are a representation of the curvature of signal, and is therefore ideally suited to highlight circular features.

1.2 Small-scale density anomalies

So far, we have seen many interesting works done with GRAIL, but most of these studies have either focused on long length scales or on specific known features. Yet looking at the filtered Bouguer gravity map (Fig. 2.1) it is clear that there is a lot of signal there that does not appear to have a clear topographic feature associated with it. In chapter 2 I investigate the source of this “random background variability”. It has been shown that ~98% of the power of the gravity signal at wavelengths less than ~68 km (spherical harmonic degrees greater than 80) is correlated with the topography (Zuber et al., 2013b). The remaining 2% of the signal cannot be explained by topography, and therefore contains important information about the subsurface. Removing the effects of topography from the free-air gravity results in the Bouguer gravity map, which contains this 2% remaining signal. The long wavelength (up to degree 80) part of the Bouguer signal is the result of crustal thickness variations (Wieczorek et al., 2013). At the shorter wavelengths (10’s of km) the crustal thickness models become unstable and so the remaining gravity anomalies must arise from shallower depths. This variability should be considered to be a geologic signature that contains real and important signal due to small variations arising from the natural complexity of the crust.

In order to determine the dominant source of the gravity anomalies, density models are calculated by inverting the data. No a priori information is assumed other than a smoothness
constraint. First, the model is tested by applying it to the mantle uplift beneath the Freundlich-Sharonov impact basin. Next, it is compared to the model results to previous work done on narrow linear features found in the gravity gradients of the GRAIL data (Andrews-Hanna et al., 2013). Then, synthetic data is used in order to investigate the sensitivity of the model results. The gravity for three areas where this random background variability exists is inverted in order to obtain 3D density models. These three density models are interpreted in terms of three end-member scenarios. In the first scenario, all the density anomalies are due to changes in porosity alone. From lunar samples (Kiefer et al., 2012a) it is known that porosity varies from 2-20%, therefore if the observed density anomalies can result from changes in porosity that fall within that range we consider the end-member scenario possible. The second end-member scenario is assuming an intrusion into the crust. From literature, we can find constraints on the fractional volume of intrusions that is expected / observed, which falls within less than 45% (Crites and Lucey, 2015). The third end-member scenario assumes that basic crustal compositional differences explain the observed density anomalies. From remote sensing we know that the lunar crust is composed mainly of anorthosite, formed as a floatation crust from the lunar magma ocean (Smith et al., 1970; Wood et al., 1970; Warren and Wasson, 1977, 1979; Warren, 1985). However, the samples obtained from the Apollo missions, as well as lunar meteorites, show that the lunar crust is compositionally heterogeneous, with major rock types including anorthosite, norite, troctolite, and sometimes gabbro (Dymek et al., 1976; Papike et al., 1998; Warren, 1993; Wieczorek et al., 2006). The difference in density between an anorthosite and a noritic anorthosite is 200 kg/m$^3$, so this could explain the observed density difference, but only if the total fractional volume of norite does not exceed 100%. The areas of interest are also compared to results from remote sensing to see if there is any correlation between minerals and density.
In chapter 3 I take a closer look at the ejecta blanket around the Orientale Basin. Topography and imagery around Orientale shows chains of small, fresh secondary craters form linear structures or catenae analogous to the rays distant from young craters such as Tycho, interpreted to have formed from the re-impact of material ejected from the basin (Shultz, 1976; Wilhelms, 1976; Scott et al., 1977; Wilhelms et al., 1978; Spudis et al., 1984; Spudis et al., 2014). However, the specific details of catenae formation and their relation to the basin ejecta remain unclear (Wilhelms et al., 1978). Now with GRAIL one can see the gravity structure of the ejecta blanket and find linear gravity anomalies that radiate out from the center of the Basin, similar to the secondary crater chains. The associated density anomalies could be a result of the effects of the impacting ejecta on the underlying crust, through the compaction or creation of porosity as observed for primary impacts (Soderblom et al., 2015; Milbury et al., 2015). Alternatively, the density anomalies may be an effect of density variations within the ejecta blanket of Orientale itself, which has been shown to have a much lower density than typical lunar crust (Wieczorek et al., 2013).

In chapter 3 I investigate the nature and source of the linear gravity anomalies that radiate out from the Orientale Basin as revealed by GRAIL. Secondary crater chains observed in images from the Lunar Reconnaissance Orbiter Wide Angle Camera (WAC; Robinson et al., 2010) are compared to linear gravity features observed in GRAIL data (Zuber et al., 2013b; Lemoine et al., 2013, 2014). An automated mapping technique is used based on radial smoothing of the data to highlight potential radial gravity anomalies. Creating average profiles over both radial gravity anomalies and crater chains in topography data from the Lunar Orbiter Laser Altimeter (LOLA; Smith et al., 2016) aboard the Lunar Reconnaissance Orbiter (LRO) spacecraft (Chin et al., 2007), as well as in the Bouguer corrected gravity and gravity gradients from GRAIL help to aid
in the interpretation of the gravity anomalies. iSALE is used to model low-velocity impacts in order to constrain the fate of the ejecta material and its effect on the underlying crust. Finally, continuous density inversions (Li and Oldenburg, 1996; 1998; Liang et al., 2014; Jansen et al., 2017) are calculated to model the density structure in some key areas around the Orientale Basin. The results of these inversions are compared to forward models of a rectangular prism as a second independent analysis.

The results in chapter 2 demonstrate that the Moon is not homogenous in density at any scales, in contrast to the assumptions of most studies, and that studies of small scale anomalies may be hindered by this random background variability. This variability is not noise in the data, and must be taken into consideration when looking at gravity models beyond degree ~80. The results in chapter 3 show that the ejecta blanket and secondary crater chains of large basins such as Orientale contribute substantially to the density variations in the upper crust. That work also shows that although background variability is a consideration, there are ways to enhance the signal in order to see gravitational features that stand out from the background. Together, this work reveals important information about the small-scale density variability of the lunar crust as now revealed by the high-resolution data from GRAIL.

1.3 Contributions from co-authors

In the first paper (chapter 2), I owe the initial idea and many helpful hints to Dr. Jeff Andrews-Hanna. The inversion model GRAVPLOT3D was provided by Dr. Yaoguo Li, who has allowed me to use his program, and gave me some insight into how best to use it. Dr. Jeff Taylor provided the remote sensing data and an explanation. All of the co-authors, Jeffrey C. Andrews-Hanna, Yaoguo Li, Paul G. Lucey, G. Jeff Taylor, Sander Goossens, Frank G. Lemoine, Erwan
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CHAPTER 2
SMALL-SCALE DENSITY VARIATIONS IN THE LUNAR CRUST REVEALED BY GRAIL


Abstract

Data from the Gravity Recovery and Interior Laboratory (GRAIL) mission have revealed that ~98% of the power of the gravity signal of the Moon at high spherical harmonic degrees correlates with the topography. The remaining 2% of the signal, which cannot be explained by topography, contains information about density variations within the crust. These high-degree

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Bouguer gravity anomalies are likely caused by small-scale (10’s of km) shallow density variations. Here we use gravity inversions to model the small-scale three-dimensional variations in the density of the lunar crust. Inversion results from three non-descript areas yield shallow density variations in the range of 100-200 kg/m$^3$. Three end-member scenarios of variations in porosity, intrusions into the crust, and variations in bulk crustal composition were tested as possible sources of the density variations. We find that the density anomalies can be caused entirely by changes in porosity. Characteristics of density anomalies in the South Pole-Aitken basin also support porosity as a primary source of these variations. Mafic intrusions into the crust could explain many, but not all of the anomalies. Additionally, variations in crustal composition revealed by spectral data could only explain a small fraction of the density anomalies. Nevertheless, all three sources of density variations likely contribute. Collectively, results from this study of GRAIL gravity data, combined with other studies of remote sensing data and lunar samples, show that the lunar crust exhibits variations in density by ±10% over scales ranging from centimeters to 100’s of kilometers.

1. Introduction

Gravity analysis is an excellent tool for characterizing the nature of subsurface structures, since lateral density variations in the subsurface result in lateral variations in the gravity field. Gravity analyses have, for example, assisted in detecting faults, intrusions, changes in porosity or composition, variations in crustal thickness, and other structures beneath the surface (e.g., Andrews-Hanna et al., 2013; Soderblom et al., 2015; Milbury et al., 2015). High-resolution gravity data from the Gravity Recovery and Interior Laboratory (GRAIL) mission
Figure 2.1: Global Lunar Maps. A) Topography, expanded to spherical harmonic degree 1800. B) Bouguer gravity filtered using the complementary minimum amplitude filter (see text). C) Gravity Gradient map, with a high-pass filter applied at degree 50 and a low-pass cosine filter applied from degree 350 to degree 400. The outlined boxes refer to the three study areas of figure 2.6.
(Zuber et al., 2013a) provide constraints on the structure of the lunar crust at scales ranging from 10’s to 1000’s of km.

At long wavelengths, the gravity field largely reflects variations in the thickness and compensation state of the crust (Neumann et al., 1996; Wieczorek and Phillips, 1998; Wieczorek et al., 2013). It has been shown that ~98% of the power of the gravity signal at wavelengths less than ~68 km (spherical harmonic degrees greater than 80) is correlated with the topography (Zuber et al., 2013b). The remaining 2% of the signal cannot be explained by topography, and therefore contains important information about the subsurface. This subsurface signal is contained in the Bouguer gravity, which is the gravity field after it has been corrected for the effects of topography (Fig. 2.1). At half wavelengths greater than ~68 km, the correlation between topography and gravity is weak (Wieczorek et al., 2006) due to the effects of crustal thickness variations (Neumann et al., 1996; Wieczorek and Phillips, 1998; Namiki et al., 2009; Huang and Wieczorek, 2012; Wieczorek et al., 2013; Zuber et al., 2013b), mascon loading (Andrews-Hanna, 2013; Melosh et al., 2013; Freed et al., 2014), and lithospheric flexure. These long-wavelength signals are commonly inverted for global crustal thickness modeling, where the assumption is made that the crust of the Moon is characterized by either a constant density or by long-wavelength variations in regional density (Wieczorek et al., 2006; Wieczorek et al., 2013). However, downward continuation of the gravity anomalies to the crust-mantle interface causes a degree-dependent amplification, resulting in instabilities in the crustal thickness models at higher degrees. Recent GRAIL-derived crustal thickness models required a low-pass filter with an amplitude of 0.5 at degree 80 for global model stability (Wieczorek et al., 2013), despite the fact that there is negligible noise in the data even at much higher degrees (Zuber et al., 2013b). Therefore, the high-degree Bouguer anomalies cannot be entirely explained by variations in the
thickness of the crust, and are instead likely dominated by small-scale (10’s of km) shallow
density variations within the crust.

At small scales, most of the free-air gravity anomalies do correlate with topography,
and provide a constraint on the mean density of the crust. Wieczorek et al. (2013) found that the
lunar crust has an average bulk density of 2550 kg/m$^3$ and, assuming that the surface
composition of the Moon is representative of the upper lunar crust, an average porosity of 12%
down to depths of several kilometers. Regional density variations of ±250 kg/m$^3$ as revealed by
the spectral correlation between gravity and topography can be explained by a combination of
variations in composition and porosity (Huang and Wieczorek, 2012; Wieczorek et al., 2013).
Subsequent studies have shown that the density also increases with depth in the crust outside the
maria (Han et al.; 2014 Besserer et al., 2014), and is at least partially attributed to decreasing
porosity with increasing lithostatic pressure and temperature. Both studies also found evidence
for substantial lateral density variations at scales of 100’s to 1000’s of km with amplitudes of
±250 kg/m$^3$.

Gravity data have also been used to study small structures such as dikes, intrusions, and
basin-related structures using forward models in the spatial domain (Andrews-Hanna et al., 2013;
Kattoum and Andrews-Hanna, 2013; Kiefer, 2013). However, such analyses of GRAIL data
must deal with the pervasive small-scale variability in the Bouguer gravity field that does not
show any apparent correlation with known surface or subsurface structures. This variability is
most apparent in maps of the Bouguer gravity gradients (Andrews-Hanna et al., 2013), in which
the background variability is in many places comparable magnitude to signals arising from basin
rings and giant intrusions into the crust. This variability can in some cases be circumvented by
averaging along lines of symmetry, such as azimuthal averaging around basins (Andrews-Hanna,
Thus, both spherical harmonic crustal thickness modeling and spatial domain studies of discrete structures are limited by the substantial small-scale variability in the lunar gravity field. This variability can be minimized by appropriate filtering and averaging, effectively treating it as noise. However, the power and error spectra of the gravity clearly show that this background variability is not noise (Zuber et al., 2013b). Rather, this variability should be considered to be a geologic signature that contains real and important signal due to small variations arising from the natural complexity of the crust (Fig. 2.2A). These short-wavelength Bouguer signals typically cannot be attributed to either any surface structure or to relief along the crust-mantle interface, and must therefore be interpreted as arising from subsurface density variations within the crust. For crustal thickness modeling, the large scale signal is mainly of importance and so a filter is

![Figure 2.2](image)

Figure 2.2: A) Root mean square power spectrum of GRAIL gravity model (GRGM900B_Bouguer). Red dotted line indicates the degree at which the minimum amplitude filter has a magnitude one half. B) Minimum amplitude (black; Wieczorek et al., 2013) and complementary minimum amplitude (red) filters applied to the gravity data. The complementary minimum amplitude filter was used in all analyses in this work. A low-pass filter from degree 550 to degree 600 was added to the complementary filter (not shown here) to prevent ringing, applied to the Bouguer gravity in the spectral domain that removes these short wavelengths (Fig. 2.2A, black line). In order to enhance the small-scale features, the inverse of that filter can be applied, capturing all anomalies not associated with crustal thickness variations (Fig. 2.2B, red line).
From remote sensing we know that the lunar crust is composed mainly of anorthosite, formed as a floatation crust from the lunar magma ocean (Smith et al., 1970; Wood et al., 1970; Warren and Wasson, 1977, 1979; Warren, 1985. However, the samples obtained from the Apollo missions, as well as lunar meteorites, show that the lunar crust is compositionally heterogeneous, with major rock types including anorthosite, norite, troctolite, and sometimes gabbro (Dymek et al., 1976; Papike et al., 1998; Warren, 1993; Wieczorek et al., 2006). The samples vary widely in both porosity and density (Kiefer et al., 2012a), and the distribution of equivalent rocks in the crust is not constrained. Similar variations in density arising from composition and porosity would be the likely explanation for both the large-scale and small-scale density variations observed in the GRAIL data. Thus, the crustal density is variable on all length scales, from 100’s to 1000’s of km as revealed by the correlation between gravity and topography (Wieczorek et al., 2013), down to 10’s to 100’s of km as revealed by the small-scale Bouguer anomalies, even down to the scale of centimeters as revealed by lunar samples (Kiefer et al., 2012a).

The goal of this study is to investigate the nature and source of the small-scale variability in the gravity field revealed by GRAIL. We use continuous density inversions (Li and Oldenburg, 1996; 1998; Liang et al., 2014) to model small-scale density variations in the lunar crust. We first test the model by applying it to the mantle uplift beneath the Freundlich-Sharonov impact basin. Next we compare our model results to previous work done on narrow linear features found in the gravity gradients of the GRAIL data (Andrews-Hanna et al., 2013). Then we use synthetic data in order to investigate the sensitivity of the model results. Next, we invert the gravity of small areas on the Moon that lack clear large-scale features such as impact basins large enough (≥200 km in diameter) to exhibit mantle uplift that would dominate the gravity
anomalies (Neumann et al., 2015; Milbury et al., 2015). Finally, we use the derived density model to test possible interpretations of these small-scale density variations, using constraints from lunar samples, remote sensing data, and other sources of information.

