Geology and petrology of enormous volumes of impact melt on the Moon: A case study of the Orientale basin impact melt sea

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Lunar basin-forming impacts produce enormous volumes (>10⁵ km³) of impact melt. All known basin-forming impacts combined may produce ~10⁸ km³ of impact melt, ~1/20th the volume of the lunar crust. Despite their volumetric importance, the geology and petrology of massive deposits of impact melt on the Moon have been little studied, in part because most basin impact melt deposits are old and have been obscured or buried by subsequent impact cratering and mare infill. We investigate the geology and model the petrology of fresh massive impact melt deposits in the relatively young 930 km diameter Orientale basin. Models of impact melt production combined with geologic analyses based on new LOLA topographic data suggest that most of the impact melt (~2/3) produced by the Orientale-forming impact in a ~15 km thick impact melt sheet (better described as an impact melt sea) ~350 km in diameter with a volume of ~10⁸ km³. We anticipate that the Orientale melt sea has undergone large-scale igneous differentiation, since terrestrial impact melt sheets (such as Manicouagan, Sudbury, and Morokweng) less than a tenth of the thickness and a hundredth of the volume of the Orientale melt sea have differentiated. We develop a model for the cumulative stratigraphy of the solidified Orientale impact melt sea. A modeled cumulate stratigraphy (occurring below a quench crust and anorthositic fallback breccia) with an ~8 km thick layer of norite overlying a ~4 km layer of pyroxenite and a basal ~2 km thick layer of dunite produced by equilibrium crystallization of a homogenized melt sea, consistent with vigorous convection in that melt sea, is supported by remotely-sensed norite excavated by the central peak of Maunder crater from ~4 km depth. Generally, we predict that very large basin-forming impacts, including the South Pole-Aitken (SPA) basin-forming impact, produce melt seas with a cumulate stratigraphy similar to that of the Orientale melt sea. Impact melt differentiation may explain apparently anomalous lithologies excavated in the SPA basin interior. We note that impact melt differentiates are slow-cooled (the Orientale melt sea took on the order of 10⁵ years to solidify) and, if meteoritic siderophiles fractionate into metal or sulfide layers, may not be siderophile-enriched; therefore, impact melt differentiates may pass for pristine highland plutonic rocks in the lunar sample suite. These predictions can be tested with current and future mission data.

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1. Introduction

Simple and complex lunar craters <150 km in diameter contain relatively small volumes (<10⁴ km³) of impact melt (Cintala and Grieve, 1998). The geology of the small impact melt deposits in these craters is relatively well understood: impact melt floods crater floors and drapes crater central peaks, walls, and rims (Bray et al., 2010; Howard et al., 1974; Howard and Wilshire, 1973, 1975; Plescia and Cintala, 2012; Spray and Thompson, 2008); melt distribution is controlled by impact angle, post-impact ejection related to slumping, and pre-existing topography (Hawke and Head, 1977). Thin sheets of impact melt occurring in these craters freeze quickly (Warren et al., 1996) to produce glasses or cryptocrystalline lithologies chemically similar to target rocks.

Field and experimental evidence shows that the volume of shock melt produced by an impact scales approximately as the fourth power of impact crater transient cavity diameter (Abramov et al., 2012; Cintala and Grieve, 1998; Grieve and Cintala, 1992, 1997). So although craters <150 km in diameter contain relatively small volumes (<10⁴ km³) of impact melt, basin-forming impacts creating basins >300 km in diameter are predicted by these analytic scaling relationships (Abramov et al., 2012; Cintala and Grieve, 1998; Grieve and Cintala, 1992) as well as hydrocode models (Collins et al., 2002; Ivanov, 2005) to produce enormous volumes (>10⁷ km³) of impact melt. The volumes of impact melt produced by basin-forming impacts are well in excess of the volumes of many terrestrial bodies of water and igneous intrusions (Fig. 1). In fact, all
basin-forming impacts combined are modeled to produce on the order of \(10^9 \text{ km}^3\) of impact melt (see Kring et al. [2012] and our Section 4.4), so that as much as \(1/20\)th of the lunar crust may be impact melt.