2. Methods

2.1. Inversion

Bouguer gravity anomalies arise from variations in bulk density in the subsurface. These variations can be characterized by the spatially varying density anomalies \( \rho(x, y, z) \), and they produce an anomalous gravity field, \( \vec{g}_s \), that adds to the ambient gravity field. Using GRAIL data, we calculate the vertical component of this anomalous gravity field, \( g_z \), which relates to the anomalous density distribution. If one considers the anomalous gravity and density locally in a Cartesian coordinate system, the relationship is given by:

\[
g_z(\vec{r}_0) = G \int_V \rho(\vec{r}) \frac{z - z_0}{|\vec{r} - \vec{r}_0|^3} \, dv
\]

where \( G \) is the gravitational constant, \( V \) represents the volume of the density model, \( \vec{r}_0 \) is the vector indicating the location of the observation point, \( \vec{r} \) is the location of a source volume element \( dv \), \( z_0 \) is the vertical position of the observation point, and \( z \) is the vertical position of the source volume element. For each volume element in a three-dimensional density contrast model, we calculate the resulting vertical component of the gravitational acceleration acting on each point in a two-dimensional observational grid.
Equation 2.2.1 can be evaluated by discretizing the three-dimensional density distribution into model cells of unknown density anomaly. The gravity anomaly arising from the $j^{th}$ cell at the $i^{th}$ observation point can then be written as:

$$g_z(r_0) = \sum_{j=1}^{M} \rho_j \left\{ G \int_{\Delta V_j} \frac{z-z_0}{|r-r_0|^2} \, dv \right\}$$

$$= \sum_{j=1}^{M} \rho_j A_{ij}$$

where $\rho_j$ and $\Delta V_j$ are the density contrast and volume of the $j^{th}$ cell, respectively. The parameter $A_{ij}$ is a matrix that quantifies the contribution of the $j^{th}$ cell to the $i^{th}$ datum, and $M$ is the total number of model cells. In matrix notation this equation becomes:

$$\vec{g} = A\vec{\rho}$$

where $\vec{g}$ is a vector representing the anomalous gravity data, and $\vec{\rho}$ is the vector containing the density contrast values of the $M$ cells. One approach to gravity interpretation is to find the density contrast distribution that satisfies Eq. 2.3. However, the solution for a three-dimensional density distribution will be non-unique. Furthermore, any noise in the data will introduce errors into the inverse solution, further reducing the reliability of the model.

The inversion minimizes the residual vector within reasonable limits based on the expected noise in the data. One commonly used method is regularized inversion. We use the
algorithm by Li and Oldenburg (1996, 1998), in which the inverse problem is formulated as an optimization problem where an objective function of the density model is minimized. The objective function is given by:

$$\phi = \phi_d + \beta \phi_m$$

(2.4)

where $\phi_d$ is a measure of the data misfit, $\phi_m$ is the model objective function that measures the smoothness of the model, and $\beta$ is a regularization parameter. The goal is to construct a model that fits the data, but that is also characterized by smooth variations in density. Due to the non-unique nature of the problem, there are an infinite number of possible solutions, but those solutions with smoothly varying density anomalies are simpler and therefore preferred. This tradeoff between model smoothness and data misfit is controlled by the regularization parameter $\beta$. The function for the model norm used in GRAV3D (Li and Oldenburg, 1996, 1998) is:

$$\phi_m(\rho) = \alpha_s \int_v \{w^2(z)(\rho-\rho_0)^2\}dv + \alpha_x \int_v \left\{\left(\frac{\partial(w(z)(\rho-\rho_0))}{\partial x}\right)^2\right\}dv$$

$$+ \alpha_y \int_v \left\{\left(\frac{\partial(w(z)(\rho-\rho_0))}{\partial y}\right)^2\right\}dv + \alpha_z \int_v \left\{\left(\frac{\partial(w(z)(\rho-\rho_0))}{\partial z}\right)^2\right\}dv$$

(2.5)

where $\rho$ is dependent on $x$, $y$, and $z$, $\rho_0$ is the mean density contrast which in this case is assumed to be zero as there is no assumed density structure, $\alpha_s, \alpha_x, \alpha_y, \text{ and } \alpha_z$ are weighting factors that affect the relative importance of the different components of the objective function i.e. how important is model smoothness in any specific direction or overall. These terms can be adjusted
based on some *a priori* expectations of the resulting density model. We have no *a priori* information to justify different choices for $\alpha_x$, $\alpha_y$, and $\alpha_z$, and thus set them equal to one another. The ratio of $\alpha_x$ to $\alpha_s$ must scale with $dx^2$, where $dx$ is the model cell size. Testing of this inversion model during its development (Li and Oldenburg, 1998) led to a preferred choice of $\alpha_x = 4 \alpha_s dx^2$, and we adopt this weighting in our analyses. The effect of choosing a different scaling between the $\alpha$ parameters is discussed in Section 4.1. The $w(z)$ term in equation (2.5) is a depth weighting function, used to counteract the decay of the gravity with depth:

$$w(z) = (z + z_0)^{-1}$$

2.6)

where $z$ is the depth of the model point below the surface and $z_0$ is the height of the observational reference surface above the surface. The exponent of -1 is appropriate for point-source and small-scale gravity anomalies in order to counteract the decay of the gravity signal with distance (see Li and Oldenburg 1996, 1998). This depth weighting function encourages a more uniform distribution of density anomalies with depth, rather than their concentration at the surface. The data misfit is represented using a 2-norm measure (Li and Oldenburg, 1996, 1998):

$$\phi_d = \|W_d (\bar{g}_{pre} - \bar{g}_{obs})\|^2$$

2.7)

where $\bar{g}_{pre}$ is the predicted gravity data from the model density solution, and $\bar{g}_{obs}$ is the measured 2D gravity anomaly, and $W_d$ is a data weighting diagonal matrix whose diagonal elements are $1/\sigma_i$ in which $\sigma_i$ is the standard deviation of the $i^{th}$ datum.
We use the spherical harmonic GRAIL gravity model GRGM900B_BOUGUER (Zuber et al., 2013b; Lemoine et al., 2014) without the degree-0 term, where a uniform crustal bulk density of 2550 kg/m$^3$ (Wieczorek et al., 2013) was used to calculate the Bouguer gravity anomaly field. The gravity inversions assume a flat Cartesian coordinate system, but the effects of this flat geometry rather than the true spherical shape of the Moon on the resulting density anomalies should be at the ~2% level, and have no impact on our conclusions. The corresponding free-air gravity field has a signal to noise ratio that ranges from ~65,000 to 15 for degrees 80 to 550 based on the power and error spectra. Given that only ~2% of the signal arises from the subsurface, the Bouguer signal to noise ratio is reduced to ~1300 to 0.3 over the range of degrees of interest. Although this indicates that a non-trivial amount of noise is included in the data being inverted, the inversion algorithm is designed to accommodate this noise. For the gravity field used in our solution (GRGM900B_BOUGUER) no Kaula constraint was applied at degrees lower than 600. This implies that at the lower degrees the Ka Band Range Rate data completely determine the spherical harmonics coefficients. To characterize the spatial variations of the error in the data, we use clone fields (that can be obtained from: http://pds-geosciences.wustl.edu/grail/grail-l-lgrs-5-rdr-v1/grail_1001/extras/clones/), which are different solutions of the spherical harmonic coefficients that describe the data equally well within one standard deviation of the best-fit field. We use 50 statistically equivalent clone fields to calculate the error $\sigma_i$ at each datum.

The regularization parameter, $\beta$, is varied so as to optimize the tradeoff between $\phi_d$ and $\phi_m$. The model is run for a range of $\beta$ values, and the resulting model norm ($\phi_m$) is plotted against the data misfit ($\phi_d$). This plot is called a Tikhonov curve, or L-curve (Hansen, 1992) due to its typical L-shape, and the optimal $\beta$ is chosen at the elbow of the curve. The model predictions are
sensitive to the chosen value of $\beta$, and the difference in predicted density anomalies over an order of magnitude in $\beta$ can be around 40%. For the smallest values of $\beta$ considered, the models predicted unphysically large and small densities; whereas for the largest values of $\beta$ considered, the modeled density anomalies are overly smooth and do not adequately fit the gravity data.

2.2. Analysis of the Inversion Results

Given a model of the density variations within the crust, the next step is to test different interpretations of the source of those density anomalies against that model. Lunar samples exhibit a range in both composition and porosity that collectively contribute to differences in density (Dymek et al., 1976; Warren, 1993; Papike et al., 1998; Wieczorek et al., 2006; Kiefer et al., 2012a). While it is not possible to break this degeneracy between composition and porosity with gravity data alone, by testing end-member scenarios, we can evaluate which sources of density variations may have the dominant contribution to the background variability. Here, three end member scenarios are considered as possible causes of the variations in bulk density: variations in porosity, in the fractional volume of high-density volcanic intrusions, and in the composition of the primary crust itself. In reality, a combination of these scenarios likely contributes to the observed gravity field.

In evaluating the model predictions, we must consider the nature of the inversion results, which allow for long tails in the distribution of density anomalies at both high and low values. In any model, a small subset of points may exceed most reasonable constraints based on our end-member scenarios. Furthermore, orbital data tend to smooth out the smallest scales of variability, while the limited number of Apollo samples which have been analyzed (Kiefer et al., 2012a) may not capture the full range of possible properties. Rather than exclude any of our end-
member scenarios based on outliers from the inversions, we conservatively compare either the
plus or minus one standard deviation (±1σ) range in model predictions of the top 20 km of the
density model solution with observational constraints, or compare ±2σ value with the minimum
or maximum allowable value (e.g., the compositional fraction of noritic anorthosite or intrusions
in the crust cannot be less than 0% or greater than 100%).

In the first end-member scenario, we assume that all variations in density are purely due
to changes in porosity (i.e., a change in bulk density). Variations in porosity are calculated by
adding the average regional bulk density (Wieczorek et al., 2013) for each region to the modeled
density variations and calculating the porosity as follows:

\[
\phi(x, y, z) = 1 - \frac{\rho(x, y, z) + \bar{\rho}}{\rho_g}
\]

where \(\rho(x, y, z)\) is the recovered density contrast model, \(\bar{\rho}\) is the average regional bulk density, and
\(\rho_g\) is the grain density of typical lunar highland rock, which we assume to be 2900 kg/m\(^3\) (Kiefer
et al., 2012b). The porosity measured for a subset of lunar samples ranges between 2% and 20%
(Kiefer et al., 2012a), representing the porosity variations at the scale of individual rocks.
Wieczorek et al. (2013) calculated variations in the crustal bulk density by minimizing the
correlation coefficient between the Bouguer gravity and the topography in the spectral domain,
with the majority of the crust exhibiting densities between 2300 and 2800 kg/m\(^3\). The grain
density was obtained from an empirical relationship between the grain density and FeO and TiO\(_2\)
concentrations, which was then used to calculate the porosity (Wieczorek et al., 2013). The range
in porosity determined through this method of 4 – 21%, representing mean porosity variations in
the top few kilometers, at horizontal spatial scales of 100’s of km, is consistent with the observed
porosity range obtained from the lunar samples. Given the similarity in the ranges in observed
porosities at scales of 10’s of cm and 100’s of km, we expect that porosity variations at
intermediate (10-km) scales should fall within a similar range. For a given region, we assume
that a $\Delta \rho$ of 0 kg/m$^3$ from the inversion equates with the average porosity of the area (Wieczorek
et al., 2013). The porosity end-member is then considered plausible if the model-predicted $\pm 1 \sigma$
porosity range falls within the observed range of 2–21%.

The next end-member scenario considered the possibility that all variations in density
are due solely to intrusions into the crust. Evidence for intrusions into the crust has been found in
different contexts. For example, any mare extrusion must have been emplaced through a dike or
other conduit, and these solidified magma conduits likely remain today (Head and Wilson,
1992). Sills are inferred beneath floor-fractured craters, and similar intrusions could conceivably
exist even in craters lacking fractures as well as in the surrounding crust (Jozwiak et al., 2012,
2015). Cryptomaria are a form of hidden extrusive eruption that may contribute to the gravity on
a range of scales (Whitten and Head, 2013; Whitten and Head, 2015a, 2015b; Sori et al., 2016).
Plutons inferred to be the source of Mg-suite like intrusions may contribute to the density
anomalies in the crust (Prissel et al., 2014). Intrusions are also inferred beneath linear rilles
(Head and Wilson, 1993) and associated with linear gravity anomalies (Andrews-Hanna et al.,
2013). Remote sensing signatures that are typical for the highlands indicate the possibility of
some mafic material mixed into the crust (Hawke, 2003; Prettyman et al., 2006; Ohtake et al.,
2009; Pieters et al., 2009; Cheek et al., 2013; Crites and Lucey., 2015). Outcrops of pure
anorthosite are rare, and generally arise from deeper crustal levels (Warren, 1990; Hawke at el.,
2003; Longhi, 2003; Ohtake et al., 2009; Cheek et al., 2013).

For this end-member scenario, we assume that the model density $2\sigma$ below the mean
predicted by the model, equates with intrusion-free crust with the average crustal density of the
specific area (Wieczorek et al., 2013), and increases in density are due to the presence of intrusions. The $2\sigma$ value was chosen as it contains most (95%) of the signal, but excludes any outliers as discussed above, though admittedly this choice is somewhat arbitrary. Since the model density distribution is inherently smooth and the model cell size (10x10x2 km) is larger than typical intrusions, we assume that each model cell is only partially composed of intrusions.

We calculate the intrusion fraction in each cell from the density anomalies by assuming a density contrast of 600 kg/m$^3$ for intrusions, based on an intrusion bulk density of 3150 kg/m$^3$ (Kiefer et al., 2012a), appropriate for a lunar gabbro, and the appropriate regional crustal density at the locations of interest of 2550-2600 kg/m$^3$ (Wieczorek et al., 2013). The amount of FeO derived from remote sensing studies yields a range of 15%-45% of intrusive materials in typical highland terrain, under the assumption that the fraction of material in the regolith is representative of the fraction of intrusions at depth (Crites and Lucey, 2015). This range represents the two end-member assumptions made, either pure intrusions due to post-magma ocean igneous activity, or material that originated in the mantle. Floor-fractured craters (Jozwiak et al., 2012, 2015, 2016; Thorey et al., 2015) and other structures also provide evidence for shallow magmatic intrusions and sill formation, though they do not provide specific constraints on the fractional volume of intrusives in the lunar crust more generally. This end-member interpretation is tested against the model results by assuming that the modeled mean $\pm 2\sigma$ fraction range of intrusive material should not exceed 100%, but with the preferred result that the modeled mean $\pm 1\sigma$ fraction of intrusive material should not exceed the expected range obtained from remote sensing (Crites and Lucey, 2015) of 45%. This approach is conservative in that the GRAIL-derived densities are most sensitive to the upper several kilometers, whereas density is expected to increase with depth (Besserer et al., 2014), which would reduce the density contrast between the crust and the...
intrusions. For a given allowable range for the fraction of intrusive material, the increase in density with depth would place a stricter constraint on the allowable model density contrasts due to intrusions alone.

Remote sensing data indicate that the bulk of the lunar highlands surface is composed of a mixed feldspathic layer of a few 10’s of km thick, likely composed of a mixture of ferroan anorthosite, anorthositic norite, and noritic anorthosite (Cheek et al., 2013; Hawke, 2003). In the previous scenario, such a composition was interpreted as representing intermixed primary crust and intrusive material, but it may also arise from compositional variations (and the FeO content in particular) of the primary crust itself. In the previous end-member scenario, we assume that the predicted variations in density are due to compositional variations between pure anorthosite, anorthositic norite or noritic anorthosite, each of which is represented in the lunar samples (Kiefer et al., 2012a). Assuming that the -2σ density in the inversion solution corresponds to pure anorthosite, which has a grain density between 2710-2750 kg/m³ (Kiefer et al., 2012a; 2012b), and any increase in grain density is due to an increased fractional content of anorthositic norite or noritic anorthosite, which have similar grain densities, in the range of 2840-2910 kg/m³ (Kiefer et al., 2012a), the maximum density contrast expected for this end-member scenario is ~200 kg/m³. The predicted density anomalies are used to calculate the fractional abundance of anorthositic norite relative to an anorthosite background (with pure anorthosite defined as the -2σ model density, and with the constraint that the mean +2σ anorthositic norite fraction from the model cannot exceed 100%).
3. Model tests

This inversion model has been extensively benchmarked in previous studies, including the inversion of forward-modeled gravity from synthetic subsurface density anomalies in simple geometric shapes (a cube), and more complex shapes (step-like density anomalies). Li and Oldenburg (1998) found in their synthetic examples that the inversion method recovers the general shape and depth accurately. However the recovered model is smoother and therefore has a lower amplitude (90% in the center and 60% at the edges) than the original model. Here, we perform additional tests more specific to the Moon and the small-scale density variations of interest in this work.