Despite their volumetric importance, the geology and petrology of massive impact melt deposits produced by basin-forming impacts are poorly understood. This is in part because basin interiors where massive impact melt deposits occur (Head, 1974; Melosh, 1989; Spudis, 1993) have been obscured and modified by impact cratering and mare basalt infill (Head and Wilson, 1992). For example, no impact melt deposits are exposed in the mare basalt-flooded Imbrium, Serenitatis, and Crisium basin interiors (Head et al., 1993). Moreover, although the thick (>1 km) sheets of impact melt predicted to occur in lunar basins (Cintala and Grieve, 1998) cool slowly and may undergo igneous differentiation (Cintala and Grieve, 1998; Grieve et al., 1991), the cumulate stratigraphy of these thick impact melt sheets has never been modeled, although several authors have carefully considered some particular petrologic consequences of impact melt differentiation (Morrison, 1998; Warren et al., 1996).

Where on the Moon should we look to improve our understanding of massive impact melt deposits? The relatively young (\(3.7\)–\(3.8 \text{ Ga}\)) (Baldwin, 1974; Whitten et al., 2011) Orientale basin (Head, 1974; Howard et al., 1974; McCauley, 1977), which has relatively few impact craters and little mare basalt infill (Greeley et al., 1993; Whitten et al., 2010, 2011), is an ideal location to investigate the distribution, thickness, and volume of massive impact melt deposits. Moreover, the relatively well-constrained structure of the farside crust (Wieczorek et al., 2006) and upper mantle (Khan et al., 2006) probably melted by the Orientale impact provide necessary compositional data (Shearer et al., 2006) to model the cumulate stratigraphy of an Orientale impact melt sheet and to consider the petrologic implications of impact melt differentiation.

In this paper, we (1) investigate the distribution, thickness, and volume of impact melt deposits produced by the Orientale-forming impact using models of impact melt production combined with recent topographic data and (2) model the cumulate stratigraphy of a solidified, differentiated Orientale melt sheet, considering the general implications for the lunar highlands crust and the lunar sample suite.

2. Distribution, thickness, and volume of massive impact melt deposits in Orientale

2.1. General geology of the Orientale basin

The deposits associated with the \(930 \text{ km diameter Orientale impact basin (Fig. 2) have been mapped and subdivided into three units (Fassett et al., 2011; Head, 1974; Howard et al., 1974; McCauley, 1977): the chaotically-textured Hevelius Formation, lying outside the \(930 \text{ km diameter Cordilleran Ring; the Montes Rook Formation, comprising knobby terrain between the Cordilleran Ring and the \(620 \text{ km diameter Outer Rook Ring; and the Maunder Formation, which occurs inside the Outer Rook Ring, around the \(480 \text{ km diameter Inner Rook Ring, and on the floor of the \(350 \text{ km diameter central depression. These three major rings (the Inner Rook Ring, the Outer Rook Ring, and the Cordilleran Ring) have each been interpreted as the Orientale basin crater rim crest (see Spudis (1993) for a review). We interpret the Orientale basin as a modified peak-ring basin (Head, 2010), taking the Inner Rook Ring to be the basin peak ring (Baker et al., 2011a, 2011b, 2012) and the Outer Rook Ring to be the basin rim crest (Head, 1974, 2010; Head et al., 1993). In the following, we take the current diameter of the Outer Rook Ring (620 km) as the best estimate of the Orientale transient cavity diameter, since there is no clear evidence for terracing within this rim that would have substantially increased its diameter (Head, 2010; Head, 2012).

2.2. Geology of impact melt-related facies in Orientale

The facies of the Maunder Formation (Fig. 3), named for the prominent 55 km diameter complex crater Maunder superposed on the north edge of the central depression of the Orientale basin, have historically been interpreted as impact melt-related (Head, 1974; Head et al., 1993; Howard et al., 1974; Spudis, 1993). The central smooth facies (Fig. 3, left), occurring within the inner depression of the Orientale basin exposed through thin mare fill (Head, 2010; Whitten et al., 2010, 2011), has been interpreted as pure impact melt (Head, 1974; Howard et al., 1974; McCauley, 1977). Wrinkle ridges (Fig. 3), fractures, and polygonal cracks are apparent in the smooth facies and under the overlying mare. At an average distance of 175 km from the Orientale basin center (Fig. 2), the topography abruptly rises \(7.5 \text{ km} \) along a series of marginal normal faults to a corrugated and fissured rough facies (Fig. 3, right). This outer corrugated, fissured facies has been interpreted as clast-rich impact melt formed from mixing of pure impact melt with brecciated debris from the Orientale impact (Head, 1974; Head et al., 1993; Howard et al., 1974; McCauley, 1977).

2.3. How much impact melt was produced by the Orientale-forming impact?

Impact melt volume scaling relationships adjusted for lunar gravity derived by Cintala and Grieve (1998) predict that a chondritic projectile with an asteroidal velocity of 10 km/s and a trajectory normal to the lunar surface giving rise to an Orientale-sized transient cavity \(620 \text{ km in diameter produces } \approx \text{6 } \times \text{10}^6 \text{ km}^3\) of impact melt. This volume is probably an overestimate: many authors have suggested that the Orientale-forming impact was oblique (Schultz and Papamarcos, 2010; Scott et al., 1977; Wilhelms et al., 1987) and oblique impacts are known to produce less melt than vertical impacts of the same size (Pierazzo and Melosh, 2000). Transverse widening of the Cordilleran Ring of Orientale (Schultz et al., 2012) may indicate that the Orientale impactor had an impact angle of 15–30°. The volume of melt produced by an impact with an impact angle of 15–30° is 10–50% of the volume of melt produced by a vertical impact (Pierazzo and Melosh, 2000); taking 22.5°, the center of this impact angle range, to be the most likely impact angle of the Orientale-forming projectile in the absence of additional constraints, we estimate that the Orientale-forming impact produced \(1.5 \times 10^6 \text{ km}^3 (25 \text{ vol.}%) \) of impact melt. We emphasize that the composition and velocity of the Orientale-forming projectile are not well determined.
constrained: variations in these parameters can change the impact melt volume by a factor of two in the Cintala and Grieve (1998) scaling model.

An independent estimate of the volume of impact melt produced by the Orientale-forming impact can be made using the impact melt volume scaling relationship for oblique impacts derived by Abramov et al. (2012). This relationship, which explicitly includes impact angle as a variable, predicts that a chondritic projectile with an impact angle of 22.5° giving rise to an Orientale-sized transient cavity 620 km in diameter produces about $2.8 \times 10^6$ km$^3$ of melt, about twice the volume estimated above by coupling the models of Cintala and Grieve (1998) and Pierazzo and Melosh (2000). Discrepancies between impact melt volumes predicted by the scaling relationships of Cintala and Grieve (1998) and Abramov et al. (2012) have previously been noted by Abramov et al. (2012). In order to assess the Cintala and Grieve (1998) and Abramov et al. (2012) estimates of the volume of impact melt produced by the Orientale-forming impact, we later compare melt deposit thicknesses determined from these calculated melt volumes (Section 2.4) to an independent determination of melt deposit thickness (Sections 2.5–2.7). For simplicity, in the following section we consider only the melt volume of $1.5 \times 10^6$ km$^3$ calculated from the scaling model of Cintala and Grieve (1998).

2.4. Accounting for the $\sim 1.5 \times 10^6$ km$^3$ of impact melt produced by the Orientale-forming impact

Where does the estimated $\sim 1.5 \times 10^6$ km$^3$ of impact melt produced by the Orientale-forming impact end up? A priori, because the surface area of the Moon is only $\sim 3.8 \times 10^7$ km$^2$, and since there is no evidence that the Moon is covered in a $\sim 40$ m (i.e.,...
1.5 × 10^6 km³ (3.8 × 10^7 km²) thick layer of Orientale impact melt, most of the impact melt produced by the Orientale-forming impact must occur in or near the Orientale basin.