3.1. Freundlich-Sharonov

We first applied the model to the Freundlich-Sharonov basin on the farside of the Moon (Fig. 2.3) in order to test the model against a large-scale Bouguer anomaly arising from deep mantle uplift that is well-resolved Moho uplift models. A recent study used a similar continuous density inversion to obtain a global model for density variations, including all spherical harmonic degrees (Liang et al., 2014). That study showed that the predicted density anomalies beneath the lunar basins are consistent with the expected density anomalies arising from an uplifted mantle plug. Another study showed that mantle uplift dominates the gravity signal for a crater diameter larger than ~200 km (Milbury et al., 2015). We performed a similar local analysis of Freundlich-Sharonov as a test of our model. For this model we do not apply a high-pass filter to the Bouguer
Figure 2.3: A) Bouguer corrected gravity anomaly of the Freundlich-Sharonov basin. B) Tikhonov curve (L-curve) with the data misfit on the y-axis and the model norm on the x-axis. The $\beta$ values vary from left to right on the curve in increments of a factor of 10. C) Cross-section of the optimal density model indicated by the black circle in B). D) Cross-section through the crustal thickness model of the same location in C).

gravity data, as we are interested in the long-wavelength structure of the basin. The Bouguer gravity (Fig. 2.3A) shows a strong positive anomaly in the basin center corresponding to the uplifted mantle plug (Neumann et al., 1996; Wieczorek and Phillips, 1998; Wieczorek et al., 2013; Melosh et al., 2013; Freed et al., 2014).

We performed the inversion for a range of $\beta$ values, and find the optimal model for a $\beta$ of $10^{-2}$ using the elbow point, or maximum curvature, of the L-curve (Fig. 2.3B). For this example, the L-curve spans a very short range in model smoothness ($\phi_m$), while the range in data misfit ($\phi_d$) spans several orders of magnitude. Due to the the large size and magnitude of this anomaly, the model norm is dominated by the range in density rather than the density gradient, and fitting the data more closely has little effect on the model norm. The resulting density model shows a strong positive density anomaly beneath the basin that is consistent with the expected central uplift of mantle material (Fig. 2.3C). For comparison to the density model, the crustal
thickness was modeled (Fig. 2.3D) using the software package SHTOOLS (Wieczorek, 2014), assuming an average crustal density of 2550 kg/m$^3$, an average mantle density of 3150 kg/m$^3$, and a mean crustal thickness of 40 km, reproducing the model of Wieczorek et al. (2013). The shape and location of the positive density anomaly from the continuous density inversion is consistent with the crustal thickness model. The inversion solution puts most of the density contrast at depth, as expected for uplifted mantle, showing that the depth weighting function behaves as expected for large, deep anomalies. The predicted density contrast of \(~600 \text{ kg/m}^3\) is consistent with an average crustal density of 2550 kg/m$^3$ and a mantle density of 3150 kg/m$^3$ (Wieczorek et al., 2013). The density anomaly extends from the base of the model domain up to \(~16 \text{ km depth}\), based on the average depth beneath the basin at which the density anomaly is at 50% of its maximum value, which is consistent with the expected depth of the crust-mantle interface obtained from crustal thickness modeling. These results are also consistent with the global continuous density inversions of the GRAIL data (Liang et al., 2014), indicating that the Cartesian geometry assumed in this study (rather than the global spherical inversion of Liang et al. (2014) is adequate for local studies.

### 3.2. Giant dike-like structures

We tested the sensitivity of the model against the results of a study of small-scale gravity anomalies originating in the shallow subsurface: long, linear gravity anomalies, interpreted to be giant dike-like intrusions (Andrews-Hanna et al., 2013). Here we examine one of these features in the northern farside highlands (centered at 68°N, 169°E) in more detail. The unfiltered Bouguer gravity reveals a linear positive gravity anomaly as expected based on the gravity gradients (Fig. 2.4A). We perform the inversion for a range of $\beta$ values, and again
find an optimal objective function $\beta$ of $10^{-2}$ (Fig. 2.4B). Compared to the L-curve of the Freundlich-Sharanov Basin, the linear gravity anomaly has a broader range in model smoothness ($\varphi_m$). The inversion result shows a narrow feature with a positive density contrast of $\sim 300$ kg/m$^3$ that extends to a depth of $\sim 25$-30 km, and has an average width of about 18-24 km (Fig. 2.4C&D). Both the dimensions and the density contrast are consistent with the results from Andrews-Hanna et al. (2013), though our continuous inversion predicts a broader, shallower, Hanna et al. (2013) fit the gravity using a single, discrete density anomaly, which requires the feature be buried more deeply in order to match the width and smoothness of the observed gravity anomaly.

While the discrete density anomaly is likely a better explanation for the linear gravity anomalies, a smooth and continuous density distribution likely provides a better explanation for

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**Figure 2.4:** A) Bouguer corrected gravity anomaly of linear feature 1 from Andrews-Hanna et al. (2013). B) Tikhonov curve (or L-curve). The $\beta$ values vary from left to right on the curve in increments of a factor of 10. C) Horizontal cross-section through the optimal density model solution indicated by the black circle in B), at 10 km depth. D) Vertical cross-section through the density model in C) at the location of the black line.

smoothly varying density distribution. We attribute the differences to the fact that Andrews-
the pervasive small-scale, and seemingly random gravity anomalies that are the subject of this study. These two tests show that the model is sensitive to, and accurately recovers, density anomalies located anywhere from the near surface, to the crust-mantle interface.

### 3.3. Synthetic models

Next, we use synthetic models to test scenarios representing continuous, smoothly varying density variations. For the first test we use a random Gaussian generator to simulate density variations with a 1-σ amplitude of 300 kg/m$^3$, which are similar in shape and magnitude to the observed density variations discussed in the next section. We use a forward-model to calculate the gravity ($g_2$) that would result from the synthetic density model. A 2% Gaussian noise field ($\epsilon$) is added to the resulting gravity field, simulating background noise in the data (exceeding the noise in the GRAIL data over most degrees (Zuber et al., 2013b; Konopliv et al., 2014)). We invert the resulting gravity field, as described above, and compare the resulting density model to the input synthetic density model. The optimal model recovers the shape of the density variations well within the top $\sim$20 km, but is substantially lower in amplitude, as it only recovers $\sim$60% of the original signal (Fig. 2.5).

This test demonstrates that, for random continuous density distributions, the model correctly reproduces the distribution of anomalies in the upper crust but underestimates their magnitude. The lack in recovery of deeper sources is likely due to the fact that gravity decays as $1/r^2$, where $r$ is the distance between the observer and the anomaly, and the attenuated anomalies from these deeper sources are overwhelmed by the anomalies arising from shallower depths. This imperfect recovery at depth is not an artifact of the depth weighting function, used in the inversion modeling, because the Freundlich-Sharanov model test in section 3.1 shows that the
inversion method does put large anomalies at depth. For isotropic density anomalies, the gravity is only sensitive to anomalies at depths less than or comparable to the wavelength of density variation (~20 km at degree 550). The example case of the Freundlich-Sharonov Basin shows us that at greater amplitudes and longer wavelengths, the density signature can also be detected at greater depths. However, these amplitudes and wavelengths are not reached for the small-scale density anomalies that we are considering in this work. The loss of amplitude of the shallow anomalies may be due to vertically superimposed, or laterally adjacent, anomalies of opposing sign partially cancelling one another. This can be seen in the model results where large magnitude density anomalies arising at depths greater than ~20 km were modeled as lower magnitude anomalies at shallower depths. Some loss in amplitude may also arise from the smoothing of small-scale anomalies by the inversion. In evaluating the inversion results from the small-scale gravity anomalies in the next section, the recovery of signals only in the top 20 km, and the loss in amplitude of these near-surface anomalies, must be taken into consideration.
4. Small-scale density anomalies

4.1. Small-scale density anomalies in non-descript areas of the crust

In the previous sections, distinctive structures in the subsurface were analyzed to demonstrate that the continuous density inversion gives results that, while non-unique, are both physically reasonable and consistent with expectations based on previous studies. In this section, we use GRAIL data to examine the nature and origin of the pervasive small-scale gravity anomalies that do not appear to be associated with particular structures. The Bouguer gravity is dominated by impact basins (Fig. 2.1B; Neumann et al., 2015). However, the aim of this study to model and determine the source of the small-scale gravity anomalies observed everywhere in the data. For this reason we investigate regions that are distant from known impact basins. Even away from impact basins, the long-wavelength Bouguer anomalies are thought to be associated with variations in crustal thickness (Wieczorek et al., 2013). To isolate the signal from these long-wavelength anomalies, we apply a high pass filter to the data. As discussed previously, GRAIL-derived crustal thickness models require substantial filtering to avoid instabilities arising from the amplification of short-wavelength gravity anomalies during downward continuation to the crust-mantle interface. These high-degree anomalies must originate at shallower depths than the crust-mantle interface. The GRAIL crustal thickness models were calculated using a minimum amplitude filter (Wieczorek and Phillips, 1998), which is a smoothly varying low-pass taper (Fig. 2.2B, black line) with an amplitude of 0.5 at degree 80 for the GRAIL crustal thickness models (Wieczorek et al., 2013).

Under the assumption that all of the gravity signal results either from variations in crustal thickness or from density anomalies within the crust, it can be assumed that the portion of the signal not used for crustal thickness modeling is the result of the small-scale density anomalies.
variations. Therefore, we define a high-pass filter equal to one minus the minimum amplitude filter (the “complementary minimum amplitude filter”; Fig. 2.2B, red line), which uses those spectral components of the Bouguer gravity that is not being used for crustal thickness modeling. The complementarity is only approximate, as the minimum amplitude filter is applied to the relief along the crust-mantle interface rather than the Bouguer anomaly itself. There is no unique filter to isolate signals arising within the crust, just as there is no unique filter to isolate signals arising from the crust-mantle interface. However, for globally stable models of crustal thickness and crustal density anomalies, some pair of filters resembling the minimum amplitude filter used by Wieczorek et al. (2013), and the complementary minimum amplitude filter used here, is required. Although it is possible to attribute Bouguer gravity anomalies of all wavelengths to continuous subsurface density variations (e.g., Liang et al., 2014), it is not possible to attribute anomalies of all wavelengths to variations in relief of along the crust-mantle interface. Since our focus is on the small-scale density variations in the crust that cannot be ascribed to relief along the crust-mantle interface, our results are not sensitive to the choice of filter, provided that it is complementary to a stable crustal thickness model. Different types of filter have also been considered, such as a cosine taper, and though the results vary slightly, the overall conclusions are not sensitive to the chosen type of filter. We also apply a low-pass cosine taper from degree 550 to 600, where noise begins to dominate the signal (Zuber et al., 2013a). Although other filters could be designed, the density inversions in this section provide a set of models that, together with the GRAIL crustal thickness model (Wieczorek et al., 2013), can account for all the signal in the gravity data.

A spatial map of the GRAIL gravity model GRGM900B_BOUGUER (Zuber et al., 2013b; Lemoine et al., 2014) was created using the complementary minimum amplitude filter
(Fig. 2.1B). Three areas were chosen for study (Figs. 2.1, 2.6, 2.7, 2.8, 2.9) because they lack large impact basins that would be expected to exhibit prominent mantle uplift (Wieczorek et al., 2013; Neumann et al., 2015) and to contain solidified impact melt pools (Vaughan et al., 2013; Spudis et al., 2014). The filtered Bouguer gravity maps of these areas are similar in character, with substantial variability but no clear discrete structures. Regions that are 150 km to 400 km in size were isolated from the gravity data at a horizontal resolution of 5 km per pixel, centered on 32°S, 57°E (area 1, Fig. 2.7), 19°N, 220°E (area 2, Fig. 2.8), and 0°N, 122°E (area 3, Fig. 2.9).

A typical L-curve for the inversion models show that the optimal $\beta$ is very similar to that found in the examples in section 3 (Fig. 2.10). Though L-shape of this curve is not as pronounced as in the other cases, the optimal $\beta$ was chosen by finding the maximum second derivative. The best-fit solutions from the inversions yield density anomalies that are similar in magnitude at all three areas (Figs. 2.7, 2.8, and 2.9), with a typical range of approximately $\pm 100-200 \text{ kg/m}^3$. The inversion results also show that density variations are only predicted in the top

![Figure 2.6: Topography (A-C) and Bouguer gravity (D-F) maps of non-descript areas (locations indicated in Fig. 2.1). Gravity data was filtered with the complementary minimum amplitude filter.](image-url)
Figure 2.7: Optimal inversion result for non-descript Area 1. The horizontal cross-section (A) is taken at 2 km depth, and the vertical cross-section (B) is taken at the location of the black line in A. The color bars each refer to a different analysis of the solution. FVN indicates the change in composition of the bulk lunar crust as represented by the fractional volume of noritic anorthosite, FVB indicates the fractional volume of a basaltic intrusion into the crust.

Figure 2.8: Optimal inversion result for non-descript Area 2 (details are the same as Fig. 2.7).

Figure 2.9: Optimal inversion model for non-descript Area 3 (details are the same as Fig. 2.7).
Figure 2.10: Tikhonov/L-curve for area 3. The optimal $\beta$ for this area is the knee point where $\beta=10^{-2}$ and the two extremes are also indicated.

~20 km of the crust. However, as discussed above, this is an expected result for a random, continuous variation in density due to the attenuation of gravity with distance, such that deeper small-scale anomalies are masked by the near-surface anomalies and are not resolved in the model. Therefore, we only consider the results in the upper 20 km of the model.

We next test the three end-member scenarios for the sources of the density anomalies. As discussed above, lunar samples exhibit a porosity range of 2-20% (Kiefer et al., 2012a). The inversion results predict 1$\sigma$ ranges in porosity from 8%-19% (Table 1) varying around the mean porosity obtained from Wieczorek et al. (2013), which is consistent with the range observed in the lunar samples. Thus, the end-member scenario in which the majority of density anomalies arise from variations in porosity is a possible explanation for the small-scale density variations in these non-descript areas.

The second end-member scenario considered is that the majority of variations in density are due to intrusions into the crust. For areas 2 and 3, we calculate a fractional volume of intrusions with a 1$\sigma$ range of 8%-50% (Table 2.1). We assume a lunar gabbro intrusion with a bulk density of 3150 kg/m$^3$ (Kiefer et al., 2012a). The predicted fractions somewhat exceed the
Table 2.1: The mean and standard deviation values of the three non-descript areas for density anomaly, porosity, fractional volume of norite (FVN), and fractional volume basalt (FVB).

<table>
<thead>
<tr>
<th></th>
<th>Density anomaly (kg/m$^3$)</th>
<th>Porosity (%)</th>
<th>FVN (%)</th>
<th>FVB (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area 1</td>
<td>$\mu = -1.7$</td>
<td>$\mu = 15.0$</td>
<td>$\mu = 146.2$</td>
<td>$\mu = 73.1$</td>
</tr>
<tr>
<td></td>
<td>$\sigma = 110.5$</td>
<td>$\sigma = 3.7$</td>
<td>$\sigma = 73.7$</td>
<td>$\sigma = 36.8$</td>
</tr>
<tr>
<td>Area 2</td>
<td>$\mu = -0.2$</td>
<td>$\mu = 11.0$</td>
<td>$\mu = 91.7$</td>
<td>$\mu = 33.3$</td>
</tr>
<tr>
<td></td>
<td>$\sigma = 91.8$</td>
<td>$\sigma = 3.1$</td>
<td>$\sigma = 45.9$</td>
<td>$\sigma = 16.7$</td>
</tr>
<tr>
<td>Area 3</td>
<td>$\mu = 0.1$</td>
<td>$\mu = 15.0$</td>
<td>$\mu = 49.1$</td>
<td>$\mu = 16.4$</td>
</tr>
<tr>
<td></td>
<td>$\sigma = 49.1$</td>
<td>$\sigma = 1.6$</td>
<td>$\sigma = 24.5$</td>
<td>$\sigma = 8.2$</td>
</tr>
</tbody>
</table>

expected ±1σ range of 15%-45% (Crites and Lucey, 2015) and the +2σ fraction of 67% is below the upper bound of 100% (values above which are physically impossible). Area 1 however, has a calculated fractional volume of intrusions with a 1σ range of 36%-110%. This is outside the expected range, and the +2σ fraction is 148% (physically impossible). Therefore, intrusions could also explain the majority of the small-scale density variations for areas 2 and 3, but this explanation is unlikely for area 1.

For the third end-member scenario, we assume that the majority of the variations in density in the subsurface are due to compositional variations between ferroan anorthosite and anorthositic norite or noritic anorthosite. The Moon may have a two-layered crust, consisting of an upper pure anorthosite crust and a lower noritic crust (Wieczorek et al., 2006). If this is the case, reworking of the crust by impactors may have mixed these two layers creating a compositional and density variations that we can detect with GRAIL data. Differences between pure anorthosite and anorthositic norite yields a density contrast of about 200 kg/m$^3$ (Kiefer et
We calculate a fractional volume of anorthositic norite or noritic anorthosite with ±2σ ranges between 0% and 98% for area 3, but with +2σ values of 290% and 180% for areas 1 and 2, respectively (Table 2.1). The ±2σ values for area 3 is close to 100%, which implies that the majority of the observed density contrast can be explained by changes in bulk crustal composition. However, areas 1 and 2 require fractional volumes of anorthositic norite much greater than 100%, which implies that the majority of the density contrast cannot be explained by changes in crustal composition alone.

These analyses, however, assume that the inversion accurately retrieved the magnitude of the density anomalies. The forward model we describe in section 3.3 suggests that the inversions only recover 60% of the magnitude of the anomalies. Therefore, we repeat the end-member scenario calculations, but with an increase in amplitude in the modeled density (Table 2.2). For the first end-member scenario, in which changes in density are mainly due to porosity, we calculate a porosity range of 9%-21%. The calculated porosity is still within the accepted limits, and so the majority of the density variations can be explained by changes in porosity. For the second end-member scenario, where changes in density are due to compositional variations between pure anorthosite and noritic anorthosite or anorthositic norite, the ±2σ value for all three areas greatly exceeds 100%. Area 3 has a fractional volume of noritic anorthosite with a mean ±1σ range of 41%-123%, and areas 1 and 2 have a range of 46%-367%. The 1σ range for areas 1 and 2 greatly exceed 100%, so this end-member scenario alone cannot explain the majority of observed density variations. For the last end-member scenario, where the density variations are due to intrusions into the crust, the ±1σ range for area 3 of 14%-41% is in the expected range of 15%-45% (Crites and Lucey, 2015), while the range for areas 1 and 2 of 58%-183% and 29%-85%, respectively, greatly exceed the expected range. These results indicate that this end-
Table 2.2: The mean and standard deviation values of the three non-descript areas for density anomaly, porosity, fractional volume of norite (FVN), and fractional volume basalt (FVB), corrected for the assumption of only recovering 60% of the original model.

<table>
<thead>
<tr>
<th>Density anomaly (kg/m$^3$)</th>
<th>Porosity (%)</th>
<th>FVN (%)</th>
<th>FVB (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Area 1</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\mu = -2.9$</td>
<td>$\mu = 15.0$</td>
<td>$\mu = 243.7$</td>
<td>$\mu = 121.8$</td>
</tr>
<tr>
<td>$\sigma = 184.2$</td>
<td>$\sigma = 6.1$</td>
<td>$\sigma = 122.8$</td>
<td>$\sigma = 61.4$</td>
</tr>
<tr>
<td><strong>Area 2</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\mu = -0.4$</td>
<td>$\mu = 11.0$</td>
<td>$\mu = 152.8$</td>
<td>$\mu = 55.6$</td>
</tr>
<tr>
<td>$\sigma = 153.1$</td>
<td>$\sigma = 5.2$</td>
<td>$\sigma = 76.5$</td>
<td>$\sigma = 27.8$</td>
</tr>
<tr>
<td><strong>Area 3</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$\mu = 0.1$</td>
<td>$\mu = 15.0$</td>
<td>$\mu = 81.8$</td>
<td>$\mu = 27.3$</td>
</tr>
<tr>
<td>$\sigma = 81.8$</td>
<td>$\sigma = 2.7$</td>
<td>$\sigma = 40.9$</td>
<td>$\sigma = 13.6$</td>
</tr>
</tbody>
</table>

These results indicate that porosity variations can explain the majority of the observed density variations in all three areas, while intrusions can explain the majority of the density variations in area 3 and possibly area 2. Compositional variations of the bulk lunar crust are unlikely to explain the majority of the density variations in any area once the imperfect recovery of the magnitude of the density anomalies is taken into account. From sample data and remote sensing observations we know that the crust varies in both composition and porosity, therefore, the most likely explanation for the density variations is a combination of the three end-member scenarios. The results of our end-member calculations, however, suggest that porosity variations likely play the dominant role.

As previously noted in section 2.1, the model results are sensitive to choice of $\beta$. We have chosen the optimal model in this work based on a tradeoff between model smoothness and fit to
the data, but we cannot rule out other models. Taking area 1 as an example, the 1σ range in density for the optimal model ($\beta=10^{-3}$) is 110.5 kg/m$^2$. An increase in $\beta$ by an order of magnitude results in a 1σ range of 71.9 kg/m$^2$ (a decrease in density range of 35%). This higher $\beta$ value would similarly decrease the required ranges in porosity, crustal composition, and intrusion fraction. A decrease in $\beta$ by an order of magnitude yields a 1σ density range of 188.6 kg/m$^2$ (an increase in density range of 41%). This lower $\beta$ value would increase the required ranges in porosity, crustal composition, and intrusion fraction. However, the models that have a higher $\beta$ are overly smooth, lack detail, and provide poor fits to the data, effectively not taking advantage of the extraordinarily high resolution of the GRAIL data. The models that have a lower $\beta$ show a great deal of detail but loose coherency and require unrealistically large variations in density, and therefore are not geologically likely. Nevertheless, the range in porosity can be stretched to beyond what is observed in lunar samples by choosing a smaller $\beta$, or the range in predicted fractional volume basalt and compositional variations can be reduced so as to allow these end-members to explain all the density variations by choosing a larger $\beta$. The inversion is also sensitive to the relative weighting of the different terms in $\phi_m$ ($\alpha_z$ in comparison to $\alpha_x$, $\alpha_y$, and $\alpha_z$). If we increase or decrease $\alpha_z$ by an order of magnitude relative to $\alpha_z$, the magnitude of the resulting density anomalies decrease or increase by approximately 40%, respectively. Although our results from the optimal value of $\beta$ and the nominal values for the weighting of the terms in the objective function are our preferred results, we cannot rule out density anomalies ~40% larger or smaller. Therefore, it is important to perform additional analyses to test the consistency of the inversion results, which we do in the next sections using analyses of the floors of impact basins and comparison of the models to remote sensing data.
4.2. Basin floors

To investigate the timing and nature of the emplacement of the small-scale density anomalies, we turn our attention to the floors of large basins. If the formation of these small-scale gravity and density anomalies predates the formation of a basin, we should expect to see no evidence for such anomalies in the floor of the basin, since pre-existing density anomalies of any sort would be expected to be homogenized in the impact melt pool. If, however, the formation of the small-scale density variations is a constant process that continued beyond the formation of the major basins, smaller amplitude anomalies would be expected.

Two basins are analyzed: Imbrium basin and South Pole-Aitken basin (hereafter referred to as SP-A). The gravity gradients within the Imbrium basin are nearly uniform (Fig. 2.1), suggesting a lack of small-scale variability. This could potentially represent a basin that postdates the formation of the small-scale density anomalies found elsewhere in the crust. However, the inner depression within 195 km of the basin center has been filled with mare up to ~8-9 km in thickness (Solomon and Head, 1980), though later estimates put the mare thickness between 1.5 and 7 km (Gong et al., 2016). The low amplitude of anomalies in the gravity gradients suggests that this basin fill is nearly uniform in density, as might be expected for volcanic fill, and indicates that the density variability seen elsewhere in the crust is not due to late stage (post-mare) generation of porosity variations. Gravity gradients decrease in amplitude proportional to the cube of the distance between the observer and the source, and thus the 2-9 km of mare fill would greatly diminish the amplitude of the gravity gradients at the surface arising from density anomalies beneath the mare.

In order to examine the small-scale density anomalies in the feldspathic crust beneath Mare Imbrium, we calculate the Bouguer gravity at the base of the mare infill (assumed to be at either
2 or 9 km depth) in the spherical harmonic domain. To prevent amplification of noise in the form of orbit parallel striping in the data at this location, the gravity was expanded out to spherical harmonic degree 350. The gravity data were then inverted and the mean and standard deviation of the density was calculated (Table 2.3). If we assume the mare thickness to be ~2 km, the standard deviation is 16 kg/m³, which is significantly smaller than that of the non-descript area 2 of 33 kg/m³ with the same filter applied (hence the results for area 2 differ somewhat from the discussed results in section 4.1 for this area). This would imply less variability under the Imbrium Basin. However, if we assume a mare thickness of 9 km the standard deviation increases to 52 kg/m³, which is significantly greater than that of non-descript area 2 of 33 kg/m³. The larger variability for Imbrium in this case may be either an effect of burial of pre-existing

Table 2.3: The mean and standard deviation values of Imbrium, downward continued to 2 km depth and to 9 km depth, for density anomaly, porosity, fractional volume of norite (FVN), and fractional volume basalt (FVB), compared to Area 2 and a small patch on South Pole-Aitken basin

<table>
<thead>
<tr>
<th></th>
<th>Density anomaly (kg/m³)</th>
<th>Porosity (%)</th>
<th>FVN (%)</th>
<th>FVB (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Imbrium (DC 2 km)</strong></td>
<td>µ = 1.3</td>
<td>µ = 6.9</td>
<td>µ = 22.1</td>
<td>µ = 11.0</td>
</tr>
<tr>
<td></td>
<td>σ = 15.9</td>
<td>σ = 0.5</td>
<td>σ = 10.6</td>
<td>σ = 5.3</td>
</tr>
<tr>
<td><strong>Imbrium (DC 9 km)</strong></td>
<td>µ = 0.0</td>
<td>µ = 6.1</td>
<td>µ = 88.0</td>
<td>µ = 44.0</td>
</tr>
<tr>
<td></td>
<td>σ = 51.8</td>
<td>σ = 1.5</td>
<td>σ = 34.6</td>
<td>σ = 17.3</td>
</tr>
<tr>
<td><strong>ND2</strong></td>
<td>µ = -0.2</td>
<td>µ = 11.0</td>
<td>µ = 33.2</td>
<td>µ = 12.1</td>
</tr>
<tr>
<td></td>
<td>σ = 33.3</td>
<td>σ = 1.1</td>
<td>σ = 16.7</td>
<td>σ = 6.1</td>
</tr>
<tr>
<td><strong>-SP-A</strong></td>
<td>µ = -1.2</td>
<td>µ = 11.0</td>
<td>µ = 36.9</td>
<td>µ = 13.4</td>
</tr>
<tr>
<td></td>
<td>σ = 37.5</td>
<td>σ = 1.3</td>
<td>σ = 18.8</td>
<td>σ = 6.8</td>
</tr>
</tbody>
</table>
topography by the mare, or an effect of small density variations in the mare that have been amplified by the downward continuation. Due to this large range in possible standard deviation values, we cannot for certain say whether the small-scale variations predate or postdate the Imbrium Basin formation. It is conceivable that the small-scale variations are a continuous process and would therefore both predate and postdate the formation of Imbrium.

The same filter and inversion was then applied to a small area inside the SPA basin, where no subsequent large craters or basins dominate the signal (Table 3). The resulting density anomalies are 13% greater than those in non-descript area 2 (37.5 and 33.3 kg/m$^3$, respectively). This implies that the formation of the small-scale density variations postdate the formation of the SP-A basin. Furthermore, the bulk density of the crust in SP-A (2800 kg/m$^3$) is larger than the average density of the lunar highlands (2550 kg/m$^3$; Wieczorek et al., 2013). If the small-scale density anomalies were mainly caused by either intrusions into the crust or locally greater concentrations of norite, the density anomalies relative to the mean crustal density within SP-A should be smaller than in the highlands, in conflict with the model results. If the small-scale density anomalies are mainly due to variations in porosity, the magnitude of the density anomalies in SP-A should be ~10% greater than those in the farside highlands, in agreement with the density model results. Some of the lateral variations in density in the SP-A Basin could also be caused by differentiation of the deep impact melt sea (Vaughan et al., 2013; Vaughan and Head, 2014), but these variations are expected to be of a larger scale than the small-scale density anomalies we are considering here. Thus, the results from SP-A support the hypothesis that the main cause of the small-scale density anomalies, at least within these regions of the Moon, is variations in porosity.
4.3. Remote sensing

An alternative approach to determine the degree to which compositional variability contributes to the small-scale density variations, is to compare our results with constraints placed on the surface composition from remote sensing data. Such a comparison is complicated by the fact that the density models are sensitive to the top ~20 km of crust with a vertical resolution of 2 km, while remote sensing is only sensitive to the top few microns of the crust. However, impact gardening of the surface will tend to mix the upper regolith layer (Cashore and Woronow, 1985), which would make remote sensing observations representative of deeper levels. Here we assume that the composition inferred from remote sensing is representative of the top layer of the density inversion model. We use plagioclase, olivine, pyroxene, and iron oxide maps derived from a radiative transfer analysis of 400,000 Clementine UVVIS spectra (Figs. 2.11 and 2.12; Lucey, 2004). From these maps, we generate a global map of the (olivine+pyroxene)/plagioclase ratio (Figs. 2.11 and 2.12). The (olivine+pyroxene)/plagioclase ratio is representative of density because the ratio of mafic material to plagioclase in the simple mineralogy on the Moon is proportional to grain density, as the mafic materials have a higher density than the crustal materials. The top layer of each density model was resampled at the resolution of the remote sensing data (4 pixels per degree) and then compared to the four different mineral maps as well as the mineral ratio map. We tested for a correlation between the density model and the different mineral abundances by calculating the $R^2$ value and p-value for a least squares linear fit between each pair (Figs. 2.11 and 2.12). The p-value is the probability that obtaining the result is equal to or more extreme than what is actually observed for a given null hypothesis (in this case, that the
Figure 2.11: (A-E) Mineral abundance maps derived from remote sensing data for Area 2 of pyroxene, plagioclase, olivine, iron oxide, and a (olivine+pyroxene)/plagioclase ratio, respectively. (F-J) The corresponding scatter plots, showing the best-fit line along with the associated the p-values and $R^2$ values.
Figure 2.12: Same as Fig. 2.11, but for Area 3.
density is correlated to the mineral map). We adopt a 2-σ threshold of 0.05. The $R^2$ value indicates what fraction of the variance is explained by this trend.