This is in accordance with what is known about the process of impact melt generation (Fig. 4). Impact melt forms in a melt cavity interior to (although, in multi-ring basins, likely extending below) what will become the crater transient cavity (Fig. 4a) (Cintala and Grieve, 1998). Some of the generated impact melt is excavated in forming the transient cavity; additional melt is ejected from the crater interior during modification (Fig. 4b) (Hawke and Head, 1977; Osinski et al., 2011). Studies of large terrestrial craters such as Sudbury demonstrate that a relatively small proportion of melt (<10% by volume) is ejected from the crater interior during modification (Grieve and Cintala, 1992). Excavation flow modeling suggests that up to 70–80 vol.% of impact melt remains inside the transient cavity in large lunar craters (Cintala and Grieve, 1998). Based on these models, we estimate that ∼25 vol.% of impact melt was ejected from the Orientale basin (perhaps incorporated in Hevelius Formation ejecta). This implies that ∼1.1 × 10^6 km³ (75 vol.% of 1.5 × 10^6 km³) of impact melt produced in the Orientale-forming impact remains inside the Orientale transient cavity rim, here taken to be the Outer Rook Ring (Head, 1974, 2010, 2012).

Occurring inside the Outer Rook Ring are the impact melt-related rough and smooth facies (Fig. 3) of the Maunder Formation described in Section 2.2 above. About 1.1 × 10^6 km³ of impact melt must be apportioned between these two facies. How much impact melt is contained in each facies?

The rough facies mantles topography similar to the peaks of the Inner Rook Ring (Head, 1974), which stand up to 2–3 km above surrounding terrain (Figs. 2 and 3). This constrains the volume of impact melt contained in this facies: the relief of the mantled topography is <1 km (Fig. 5), but the original relief was likely no greater than the relief of the nearby peaks of the Inner Rook Ring, 2–3 km; therefore, the mantling deposit is ∼2 km thick at most. The area of the rough facies—approximately the area of the annulus between the edge of the central depression (175 km from the basin center) and the Outer Rook Ring (310 km from the basin center)—is ∼2 × 10^5 km². If the clast-rich rough facies is 2 km thick (which for 50 vol.% clasts is the equivalent of a 1 km thick layer of pure impact melt) it contains ∼1 × 10^5 km³ (50 vol.% pure melt × 2 km × ∼2 × 10^5 km²) of impact melt. The remaining ∼1.0 × 10^6 km³ (∼1.1 × 10^6 km³ – ∼1 × 10^5 km³) of impact melt must be contained in the central smooth facies, thought to be a pure impact melt sheet (Section 2.2). This smooth facies is 175 km in radius (Fig. 5) or ∼10^5 km² in area, so the impact melt

![Fig. 4. The material melted by the Orientale-forming impact (a), the Orientale melt cavity and transient cavity geometry immediately following hypervelocity impact (b), and the final crater geometry of Orientale (c) (Andrews-Hanna and Stewart, 2011; Hikida and Wieczorek, 2007) superposed on the lunar highlands crust and upper mantle (Khan et al., 2006; Wieczorek et al., 2006; Zuber et al., 1994). Melt fills a truncated sphere with radius/truncation height taken as 0.9/1.1, following Barr and Citron, (2011), and a total volume of 1.5 × 10^6 km³ (Cintala and Grieve, 1998), following our discussion in Sections 2.3–2.7. The basin transient cavity depth-diameter ratio is taken to be 1/10 (Wieczorek and Phillips, 1999) and the transient cavity diameter has been decreased by a factor of ½ for clarity. The excavation flow curve (in blue) is given by the Maxwell Z-model taking Z = 3 (Melosh, 1989). Note that (1) the transient cavity is gravitationally unstable, so the geometry shown in (b) is never fully realized at any time and (2) the final basin diameter (Fig. 2) is much larger than the transient cavity diameter in (b). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)](image-url)
sheet must be on the order of 10 km thick to account for the remaining \( \sim 1.0 \times 10^8 \) km\(^3\) of impact melt.

A \( \sim 10 \) km thick impact melt sheet is not necessarily unrealistic: the Sudbury Igneous Complex, a terrestrial impact melt sheet occurring in a \( \sim 250 \) km diameter crater (rim diameter 140 km), is \( \sim 2.5–3 \) km thick (Thierrault et al., 2002). More compellingly, there is independent geologic evidence for a \( \sim 10 \) km thick impact melt sheet occurring in the Orientale basin’s central depression.