We find p-values <0.05 for Area 2 (Fig. 2.11) for the iron oxide, plagioclase and mineral ratio data, but the $R^2$-values are very small (below $10^{-2}$), suggesting that there is a significant correlation, but this correlation only explains a small portion of the variance in the data. The p-values obtained for Area 3 (Fig. 2.12) are all well above the threshold for all mineral maps considered with the exception of olivine. However, the $R^2$ value for olivine is again very small (below $10^{-2}$), so here we can also conclude that compositional variations can explain only a very small fraction of the observed density anomalies. These results reveal that, at least for the top layer, compositional variations contribute to <1% of the observed small-scale density variations. This result then supports the conclusion that variations in porosity are the dominant factor driving the small-scale density variations in the near surface.

4.4. Discrete vs continuous density variations

We next investigate whether the density anomalies are in the form of smoothly varying density anomalies, or discrete density anomalies, such as large (relative to the model resolution) bodies of intrusions, noritic anorthosite, or buried porous materials such as impact ejecta. Because the density inversion approach we have employed only recovers smoothly varying density anomalies, we must compare our inversions of the GRAIL data to density models derived from inversions of synthetic models. For this analysis, we again consider only the upper 20 km, since the inversions are not sensitive to small-scale anomalies at greater depths. Three different sets of inversions are compared. The first density model is the inverse solution to the gravity predicted by continuous and normally distributed random densities from Section 3.3. The
second density model begins with the inverse solution to the gravity field predicted by a related density forward model, but this continuous density model is discretized by setting all positive density anomalies to be +150 kg/m$^3$, and setting all negative density anomalies to be -150 kg/m$^3$.

The last density model is the inversion result for area 1.

We first compared the histograms of the three different data sets (Fig. 2.13), and used the statistical program R to perform the Shapiro-Wilks test for normality using a p-value threshold of 0.05 (Shapiro and Wilk, 1965). The p-value is the probability of obtaining a result is equal to or more extreme than what is actually observed given the null hypothesis that the data is normal, and we reject the null hypothesis for p-values less than 0.05. The models were also compared using a normal probability plot (Fig. 2.14), which is a special case of the quantile-quantile plot (or “q-q plot”). In this plot the quantiles (the data divided into bins equally spaced for a perfect normal distribution), are plotted against the theoretical quantiles that the data would have if the data were normal (Tamhane and Dunlop, 2000). Normally distributed data should fall

Figure 2.13: Histograms of the density inversions derived from the continuous density anomaly forward model (A), the discrete density anomaly forward model (B), and the GRAIL data over the non-descript area (C).
Figure 2.14: Normal probability quantile-quantile (QQ) plots of the density inversions of the synthetic gravity from the continuous (A) and discrete (B) density models, and of the true gravity data (C). The thin line in each panel represents a normal distribution.

on a line with a slope of 1 passing through the origin, and thus any deviations from this line implies a deviation from normality.

The histogram of the density inversion arising from the synthetic continuous density model resembles a normal distribution, as would be expected. The p-value from the Shapiro-Wilks test of 0.062 is above the 2-σ threshold of 0.05, so we fail to reject the null hypothesis of normality for this data set. The results of this inversion in the q-q plot lie very close to the expected line for a normal distribution, further supporting normal behavior.

The histogram of the inversion of the discrete density model is broader than one would expect from a normal distribution. This broader distribution might reflect the presence of two overlapping distributions with different means, as might be expected to arise for an inversion of a discrete density model. However, the histogram does not resolve two discrete peaks representing the two different densities. The p-value from the Shapiro-Wilks test is $6.2 \times 10^{-15}$, clearly low enough to reject the null hypothesis of normality, leading us to conclude that there is significant
evidence of non-normal behavior in this density model. In the q-q plot, the densities do not plot on the normal line, instead resembling an S-shape, plotting above and below the line in the negative and positive quantiles, respectively. This q-q plot indicates that the distribution is deficient in the tails, consistent with the assessment of the histogram.

The histogram for the inversion of the actual data from area 1 resembles a normal distribution, though it is heavier in the tails. The p-value of \(4.71 \times 10^{-12}\) is well below the threshold, leading us to reject the null hypothesis of normality. In the q-q plot, the results again deviate from a normal distribution in the tails, but in the opposite sense to that found in the discrete density model. The histogram for the inversion of the actual data is heavier in the tails. This analysis does not support the presence of discrete density anomalies at the lower limit of the GRAIL resolution (~10 km). These results therefore are consistent with the previous conclusion that continuous variations in porosity are the dominant factor in explaining the small-scale density variations in the upper crust of the Moon. However, this analysis does not, by itself, rule out the presence of discrete anomalies on scales smaller than the gravity can resolve (e.g., variations in porosity or composition at scales much less than 10 km).

This test shows that for the scale of the density anomalies supported by the inversions, the data and inversions favor continuous rather than discrete density anomalies. Conceptually, for subsurface sources with a horizontal scale comparable to the observed gravity anomalies with a density contrast compatible with intrusions into the crust, the histogram of the modeled density anomalies should be heavier in the tails than a normal distribution. Although we cannot rule out the possibility that models of discrete anomalies could be generated that would match the data equally well, this test favors continuous and smoothly varying density variations.
5. Conclusions

High-resolution GRAIL gravity data show a ~98% correlation in the power with topography at short wavelengths (Zuber et al., 2013b). The longer wavelength Bouguer gravity anomalies correspond with variations deeper in the lunar interior, and have been used for crustal thickness modeling (Wieczorek et al., 2013), while the shorter wavelengths must arise at depths shallower than the crust-mantle interface. Most of these short-wavelength Bouguer gravity signatures show little correlation to topographic features, with the exception of some small-scale anomalies that are correlated with basin rings or other discrete features. In this study we focused on these small-scale anomalies as an important geologic signal, rather than as noise, as they are commonly treated.

Continuous density inversion with Tikhonov regularization (Li and Oldenburg, 1996, 1998) was used to model these small-scale density variations in the lunar crust by inverting the gravity anomalies in order to solve for the underlying variations in density. A set of test cases using both previously studied lunar structures and synthetic data demonstrate that the inversion model is sensitive to both small-scale density anomalies in the top 20 km of the crust and large-scale anomalies at the crust-mantle interface.

Inversions were performed on three different small areas on the Moon, chosen because they show no clear topographic features, such as impact basins, which would dominate the gravity anomalies. The solutions reveal pervasive density variations of approximately ±100-200 kg/m³ with characteristic length scales of 10’s of km. The derived best-fit density models were used to test possible sources of these small-scale density variations, including variations in porosity, intrusions, and bulk crustal composition.
If we assume that all variations in density are due to changes in porosity alone, the
inversion results predict variations in porosity from ~6% to ~19%, which is consistent with the
range of porosities observed in the lunar samples (Kiefer et al., 2012a). If we assume that all
variations in density in the subsurface are due to compositional variations between ferroan
anorthosite and anorthositic norite or noritic anorthosite, the data require noritic anorthosite
fractions around 100% for area 3 and values greater than 100% for areas 1 and 2. All areas
greatly exceed 100% if accounting for the imperfect recovery of the density model, indicating
that variations in primary crustal composition cannot account for all of the observed small-scale
gravity anomalies. Finally, if we assume that all variations in density are due to intrusions into
the crust, we calculate a fractional volume of gabbroic intrusions no more than 50%, only
slightly exceeding the expected range of 15-45% (Crites and Lucey, 2015), though greater
intrusion fractions are required if we account for the imperfect recovery of the magnitude of the
anomalies in the inversion. Thus, the small-scale gravity anomalies revealed by GRAIL may be a
result of variations in porosity, intrusion fraction, or bulk crustal composition, but only porosity
variations can account for the full magnitude of the anomalies alone. A combination of all three
sources is likely. The density model solution for one area was compared to synthetic data sets
representing continuous and discrete density anomalies, finding that the density anomalies depart
from a normal distribution, but do not match expectations for large-scale discrete anomalies.
Thus, the sources of the density anomalies are continuous and smoothly varying at the scale of
the GRAIL gravity resolution (~10 km).

An analysis of the floors of the Imbrium and SP-A basins has shown that the density
variations observed elsewhere on the Moon are persistent in the basin floors, leading to the
conclusion that the small-scale density variations were emplaced (or continued to be emplaced)
after the formation of these basins. The somewhat larger variability in density found within the high-density crust on the floor of the SP-A basin suggests that variations in porosity may be the dominant factor in formation of the small-scale density anomalies. Furthermore, a comparison between surface composition inferred from remote sensing and the density inversion results for the non-descript areas in the highlands found that variations in the abundances of minerals indicative of high density can only explain a small fraction (<1%) of the density anomalies. Therefore, we conclude that the small-scale density variations are dominated by variations in porosity, though remote sensing data indicates that composition also plays a small role.

We acknowledge that gravity inversions are non-unique and sensitive to $\beta$, and though we have chosen a certain tradeoff between model norm and model smoothness as an optimal model, models with a different tradeoff (smaller or larger $\beta$) may also be valid. As a result the range in porosity can be stretched to beyond what is observed in lunar samples, or the predicted ranges in the fractional volume of basalt and compositional variations can be compressed to explain all the density variations. Yet our conclusion from the inversion modeling in section 4 that variations in porosity are the dominant factor in explaining the small-scale gravity inversions are supported by the basin floor models and remote sensing analyses.

We emphasize that three-dimensional density inversions based on gravity data are under-constrained and the results are non-unique. Although it may not be possible (or correct) to uniquely ascribe a single source to these gravity anomalies, we have endeavored to test different possible interpretations to the best of our abilities given available methodologies. While the model results are sensitive to assumptions, such as choice in $\beta$ and weighting parameters in the inversion, all of our approaches favor porosity as the dominant source, though we feel it likely that each of the end-members tested contributes to some extent. More importantly, these analyses
have revealed the existence and placed constraints on the magnitude of pervasive small-scale density variations in the upper crust of the Moon.

This study has shown that, not only does lunar crustal density vary systematically with depth and on large horizontal length scales (Besserer and Nimmo, 2013; Wieczorek et al., 2013), but it also varies horizontally and vertically on small ~10-km scales. These variations are the dominant signal in the high-degree component of the GRAIL Bouguer gravity data, and are important to consider when interpreting the lunar gravity field. These small-scale variations are frequently treated as noise, but these variations are real and should not be neglected when investigating localized problems. This small-scale variability is important to take into account in error analysis. In particular, studies focusing on structures with dimensions of ~10-100 km (or spherical harmonic degrees greater than ~80) with density anomalies on the order of ±200 kg/m³ may be frustrated by this background variability. Such potential target structures include dikes, lava tubes, buried craters, magma chambers, and faults. For structures at these scales, an absence of evidence in the gravity data cannot be taken as evidence for the absence of the structures. At the same time, gravity anomalies arising from this random small-scale variability could be misattributed to discrete subsurface structures (e.g., a negative or positive anomaly beneath a crater could simply be a part of this background variability with no relation to the crater). One approach to circumvent this limitation is by using large numbers of structures (e.g., Soderblom et al., 2015; Bierson et al., 2016) or averaging along lines or axes of symmetry (e.g., Andrews-Hanna et al., 2013).

In addition to acting as “noise” for studies of discrete surface structures, this small-scale density variability is an important signal in itself. These variations are an indication that many more processes are (or have been) operating in the lunar crust than those reflected in the
surface geology alone. The timing of the formation of these small-scale variations appears to postdate the formation of the major impact basins on the Moon, but to largely predate the emplacement of the majority of the basin-filling maria. Cratering provides one ongoing mechanism for generating this heterogeneity (Milbury et al., 2015), though a study of the gravity signal of complex craters found the scatter about the trend to be greater than the crater signal itself at small diameters (<100 km) (Soderblom et al., 2015). However, the net effect of the long history of impact cratering on the Moon was likely the generation of a complex heterogeneous crust, consisting of discrete zones of impact melt, ejecta, and fractured and brecciated rocks.

The GRAIL mission has given us high-resolution gravity data exceeding expectations, allowing us to investigate density variations on scales much smaller than can be resolved on other planets. The other terrestrial planets have experienced equally complex histories of processes affecting their surfaces and interiors, and thus are likely similarly heterogeneous. Previous work has shown variability on large scales of 100-1000’s of km (Besserer and Nimmo, 2013; Wieczorek et al., 2013). Here we show that variability exist on the smallest scales resolved by GRAIL of 10’s of km. From lunar samples we know that there is variability on the smaller scales of 10’s of cm and smaller, as well. The density anomalies at all scales ranging from centimeters to 1000’s of km are comparable in magnitude. Although the assumption of uniform density in analyses of structures on the Moon and other planets is often a necessary simplification, it must be recognized that this assumption is invariably incorrect. This density variability has a substantial impact on any interpretations of planetary gravity data, across all scales.
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CHAPTER 3

RADIAL GRAVITY ANOMALIES ASSOCIATED WITH THE EJECTA OF THE ORIENTALE BASIN

In preparation for submission to Icarus. Johanna C. Jansen*, Jeffrey C. Andrews-Hanna, Colleen Milbury, James W. Head III, Yaoguo Li, H. Jay Melosh, and Maria T. Zuber

Abstract

GRAIL gravity data shows small-scale gravity anomalies radiating out from the Orientale basin between distances of ~550 km (average distance to the Cordillera ring) and ~1000 km. These radial gravity lineations are sometimes associated with secondary crater chains or catenae, of which the specific details of formation and their relation to the basin ejecta remain unclear. However, many of the radial gravity anomalies have no clear topographic signature at all. Typical anomalies have an average width of about 20 km, with a gravity anomaly of ±20 mGal, and a length of 50-200 km. Here we use gravity inversions, hydrocode modeling, and observations to investigate the radial gravity anomalies in more detail. Density inversion models

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show that the gravity can be matched by solutions ranging from broad low amplitude anomalies with density contrasts of ±20 kg/m³ extending to depths of tens of kilometers, to shallow high amplitude anomalies with density contrasts of about ±200-400 kg/m³ confined to the top ~1-2 km of the crust. Hydrocode models of the low-velocity secondary impacts of basin ejecta show that the impacting material does not disperse as happens in the case of hypervelocity impacts, but instead remains as a thin layer within the secondary craters. This remaining secondary projectile material derived from Orientale ejecta can explain the linear gravity anomalies associated with secondary crater chains and catenae. However, this does not explain the majority of linear gravity features that do not have secondary crater chains associated with them. Therefore, we conclude that the majority of radial gravity anomalies represent the structure of the ejecta blanket. This density variability within the ejecta of basins likely contributes substantially to the density variability of the shallow upper crust as a whole.

1. Introduction

The Gravity Recovery and Interior Laboratory (GRAIL) mission has generated unparalleled, high-resolution global gravity data for the Moon (Zuber et al., 2013b). This data has led to new discoveries including linear gravity anomalies (Andrews-Hanna et al. 2013), gravity anomalies associated with impact craters (Soderblom et al., 2015, Bierson et al., 2016), buried craters (Evans et al., 2016), new impact basins (Neumann et al., 2015), structures associated with the Orientale basin (Zuber et al., 2016), and a better estimate of the Love numbers (Lemoine et al., 2013; Williams et al., 2013; Lemoine et al, 2014). In addition to these
discoveries, the GRAIL data provide an opportunity to examine a variety of other structures known from their surface expression but not previously resolved in gravity data.

Topography and imagery around Orientale shows chains of small, fresh secondary craters form linear structures or catenae analogous to the rays distant from young craters such as Tycho, interpreted to have formed from the re-impact of material ejected from the basin (Shultz, 1976; Wilhelms, 1976; Scott et al., 1977; Wilhelms et al., 1978; Spudis et al., 1984; Spudis et al., 2014). However, the specific details of catenae formation and their relation to the basin ejecta
remain unclear (Wilhelms et al., 1978). A closer look at the gravity field of the Orientale basin and its surroundings (Fig. 3.1) shows a pattern of both positive and negative Bouguer signatures, radiating out from the center of the basin outside of the Cordillera ring that are sometimes, but not always, associated with catenae (Wilhelms, 1976). The anomalies are found at distances of ~450-650 km (distance to the Cordillera ring; Scott et al., 1974) to ~1000 km from the center of the Orientale Basin. This signature in GRAIL data makes it possible to get a more detailed picture of the density structure under the crater chains and of the ejecta blanket more generally, shedding more light on the possible formation of these structures.

The gravity anomalies are clearly associated with the ejecta blanket of Orientale. The associated density anomalies could be a result of the effects of the impacting ejecta on the underlying crust, through the compaction or creation of porosity as observed for primary impacts (Soderblom et al., 2015; Milbury et al., 2015). Alternatively, the density anomalies may be an effect of density variations within the ejecta blanket of Orientale itself, which has been shown to have a much lower density than typical lunar crust (Wieczorek et al., 2013).

The goal of this study is to investigate the nature and source of the linear gravity anomalies that radiate out from the Orientale Basin as revealed by GRAIL. We compare secondary crater chains observed in images from the Lunar Reconnaissance Orbiter Wide Angle Camera (WAC; Robinson et al., 2010) to linear gravity features observed in GRAIL data (Zuber et al., 2013b; Lemoine et al., 2013, 2014). We also use an automated mapping technique based on radial smoothing of the data to highlight potential radial gravity anomalies. We obtain average profiles over both radial gravity anomalies and crater chains in topography data from the Lunar Orbiter Laser Altimeter (LOLA; Smith et al., 2016) aboard the Lunar Reconnaissance Orbiter (LRO) spacecraft (Chin et al., 2007), as well as in the Bouguer corrected gravity and
gravity gradients from GRAIL. To aid in the interpretation of the gravity anomalies, we use iSALE to model low-velocity impacts in order to constrain the fate of the ejecta material and its effect on the underlying crust. Finally, we use continuous density inversions (Li and Oldenburg, 1996; 1998; Liang et al., 2014; Jansen et al., 2017) to model the density structure in some key areas around the Orientale Basin.

2. Methods

2.1 Gravity data and averaging linear anomalies

We use the spherical harmonic GRAIL gravity model JGGRAIL_1200C12A (Zuber et al., 2013b; Lemoine et al., 2014), correcting for the effects of topography using a spatially varying model of the surface density (Wieczorek et al., 2013) to calculate the Bouguer gravity anomaly field. The gravity field and gravity gradients were calculated at the surface of a degree 50 representation of the topography, so as to minimize the effects of different elevations on the magnitude of the anomalies (Andrews-Hanna et al., 2018).

In order to emphasize small-scale crustal density anomalies, the Bouguer gravity model is filtered using a high-pass filter that is complimentary to the filter used in crustal thickness modeling (Wieczorek and Phillips, 1998). While long-wavelength (low degree) gravity anomalies can be explained as a result of variations in the relief along the crust-mantle interface (Wieczorek et al., 2013), shorter wavelength (high degree) anomalies are unstable if downward continued to the base of the crust and require a shallow origin (Jansen et al., 2017). The GRAIL-derived crustal thickness models employed a smoothly varying low-pass filter that emphasizes the lower degrees of the spherical harmonics, and decreases the effects of the higher degrees. This filter had an amplitude of 0.5 at degree 80, which is the highest degree at which the largest
fraction of the gravity anomalies can be ascribed to the relief along the crust-mantle interface before the models become unstable (Wieczorek et al., 2013). Therefore, the majority of the signal at higher degrees must arise at shallower depths (Jansen et al., 2017). To isolate the small-scale density variations that are responsible for the small-scale gravity anomalies, we define a high-pass filter as one minus the filter applied for the crustal thickness models. This filter should highlight the density variations that occur at these shallow depths in the crust of the Moon. Next, we remove all high degrees where noise becomes apparent at some locations in the spherical harmonic gravity model using a simple low-pass cosine filter from degrees 500 to 550. Beyond degree 550, the power spectrum of the Bouguer gravity increases with degree, indicating that the gravity model is beginning to fit noise.

The filtered Bouguer gravity is also compared to the gravity gradients (Andrews-Hanna et al., 2013), which are calculated by taking the second horizontal derivatives of the Bouguer gravity ($\Gamma_{xx}$, $\Gamma_{xy}$, and $\Gamma_{yy}$). Next, the eigenvalues are calculated, and the eigenvalue of maximum absolute value at each point is used to generate a map of the maximum gravity gradient in the horizontal directions. This enhances the locations where the Bouguer gravity is dominated by small-scale short-wavelength structures, emphasizing structures such as the radial anomalies around Orientale. Note that a positive Bouguer gravity lineation appears as a negative Bouguer gravity gradient, and vice versa.

To more objectively identify and better highlight the radial gravity anomalies, we process the gravity and gravity gradient maps using a moving average window to smooth the data in the direction radial to the basin center. The best window is investigated and discussed along with the results in section 3.1. This approach will reduce the magnitude of both random small-scale anomalies (Jansen et al., 2017) and circumferential gravity anomalies associated with the basin.
rings (Andrews-Hanna et al., 2018), while preserving the magnitude of any radially oriented anomalies greater than or comparable in length to the smoothing kernel. After applying the smoothing function, we identify all remaining positive and negative anomalies exceeding the ±2.5 standard deviation (σ) range. Although this approach has the potential to connect unrelated point source anomalies, it also provides a more objective mapping of radial lineations in the gravity field.

The filtered gravity model and gravity gradients are used to create average profiles orthogonal to both the crater chains observed in WAC images and the gravity lineations. Average profiles over crater chains from topography are also obtained using the LOLA data (Smith et al., 2016). The standard error on the average profile is calculated as:

\[ SE = \frac{\sigma}{\sqrt{n}} \]  

where \( \sigma \) is the standard deviation and \( n \) is the number of observations (adjusted for the spatial resolution of the filtered GRAIL data for the gravity profiles).

2.2. Hydrocode models of the density structure of secondary crater chains

Inversions of gravity data are fundamentally non-unique, particularly with respect to the depths of the anomalies. In order to guide our interpretations of the gravity data, we first consider the origin of the gravity lineations. The correlation of some gravity lineations with chains of secondary craters on the surface indicates that these secondary impacts play a role in generating the gravity anomalies. In order to better understand the effects that low velocity secondary impactors have on the density structure of the lunar crust and their role in producing the observed gravity lineations, we look at the effects that low velocity impactors have on the density and
porosity of the crust, using hydrocode impact modeling (Milbury et al., 2015). The range of
distances (x) of the secondary crater chains from the center of Orientale is measured and used to
calculate the range in vertical velocities (u_z) upon impact for a range of assumed ejection angles
(\theta):

\[
u_z = \sqrt{\frac{g \cdot x \cdot \tan \theta}{2}}
\]

where the gravity on the Moon, g, is given by \(g = 1.64 \text{ m/s}^2\). We assume that the ejection angle
varies from 35° to 55°. The distance of the prominent secondary crater chains from the center of
Orientale was measured to be between 500 km and 1000 km. We calculate the vertical
component of the impact velocity because iSALE uses vertical impacts in its simulations.
Although the actual impacts of secondary material are at oblique angles by definition, we do not
expect the impact angle to fundamentally change the resulting density anomalies. Next, we
calculate projectile diameters responsible for observed secondary craters using the range of
velocities calculated above and the measured diameters of secondary craters using the pi-scaling
relationship adapted from Holsapple (1993) for a vertical impact:

\[
D_{cr} = 1.6 C_d d_p^{1-\beta} \left[ \frac{1}{6} \pi \frac{\rho_p}{\rho_t} \right]^{1/3} \left( \frac{1.61 g}{u_y^2} \right)^{-\beta}
\]

where \(D_{cr}\) is the final crater diameter for which we use a range between 5 to 25 km, \(C_d\) is the
drag coefficient (typically 1.6 for rock), \(d_p\) is the projectile diameter, \(\beta\) is an empirical constant
(taken to be 0.22), \(\rho_p\) and \(\rho_t\) are densities of the projectile and target (taken to be 3000 kg/m³ or
2400 kg/m³, and 2400 kg/m³, respectively), and \( u_z \) is the vertical component of the impact velocity. The chosen densities represent either low-density material ejected from shallow depths or potentially more dense material ejected from deeper levels within the crust, impacting on top of the low-density ejecta blanket of Orientale (Wieczorek et al., 2013).

The calculated ranges in impact velocity and projectile diameter are then used as inputs in iSALE hydrocode models of impacts for a range of initial target porosities (7-17%) and projectile densities of 3000 and 2400 kg/m³ to examine the predicted changes in density for comparison with the gravity inversions. The iSALE shock physics code is a program used to simulate impact processes in solid materials (Collins et al., 2004; Wunnemann et al., 2006; Milbury et al., 2015) as an extension of the SALE hydrocode (Amsden et al., 1980), typically used for hypervelocity impacts, but here applied to low velocity impacts. iSALE includes an elastic-plastic constitutive model, fragmentation models, equations of state for multiple materials, a modified strength model, a porosity compaction model, and the creation of porosity by dilatancy (Milbury et al., 2015). These models will shed light on both the changes in density of the target material and the fate of the projectile material, as well as the role that each may play in producing the observed gravity anomalies.

2.3. Gravity inversions

Since the short wavelength (high degree) gravity anomalies cannot be ascribed to relief along the crust-mantle interface, we must examine continuous variations in density within the crust (see Jansen et al., 2017 for more discussion). To investigate the radial features around Orientale, we use three-dimensional gravity inversions (Li and Oldenburg, 1996; 1998) in order to get a better understanding of the density structure of the subsurface. Inversions of gravity data
are inherently non-unique, therefore we try to find a model that provides the best compromise between fitting the data accurately and being geologically realistic, which is defined here as being smooth. This is done by defining an objective function that combines the data misfit and the model smoothness using a regularization parameter. This model was previously used to examine the nature of the small-scale density variability in the lunar crust (Jansen et al., 2017), and a similar model was applied to the global GRAIL data across all degrees (Liang et al., 2014).

Bouguer gravity anomalies result from variations in bulk density in the subsurface. These variations can be characterized by the spatially varying density anomalies $\rho(x, y, z)$, which produce an anomalous gravity field, $g'_s$, that adds to the ambient gravity field. GRAIL data provides a measure of the vertical component of this anomalous gravity field, $g_z$, which relates to the anomalous density distribution in a local Cartesian coordinate system by (Li and Oldenburg, 1996; Li and Oldenburg, 1998; Jansen et al., 2017):

$$
g_z(\vec{r}_0) = G \int_V \rho(\vec{r}) \frac{z-z_0}{\sqrt{|\vec{r}-\vec{r}_0|^3}} dV
$$

where $G$ is the gravitational constant, $V$ represents the volume of the density model, $\vec{r}_0$ is the vector indicating the location of the observation point, $\vec{r}$ is the location of a source volume element $dV$, $z_0$ is the vertical position of the observation point, and $z$ is the vertical position of the source volume element.

Equation 4 can be evaluated by discretizing the three-dimensional density distribution into model cells of unknown density anomaly, and calculating the resulting vertical component
of the gravitational acceleration acting on each point in a two-dimensional observational grid. The gravity anomaly arising from the $j^{th}$ cell at the $i^{th}$ observation point can then be written as:

$$g_{zi}(\vec{r}_i) = \sum_{j=1}^{M} \rho_j \left\{ G \int_{\Delta V_j} \frac{z_j - z_i}{|\vec{r}_j - \vec{r}_i|^3} d\nu \right\} \equiv \sum_{j=1}^{M} \rho_j A_{ij}$$

3.5)

where $\rho_j$ and $\Delta V_j$ are the density contrast and volume of the $j^{th}$ cell, respectively, and $r_i$ and $r_j$ are the position vectors for the observation point and density model cell. The parameter $A_{ij}$ is a matrix that quantifies the contribution of the $j^{th}$ cell to the $i^{th}$ datum, and $M$ is the total number of model cells. One approach to gravity interpretation is to find the density contrast distribution that satisfies Eq. 3.2. However, the solution for a three-dimensional density distribution will be non-unique. Furthermore, any noise in the data will introduce errors into the inverse solution, further reducing the reliability of the model (Li and Oldenburg, 1996; Li and Oldenburg, 1998; Jansen et al., 2017).

Regularized inversion is an approach that minimizes the residual vector within reasonable limits based on the expected noise in the data. We use the algorithm by Li and Oldenburg (1996, 1998), in which the inverse problem is formulated as an optimization problem where an objective function of the density model is minimized. The objective function is given by:

$$\phi = \phi_d + \beta \phi_m$$

3.6)
where $\phi_d$ is a measure of the data misfit, $\phi_m$ is the model objective function that measures the smoothness of the model, and $\beta$ is a regularization parameter. The model objective function is taken to be the sum of the squares of the density anomalies and the squares of the density gradients together with a depth weighting function, and the data misfit function is the sum of the squares of the differences of the observed and modeled gravity anomalies (for full details, see Jansen et al., 2017; Li and Oldenburg 1996; Li and Oldenburg 1998). The goal is to construct a model that fits the data, but that is also characterized by smooth variations in density. Due to the non-unique nature of the problem, there are an infinite number of possible solutions, but those solutions with smoothly varying density anomalies are simpler and therefore preferred. This tradeoff between model smoothness and data misfit is controlled by the regularization parameter $\beta$ (Li and Oldenburg 1996; Li and Oldenburg 1998; Jansen et al., 2017).

The regularization parameter, $\beta$, is varied so as to optimize the tradeoff between $\phi_d$ and $\phi_m$. The model is run for a range of $\beta$ values, and the resulting model norm ($\phi_m$) is plotted against the data misfit ($\phi_d$). This plot is called a Tikhonov curve, or L-curve (Hansen, 1992) due to its typical L-shape, and the optimal $\beta$ is chosen at the elbow of the curve. The model predictions are sensitive to the chosen value of $\beta$, and the difference in predicted density anomalies over an order of magnitude in $\beta$ was found to be ~40% in a previous inversion of the GRAIL data (Jansen et al., 2017). For the smallest values of $\beta$ considered, the models predicted unphysically large and small densities; whereas for the largest values of $\beta$ considered, the modeled density anomalies are overly smooth and do not adequately fit the gravity data.

For inversion modeling we use the aforementioned filtered Bouguer gravity model. We isolate three rectangular regions from the gravity data that are ~450 km in length and width, each of which contains Orientale-radial gravity anomalies. The gravity inversions assume a flat
Cartesian coordinate system, but the effects of this flat geometry rather than the true spherical shape of the Moon on the resulting density anomalies should be at the ~2% level, and have no impact on our conclusions (Jansen et al., 2017). The gravity data is taken out to spherical harmonic degree 550, which results in a half-wavelength resolution of 10 km, which we use as the density model cell size. In the vertical direction, the resolution is 2 km for the first set of density models, with a total vertical extent of the model of 60 km. Based on the results of the hydrocode models in Section 3.2, we also consider a second set of simulations in which density anomalies are forced to remain at shallow depths, with a vertical model resolution of 100 m and a total vertical extent of the model of 2 km.

As an independent check of the density anomalies that could cause the observed gravity signatures, we use a forward model of the gravity resulting from a rectangular prism, representing the projectile material remaining within a crater chain or catena or deposited along a gravity lineation. The gravity anomaly measured at the origin arising from a simple rectangular prism with density contrast $\Delta\rho$ extending between $x_1$ and $x_2$, $y_1$ and $y_2$, and $z_1$ and $z_2$, is given by (Blakely, 1995):

$$g = G\Delta\rho \sum_{i=1}^{2} \sum_{j=1}^{2} \sum_{k=1}^{2} \mu_{ijk} \left( \frac{z_k}{R_{ijk}} \arctan\left( \frac{x_i y_j}{x_i^2 + y_j^2 + z_k^2} \right) \cdot x_i \ln(R_{ijk} + y_j) - y_j \ln(R_{ijk} + x_i) \right)$$

$$R_{ijk} = \sqrt{x_i^2 + y_j^2 + z_k^2}$$

$$\mu_{ijk} = (-1)^i (-1)^j (-1)^k$$

where the x-direction is defined to be horizontal and perpendicular to the long axis of the radial anomaly, the y-direction is horizontal and parallel to the long axis of the anomaly, and the z-
direction is positive downward. The length of the anomalies in the y-direction is considered sufficiently long that end effects were neglected, so the prisms where extended ±1000 km in either direction. Based on the results of the hydrocode models and the expected thickness of the Orientale ejecta (Fassett et al., 2011), the top depth is set to 0 km and the bottom depth is varied between 0.3 and 3 km. The width and density contrast of the anomaly are left as free parameters.

The solutions are compared to observed average gravity profiles, and the root mean squared error (RMSE) of each model is calculated.

\[
RMSE = \sqrt{\frac{\sum_{i=1}^{n}(\hat{g}_i - g_i)^2}{n}}
\]

where \( \hat{g}_i \) are the predicted gravity values, \( g_i \) are the observed values, and \( n \) are the number of observations. Note that only the RMSE of the region where the anomaly is seen is calculated.

3. Results and discussion

3.1. Observations

A Bouguer map of the Orientale basin and its surroundings (Fig. 3.1) shows patterns of both positive and negative linear gravity anomalies radiating out from the center of the basin outside the Cordillera ring. These linear anomalies vary in length and amplitude, as well as distance from the Orientale basin, but can be found all around Orientale, outside of regions in which a subsequent basin overlaps the area and obscures the anomalies.
Figure 3.2: Radial gravity anomalies based on radially averaged Bouguer gravity (left) and gravity gradient (right) maps. All gravity anomalies that are higher or lower than $\pm 2.5\sigma$ of the background variation is plotted in red and blue, respectively. Note that positive gravity gradients correspond to negative gravity anomaly and are highlighted in blue in the panel on the right. Figure are in a radial projection centered on Orientale.
Figure 3.3: Catena Leuschner location and average profiles. The section of the catena analyzed is shown by the black line in A. Average profiles are shown derived from the topography (B), Bouguer gravity (C), and gravity gradients (D). The black and red lines in B-D indicate the mean and 1 standard error uncertainty.
To more objectively identify and better highlight the radial gravity anomalies, we process the gravity and gravity gradient maps using a moving average window to smooth the data in the direction radial to the basin center, and highlight only those anomalies exceeding the $\pm 2.5-\sigma$ (99%) level. As a test, the radial averaging centered on Orientale basin was compared to the same procedure centered on an area picked not in the center of any impact basin (not shown). Different lengths to average over were investigated, and the length that highlights the features in the Orientale Basin clearly, but doesn’t result in a comparable number of false positives in the random area was chosen. It was found that the best radial averaging length for this analysis is 60 km, compared with the typical length of the shorter anomalies of $\sim 100$ km (Fig. 3.2). For the Orientale Basin, we removed areas affected by other major basins in the area for the sake of clarity. The resulting map shows many radial features, and there is no clear preference for negative or positive features. Gaps without radial anomalies correspond to locations of major basins excluded from the analysis. Note that not all linear gravity features will show up clearly, because some are not perfectly radial to the center of the basin, and some radial structures may be false positives due to the radial stretching of an uncommonly large gravity anomaly.

We now take a closer look at linear anomalies that have a clear surface expression. Catena Leuschner (Fig. 3.3) consists of a crater chain extending toward the northwest between distances of 650 and 800 km from the center of the basin. Neither the gravity nor gravity gradients show an obvious linear anomaly associated with the most prominent secondary crater chain of the catena, and this feature was not highlighted in the automated radial anomaly map.
Figure 3.4: Catena Michelson location and average profiles. The section of the catena analyzed is shown by the black line in A. Average profiles are shown derived from the topography (B), Bouguer gravity (C), and gravity gradients (D). The black and red lines in B-D indicate the mean and 1 standard error uncertainty.
Figure 3.5: Vallis Bouvard location and average profiles. The section of the catena analyzed is shown by the black line in A. Average profiles are shown derived from the topography (B), Bouguer gravity (C), and gravity gradients (D). The black and red lines in B-D indicate the mean and 1 standard error uncertainty.
Nevertheless, closer to the basin linear anomalies along this same trend were highlighted in the automated mapping and are visible in the gravity gradient map. A positive linear gravity anomaly extends approximately 200 km from the end of Catena Leuschner toward Orientale, and several positive and negative radial anomalies are found in the vicinity. A more subtle continuation of the catena as a low relief groove or scour is seen here in WAC images and topography. In the average profile, the main catena crater chain itself shows an average depth of ~2 km deep and width of ~15 km, consistent with earlier mapping (Wilhelms, 1976). The slight positive anomaly of ~7 mGal corresponding to the location of Catena Leuschner in the average gravity profile is not sufficient to rise above the background variability (Jansen et al., 2017).

Southwest of Catena Leuschner, another secondary crater chain, Catena Michelson (Fig. 3.4), shows an average depth of ~1.5 km and a width of ~20 km similar to Catena Leuschner (Wilhelms, 1976). Again, no obvious linear anomaly is observed in the gravity or gravity gradients and the automated mapping did not highlight any structures aligned with this trend. Here there is a slight negative gravity anomaly of about ~7 mGal below the surrounding in the average profile, yet still not clear enough to stand out above the background variability (Jansen et al., 2017). Other portions of Catena Michelson closer to the basin show evidence of both positive and negative gravity anomalies aligned with the catena, but no consistent anomaly persists for the full length of the feature. More prominent positive and negative anomalies along similar trajectories are found immediately south of Catena Michelson. Several minor unnamed catenae also align approximately with gravity lineations.

Vallis Bouvard is the only structure with a prominent signature in both topography and gravity (Fig. 3.5, Fig. 3.8 C, F, and I). This structure is a prominent topographic valley of about 2 km in depth. This trough is associated with a clear positive gravity anomaly of ~15 mGal in the
Figure 3.6: P1 location and average profiles. The section of the catena analyzed is shown by the black line in A. Average profiles are shown derived from the topography (B), Bouguer gravity (C), and gravity gradients (D). The black and red lines in B-D indicate the mean and 1 standard error uncertainty.
Figure 3.7: Location and average profiles across a negative radial gravity anomaly that is not associated with a secondary crater chain. The section analyzed is shown by the black line in A. Average profiles are shown derived from the topography (B), Bouguer gravity (C), and gravity gradients (D). The black and red lines in B-D indicate the mean and 1 standard error uncertainty.
Figure 3.8: Close up views of the areas outlined in figure 3.1. Panels A-C show topography, panels D-F show the high-pass filtered Bouguer gravity, and panels G-I show the maximum gravity gradients. The arrows next to the panel titles indicate the direction to the basin center, and the dashed outlines highlight some of the linear anomalies seen in topography and/or in the gravity. N1 and N2 are negative linear anomalies, and P1 is a positive linear anomaly, VB is Vallis Bouvard.
filtered Bouguer and approximately -20 E in the gravity gradients, but the gravity anomaly is slightly offset relative to the topography, and it is not clear if and how they are related.

Although the most prominent topographic structures are not associated with obvious gravity anomalies, a number of radial gravity anomalies are observed with little or no topographic signature. Slightly southwest of Catena Leuschner is a positive radial gravity anomaly with no secondary crater chain associated with it (Fig. 3.6, Fig. 3.8A, D, G), which we will refer to as P1. This structure is highlighted in the Bouguer gravity, gravity gradients, and in the automated mapping of radial features (Fig. 3.1 B and C, Fig. 3.2). The average profiles reveal a clear gravitational signature of 10 mGal in the filtered Bouguer and -10 E in the gravity gradients (Figure 3.6). The topographic profile shows a slope associated with the linear anomaly, but it is not clear if this is significant.

Slightly to the southeast of Catena Leuschner is a very prominent negative ray, which for convenience we will call N1 here (Fig. 3.7, Fig. 3.8A, G, and D). This structure is highlighted in the Bouguer gravity, gravity gradients, and in the automated mapping of radial features (Fig. 3.1 B and C, and Fig. 3.2). The linear gravity anomaly does not have a clear topographic signature in map view, though there is a slight ridge of ~0.5 km in the average profile that does appear to rise above the background variability. This anomaly does have a very prominent signal in the Bouguer gravity of -20 mGal, and in the gravity gradients it has a 15 E signal. The fact that these linear gravity anomalies that do not have a clear topographic expression associated with them exist implies that the process that creates them cannot have happened only after the deposition of the Orientale ejecta blanket, but also during.

In summary, the general pattern of gravity anomalies in the surroundings of Orientale reveals radial gravity anomalies up to 300 km in length. Both positive and negative anomalies
are observed with typical magnitudes are in the range of ±10-20 mGal. Although some topographic features are associated with the linear gravity anomalies, most catenae are not associated with obvious gravity anomalies, and most radial gravity anomalies are not associated with catenae. Anomalies in the averaged profiles that are not associated with topographic structures are comparable in magnitude to the anomaly in Vallis Bouvard. Both the positive and negative anomalies analyzed that were not associated with known catena showed slightly elevated topography, consistent with deposition of material ejected from the basin or a simple local thickening of the ejecta.

3.2. Hydrocode models and interpretations

Before inverting the gravity data for the associated density anomalies, we first consider the ejecta emplacement and catenae formation process in order to shed light on the physical processes at work and guide our interpretations of the gravity data. The radial pattern of the gravity anomalies indicates that they are associated with the ejecta of Orientale. This observation suggests that the gravity anomalies may arise due to density anomalies associated with the ejecta itself or the effect of the ejecta on the underlying crust. Some of the radial gravity anomalies are suggested to be associated with catenae formed from the secondary impacts of ejecta material onto the surface of the ejecta blanket (Wilhelms 1976).

For assumed ejection angles of the material forming the crater chains between 35° and 55°, and the observed range in distances of 250 to 1000 km from the center of the basin, we calculate vertical ejecta impact velocities of 380-1000 m/s. These velocities are much lower than typical hyper-velocity impacts forming primary craters on the Moon of ~13 km/s (Oberbeck, 1975; Bottke et al., 2000). Most of the secondary crater chains are found to have diameters
ranging from 5 to 25 km, yielding impactor sizes between 0.5 and 7 km from the pi-scaling relationship (Holsapple, 1993).

Building on previous work (Milbury et al., 2015), a number of iSALE impact models were run using the above ranges in projectile velocity and diameter and assuming target porosities of 7, 12, and 17%, representing low, average, and high porosity targets, respectively. We average the predicted changes in density and porosity from the surface to a depth of 17 km in order to enable us to compare the magnitude of the density anomalies due to projectile material with those due to changes in porosity throughout the crustal column. Upon removing the effects of the impactor material left behind in the crater, we find that for low to medium porosity targets (initial target porosity lower than 17%), a small decrease in the vertically averaged target density (~3-5 kg/m³) is predicted beneath the secondary craters. If the initial target porosity is 17% or higher, the resulting averaged density contrast is positive (~1 kg/m³). This shows that a low velocity impactor will create more porosity in a low porosity target, but will compact a high porosity target. However, the changes in target density and porosity due to low velocity impactors are small (<1%), and unlikely to explain the observed radial gravity anomalies around Orientale. Thus, we focus the rest of this discussion on the effect of projectile density in models assuming 12% target porosity.

Since the impacts occur at low velocities, the impacting material does not disperse as happens in the case of hypervelocity impacts. Instead, the projectile material remains as a lens on the secondary crater floor. For a representative case of a 4-km-diameter impactor, a 12% target porosity, and an impact speed of 700 m/s, a thickness of 0.9 km of projectile material remains on the floor of a 15-km-diameter secondary crater (Fig. 3.9). For the scenario of high density material impacting a low density target, including the projectile material in the average density
Figure 3.9: A typical cross section through the hydrocode model results, for a 4 km impactor, with a 12% target porosity, and an impact speed of 700 m/s. The thickness of the remnant impact material is 0.9 km.

...change results in a large increase in average density beneath the secondary crater (~50 kg/m$^3$; Fig. 3.10, red line), which is much greater than the density contrast when the projectile material is excluded from the averaging (~3 kg/m$^3$; Fig. 3.10, magenta line). The same model run with an impactor density equal to the target density of 2400 kg/m$^3$ (Fig. 3.10, green line) results in a negligible decrease in the vertically averaged density (less than 0.1%).

The effect of the remnant impacting material within the secondary craters on the gravity will depend upon the density contrast between the impactor material and the target material, which for the case of the catenae and secondary craters formed on top of the ejecta blanket, are both ejecta of the Orientale basin. The secondary impactor material is extracted from the crust upon the original Orientale-forming impact, which would imply that the density of the secondary impactor material should be, on average, comparable to that of the ejecta blanket around Orientale (Wieczorek et al., 2013). Nevertheless, the fact that some of this material creates
Figure 3.10: Averaged density from the center of the basin outward for the case of a 4 km, impactor, hitting a target with 12% porosity at 700 m/s, including the remnant projectile material in the averaging. The density anomaly is averaged from the surface to a depth of 17 km.

Secondary crater chains and catenae on top of this ejecta blanket indicates some differences in the timing and ejection process of this material, and differences in density between the late-impacting secondary material and the ejecta blanket are possible. If there is no a priori reason to expect the impacting material to be more or less dense than the target, then either positive or negative anomalies could result.

Similarly, density variations are probable within the ejecta blanket itself, and might be expected to extend in radial patterns parallel to the motion of the ejecta. The existence of both positive and negative gravity anomalies and their common occurrence without secondary crater chains suggests that the majority of radial density anomalies represent the density variability of the ejecta combined with the dominantly radial structure of the ejecta blanket. As material is ejected from the crater, it gets stretched out and is deposited so that clumps of material of similar provenance may elongate on lines radial to the basin. For gravity anomalies due to both late impacts of secondary material associated with catenae and radial variations in the density of the
Figure 3.11: Optimal density models for the vertically-restricted density inversions (in kg/m$^3$). Top panels show a planer cross-section at 500 m depth. Bottom panels show vertical cross-sections at the location of the black lines in the top panels. Models are shown for areas 1 (left), 2 (center), and 3 (right). The arrow next to the area number indicates the direction of the basin center, and the areas correspond to the areas outlined in figure 3.1.

Ejecta blanket, the hydrocode models suggest that all substantial density anomalies should be vertically confined within the ejecta blanket itself. Radial gravity lineations extend from the Cordillera ring out to as far as 1000 km from the basin center, where the ejecta ranges in thickness from $\sim$3 km to 300 m (Fassett et al., 2011). We expect the thickness of the density anomalies to be in this range.

### 3.3 Inversion results

In order to better understand the subsurface structure of these linear anomalies, the gravity was inverted to obtain a model of the underlying density anomalies. Inverting the entire Orientale region is too computationally expensive, so we focus on 3 smaller areas, hereafter referred to as O1, O2, and O3 (Fig. 3.8). Area O1 (Fig. 3.8A, D, G) includes parts of Catenae Leuschner and Michelson, and also shows two prominent linear gravity anomalies, P1 and N1,
which are radial to Orientale but offset from Leuschner and Michelson (Fig. 3.1). Area O2 (Fig. 3.8B, E, H) does not show any clear secondary crater chains, but it does have a clear example of a negative linear gravity feature, N2. Area O3 (Fig. 3.8C, F, I) contains Vallis Bouvard (VB), which is associated with a positive gravity anomaly that is slightly offset from the valley (Fig. 3.5).

If a maximum depth of 60 km is assumed for the inversion model, then most of the density anomalies (including the linear gravity features) extend to a depth of ~20 km (Fig. 3.11). However, this maximum depth is likely limited by the sensitivity of the inversion model for gravity anomalies of a set wavelength, and does not preclude the existence of density anomalies at greater depths (Jansen et al., 2017). The density contrast predicted of ±40 kg/m$^3$ for the radial gravity anomalies is equivalent to a change in porosity by ±1.6% for a bulk density of 2550 kg/m$^3$, which is well within the range of porosities observed in lunar samples (Kiefer et al., 2012a). However, based on the results of the hydrocode models and the expectation that the density anomalies be contained within the Orientale ejecta, the density contrast expected from these linear gravity anomalies should be constrained to the upper ~2 km. This disparity is likely due to the smoothness of the density inversion models. The “optimal” model was based on the tradeoff between fitting the data correctly and finding a geologically likely model, which was assumed to be smooth. However, due to the non-uniqueness of inversions of gravity data, a broad low amplitude density contrast at the surface would be virtually indistinguishable from a narrower high amplitude density anomaly distributed over a greater range of depths. By assuming a smooth model, the gravity inversion favors the latter, while the hydrocode models
prove the former to be the more likely explanation. Therefore, we now constrain the inversion model to place the density anomalies within the top 2 km of the crust.

For the density inversions constrained to shallow depths, the optimal model based on the L-curve approach indicates that $\beta = 10^{-2}$ is the model with the best trade-off between data fit and model smoothness. However, this model forces the anomalies to be much sharper and narrower, in contrast with the gravity expression. Moreover, the predicted density anomalies are too large in magnitude ($>1000 \text{ kg/m}^3$), conflicting with a priori information from the range of densities from lunar samples (Kiefer et al., 2012a) and previous gravity analyses (Wieczorek et al., 2013). Therefore, we choose to use the model corresponding to $\beta = 1$, which provides a smoother model with a more reasonable density contrast.

The density inversions of these three areas (Fig. 3.11) show density anomalies roughly in line with the orientation of the Bouguer gravity anomalies. Note that the cross-sections appear to be columnar in the figures, but this is because of the extreme stretching of the vertical axis in the figures. Because the vertical extent of the model domain is less than the wavelength of the gravity anomalies, the model cannot resolve variations in density with depth. The negative linear gravity anomaly (Fig. 3.7 and N1 in Fig. 3.8) shows up very clearly in the density model. An average profile perpendicular to the lineation shows that anomaly N1 has a negative density

Figure 3.12: Average density profiles across anomaly N1 and Vallis Bouvard from figure 3.11.
contrast of ~350 kg/m$^3$ (Fig. 3.12). This density contrast corresponds to an increase in porosity by ~13.7\% assuming a 2550 kg/m$^3$ bulk density. Anomaly P1 yields a positive density contrast of 200 kg/m$^3$, or a decrease in porosity by 7.8\%. Anomaly N2 in area O2 results in a negative density contrast of ~250 kg/m$^3$, or a porosity increase by 9.8\%. This range in density contrast and porosity are still within what we observe from lunar samples (Kiefer et al., 2012a).

Performing the same analysis on the linear gravity anomaly associated with Vallis Bouvard (Fig. 3.12) yields a density contrast of ~250 kg/m$^3$, or a decrease in porosity by ~9.8\%. Since the density does not vary with depth, the density contrast will scale with the inverse of the assumed thickness, and thus we can easily account for thinner density anomalies as might arise due to material lining the floors of secondary crater chains.

Due to the non-uniqueness of the inversion model, a simpler analytical model was also tested in which we calculate the gravity resulting from a rectangular prism density anomaly and compare it to the observed average anomalies to find the best-fit width and density contrast for a range of thicknesses. This model approximates the effects of a discrete lens of material on the floor of a chain of secondary craters, or a discrete zone of material embedded within the Orientale ejecta blanket. Thus, for simplicity we assume an anomaly thickness similar to that of the hydrocode model (Fig. 3.9) of 0.9 km and solve for the best-fit width and density contrast of the anomaly (Fig. 3.13).

For anomaly N1, the lowest RMSE is 2.34 mGal corresponding to a density contrast of -410 kg/m$^3$ and a width of 19.5 km. This density anomaly is slightly more negative than the peak anomaly in the cross section of the density inversion result constrained to the top 2 km (Fig. 3.12), but this model is restricted to a narrower range of depths and maintains that density anomaly over a greater range of widths. For an anomaly thickness of 1.75 km thick,
corresponding to the thickest lens of projectile material predicted by the hydrocode models, we obtain a minimum RMSE of 2.34 mGal for a density contrast of -220 kg/m$^3$ and a width of 19.0 km. For Vallis Bouvard the lowest RMSE is 2.90 mGal for a model with a density contrast of 425 kg/m$^3$ and a width of 17.5 km. For an anomaly thickness of 1.75 km, we obtain a minimum RMSE = 2.65 mGal for a model with a density contrast of 230 kg/m$^3$ and a width of 17.5 km. Though these values given here are for the lowest RMSE, there are a wide range of models that can fit the data nearly as well. Thicker and thinner anomaly source regions would result in lower and higher density anomalies than stated above, respectively.

4. Conclusions

The high-resolution GRAIL gravity data reveal that the Moon is far more complex than previously imagined. Previous work has shown that the density of the crust varies on small scales (10’s of km; Jansen et al., 2017). Though some of these small-scale density variations appear to
be the result of random changes in porosity (Jansen et al., 2017), some have a more defined source, such as the small-scale gravity anomalies radiating out from the Orientale Basin. These radial gravity lineations are sometimes associated with secondary crater chains or catenae, for which the specific details of formation and their relation to the basin ejecta remain unclear (Wilhelms et al., 1978). Here we investigated these radial gravity structures and the secondary crater chains in more detail by using gravity observations, inversions, and hydrocode modeling. Perpendicular profiles of the topography, filtered Bouguer gravity, and gravity gradients over several anomalies such as Catena Leuschner, Catena Michelson, Vallis Bouvard, and unnamed negative and positive linear gravity anomalies show that radial features seen in topography do not always have gravity signatures associated with them that clearly stand out from the background noise. Similarly, those features that have a clear signature in the gravity do not always show any surface expression. The widths of the radial gravity lineations not associated with catenae are similar to the widths of the catenae.

Since the secondary craters and catena form in low-velocity impacts, hydrocode modeling was used to constrain the cause of the density differences. It was found that the secondary impacts have a negligible effect on the density of the target material. However, the effect of remnant impact material left behind within the secondary craters and catenae could explain the density differences if the density of the impactor is significantly different from the target. Thus, for radial gravity anomalies associated with catenae, secondary projectile material remaining on the floors of the secondary craters is the most likely source of the anomalies. For radial gravity anomalies not associated with catenae, we conclude that the radial orientation of both positive and negative gravity anomalies likely represents the variability of the ejecta combined with the radial structure of the ejecta blanket.
In order to better understand the subsurface structure of these linear anomalies, the gravity was inverted to obtain a model of the underlying density anomalies. If a maximum depth comparable to the crustal thickness of 60 km is assumed in the inversion model then most of the density anomalies (including the linear gravity features) extend to a depth of ~20 km with a density contrast of ±40 kg/m\(^3\), which is equivalent to a change in porosity by ±1.6% for a bulk density of 2550 kg/m\(^3\). However, considering the hydrocode model results the density contrast expected from these linear gravity anomalies should be constrained to be within the ejecta blanket in the upper ~2 km of the crust. The solution to the shallow-depth inversions show density anomalies in the range of ±250-350 kg/m\(^3\), corresponding to changes in porosity of ±10-14%. Both of these ranges are within the observed porosity range of the lunar samples (Kiefer et al., 2012a). Forward modeling a rectangular prism with assumed density and depth shows that we can get a very close fit to the data for anomaly widths of ~20 km and density contrasts of approximately ±200-400 kg/m\(^3\) for an assumed thickness of 0.9-1.75 km.

Given the porosity of the Orientale ejecta blanket of ~14-18% (Wieczorek et al., 2013), the positive radial density anomalies would correspond to material with porosities approaching 0 while the negative radial density anomalies would correspond to material with porosities approaching 30%. The lowest porosities may correspond to material excavated from deeper within the crust where ductile deformation has collapsed the pore space (Besserer et al., 2014), or melted material that crystalized with little porosity. Either interpretation may reveal important details about the excavation and ejecta deposition process, with possible implications for the formation of catenae as well.

While some radial gravity anomalies are clearly visible in the filtered gravity and gravity gradients, many of the anomalies are relatively subtle features. The radial gravity anomalies are
comparable in magnitude in both Bouguer gravity and gravity gradients to the random small-scale background variation in the gravity field that is pervasive across all highland surfaces (Jansen et al., 2017). The radial anomalies only become apparent where they rise above this background variability. Thus, it is possible that a radial texture in the gravity field pervades the ejecta blanket, even where it is not readily observable.

Although a preliminary survey reveals some evidence for radial gravity anomalies around other basins, they are generally less prominent than those around Orientale. This is perhaps not surprising, given that Orientale is the youngest and best-preserved large multi-ring impact basin. Orientale is unique in the roughness (Kreslavsky and Head, 2012) and low density (Wieczorek et al., 2013) of its ejecta blanket. Other older basins likely possessed similar ejecta blanket properties earlier, but the density and roughness of those ejecta blankets gradually became less distinct from their surroundings due to the effects of seismic shaking and impact gardening (Kreslavsky and Head, 2012). Similarly, the radial gravity anomalies arising due to density variations within the ejecta blankets of older basins may become less distinct over time. Even where the radial patterns are not distinct, it is likely that these substantial discrete variations in the density of basin ejecta contribute substantially to the small-scale density anomalies that pervade the upper crust of the Moon (Jansen et al., 2017).
REFERENCES


Andrews-Hanna, J. C., 2018???


APPENDIX A

PERMISSION FROM CO-AUTHORS

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I am delighted to grant permission to use the paper for your dissertation. You did almost all of the work and writing.

Jeff

On 1/8/2018 12:07 PM, Johanna Jansen wrote:

Dear co-authors,

I hope you are all well.

It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.

Thank you very much,

JJ

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G. Jeffrey Taylor
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Permission granted! Good luck! Jim

On Mon, Jan 8, 2018 at 5:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:

Dear co-authors,

I hope you are all well.

It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.

Thank you very much,

JJ

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Google Scholar Link https://scholar.google.com/citations?user=GjULCCQAAAAJ&hl=en
Brown Planetary Geoscience website http://www.planetary.brown.edu

Office: 104 Lincoln Field Building
You are so permitted. Congratulations and good luck.

On 2018-01-08 12:07, Johanna Jansen wrote:

Dear co-authors,
I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.
Thank you very much,
JJ

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Paul G. Lucsey
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Honolulu HI 96822

Want more info? Google me at: lucsey moon

From: Kiefer, Walter <kiefer@lpi.usra.edu>
Subject: RE: Permission
Date: January 8, 2018 at 3:08 PM
To: Johanna Jansen, jeff_andrews-hanna@jhuapl.gatech.edu, yaoguo.li@mines.edu, Paul Lucsey lucsey@higo.hawaii.edu, Jeff Taylor pbtaylor@higo.hawaii.edu, Sander Goossens sander.j.goossens@nasa.gov, Frank Lemoine frank.g.lemoine@nasa.gov, Erwan Mazariro erwan.m.mazariro@nasa.gov, Jim Head james_head@brown.edu, crmbury@gmail.com, Jason Soderblom jms4@mit.edu, Maria Zuber zuber@mit.edu
Cc: Kiefer, Walter kiefer@lpi.usra.edu

JJ,

I am happy to give my permission. Good luck with the defense.

Walter Kiefer
Senior Staff Scientist
Lunar and Planetary Institute

----- Original Message -----
From: Johanna Jansen [mailto:joan.c.jansen@gmail.com]
Sent: Monday, January 08, 2018 4:07 PM
To: Jeff Andrews-Hanna; Yaoguo Li; Paul Lucsey; Jeff Taylor; Sander Goossens; Frank Lemoine; Erwan Mazariro; Jim Head; crmbury@gmail.com; Kiefer, Walter; Jason Soderblom; Maria Zuber
Subject: Permission

Dear co-authors,

I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.
Thank you very much,

JJ
You have mine. I haven’t read it but doubt that I’d have edits.

On Jan 8, 2018, at 12:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:

Dear co-authors,

I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.
Thank you very much.

JJ

Hoi JJ,

Good te horen dat je binnenkort je proefschrift kan verdedigen. Veel succes!

En uiteraard stem ik ook toe om de paper op te nemen in je proefschrift.

Of course I give you my permission as a co-author on your paper to include the paper in your thesis.

Groeten,

Sander.

On Jan 8, 2018, at 5:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:

Dear co-authors,

I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.
Thank you very much.

JJ
JJ—
This is fine with me, good luck!
Maria

On Jan 8, 2018, at 5:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:
Dear co-authors,
I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.
Thank you very much,
JJ

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http://www-geodyn.mit.edu

Not all those who wander are lost.
-- J.R.R. Tolkien

From: Yaoguo Li ygl@mines.edu
Subject: Re: Permission
Date: January 8, 2018 at 3:18 PM
To: Johanna Jansen joan.c.jansen@gmail.com
Cc: Jeff Andrews-Hanna jahanna@gmail.com, Paul Lucey lucey@hgp.hawaii.edu, Jeff Taylor gtaylor@hgp.hawaii.edu, Sander Goossens sander.j.goossens@nasa.gov, Frank Lemouine frank.g.lemouine@nasa.gov, Erwan Mazariro erwan.m.mazariro@nasa.gov, Jim Head james_head@brown.edu, cmibury@gmail.com, Walter Kiefer kiefer@lpi.usra.edu, Jason Soderblom jms4@mit.edu, Maria Zuber zuber@mit.edu

JJ,
Yes, you have my permission!

Cheers,
Yaoguo

Sent from my iPhone
(Please excuse brevity/types)

On Jan 8, 2018, at 3:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:
Dear co-authors,
I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.
Thank you very much,
JJ
Yes, you have my permission.

Jeff Andrews-Hanna
Associate Professor
Lunar and Planetary Laboratory
University of Arizona
Tucson, AZ 85721-0922

Kupfer room 438
https://www.lpi.arizona.edu/faculty/jeffrey-andrews-hanna

On Jan 8, 2018, at 3:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:

Dear co-authors,

I hope you are all well.
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Thank you very much,
JJ

Congrats!!

You absolutely have my permission.

Best,
Jason

On Jan 8, 2018, at 5:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:

Dear co-authors,

I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.
Thank you very much,
JJ
From: Lemoine, Frank G. (GSFC-61A0) frank.g.lemoine@nasa.gov
Subject: RE: Permission
Date: January 8, 2018 at 3:25 PM
To: Johanna Jansen joan.c.jansen@gmail.com

Dear Johanna,

I'm happy to your concurrence and permission to adding this published paper to your thesis. Good luck with your thesis defense. Do you know when it will be scheduled?

Regards,
Frank L.

De: Johanna Jansen [joan.c.jansen@gmail.com]
À: Jeff Andrews-Hanna; Yaoguo Li; Paul Lucey; Jeff Taylor; Goossens; Sander (GSFC-6986.0)[UNIVERSITY OF MARYLAND BALTIMORE CO]; Lemoine, Frank G. (GSFC-61A0); Mazarico; Erwan (GSFC-6980); Jim Head; cmilibury@gmail.com; Walter Kiefer; Soderblom, Jason M (GSFC-4320)[CORNELL UNIVERSITY]; Maria Zuber
Objet: Permission

Dear co-authors,

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Thank you very much,

JJ

From: Mazarico, Erwan (GSFC-6980) erwan.m.mazarico@nasa.gov
Subject: Re: Permission
Date: January 8, 2018 at 4:09 PM
To: Johanna Jansen joan.c.jansen@gmail.com

Hello JJ,

You have my permission as well of course.
Best of luck with the defense!

Erwan

On Jan 8, 2018, at 5:07 PM, Johanna Jansen <joan.c.jansen@gmail.com> wrote:

Dear co-authors,

I hope you are all well.
It has been a while since we published the "small-scale density variations in the lunar crust as revealed by GRAIL" paper. It is time for me to defend, and my school requires permission from co-authors before adding a published paper to my thesis. If you would all be so kind as to send me your permission by Friday, I would be very grateful.

Thank you very much,

JJ