### 2.5. Geologic evidence for a \( \sim 10 \) km thick melt sheet

As noted in Section 2.2, the surface of the smooth facies is on average offset \( \sim 1.75 \) km (Fig. 5) below the rough facies along marginal normal faults. This offset has been previously explained as vertical subsidence caused by thermal stresses resulting from impact-generated heat and uplift of crustal isotherms (Bratt et al., 1985). This model predicts a \( \sim 2 \) km drop over 100 km of radial distance (Fig. 14 from Bratt et al. (1985)). However, new, high-resolution LOLA altimetry (Smith et al., 2010) shows that the vertical subsidence of the central depression is abrupt: the topography drops \( \sim 2 \) km over a radial distance of \( \sim 20 \) km along the west edge of the depression (Fig. 6), five times more steeply than predicted by the model of Bratt et al. (1985).

Wilson and Head (2011) suggest that the vertical subsidence of the smooth facies, an impact melt sheet, is largely the result of vertical shrinkage upon solidification and cooling of impact melt. There is evidence for lateral shrinkage due to the same process: fractures on the smooth facies resemble deformations on the surfaces of terrestrial lava lakes (Barberi and Varet, 1970) known to form due to lateral contraction upon solidification and cooling. We now calculate the amount of vertical subsidence that results from solidification and cooling of impact melt sheets on the Moon in order to constrain the thickness and volume of the Orientale melt sheet.

### 2.6. Estimating melt sheet thickness from vertical subsidence

The vertical shrinkage of a melt sheet upon solidification and cooling is related to its total thickness. We now determine the nature of this relationship in order to calculate the thickness of the Orientale melt sheet from its observed vertical subsidence. The derivation follows Wilson and Head (2011).

A melt sheet emplaced at a temperature \( T_e \) that solidifies and cools to the ambient temperature \( T_c \) will decrease in volume both during solidification and while cooling to \( T_c \). What is the magnitude of the total proportional decrease in volume? Volume is inversely proportional to density, so \( \frac{dV}{V} = -\frac{dp}{p} \). A simple rule of thumb is that the density of silicate melt increases (and volume of melt decreases) by \( \sim 10\% \) upon solidification. This rule may be misleading for melts of the lunar crust, since the density of anorthite melt increases only \( \sim 5.5\% \) upon solidification (based on calculations of silicate liquid density following Nelson and Carmichael (1979)); however, as will be shown in Section 3.3, the Orientale melt sheet contains abundant mafic material, so we continue to use the \( \sim 10\% \) estimate. The proportional change in volume of a material at constant pressure, \( \frac{dV}{V} \), due to subsequent
cooling is equal to \(\alpha \frac{dT}{dt}\), where \(\alpha\) is the volumetric coefficient of thermal expansion, \(-3 \times 10^{-5}\) K\(^{-1}\), for silicate rock (Turcotte and Schubert, 2002). The \(dT\) is 1328 K, the difference between \(T_s\) (taken to be 1568 K, the average of the lava liquidus, 1713 K, and solidus, 1423 K, temperatures suggested by Williams et al. (2000)) and \(T_c\) (taken to be the average of the lunar day and night temperatures, \(-240\) K); therefore, the volume of the melt sheet must decrease by \(-4\%\) upon cooling. A \(-10\%\) volume decrease compounded with a \(-4\%\) volume decrease results in approximately a \(-14\%\) volume decrease, since these percentages are both small.

What portion of this \(-14\%\) volume decrease results in vertical subsidence as opposed to lateral shrinkage? If the melt sheet contracts equally in all directions, the percentage thickness decrease will be one-third of the percentage volume decrease, about \(-4.7\%\). However, this is physically and geologically unrealistic: the melt sheet does not cool at an equal rate at each of its boundaries, but cools primarily through radiation at its top, and will therefore not contract equally in all directions; moreover, the contraction features (cracks and depressed polygons) present in the S and SW portion of the smooth facies occupy only 1–2\% of the basin floor and are not consistent with a 9.3\% decrease in surface area. A more detailed treatment of lava lake solidification and cooling (Worster et al., 1990, 1993) gives a better estimate of vertical subsidence. The upper surface of a lava lake cools by radiation and loses heat much faster than the other boundaries of the lava lake. Initially, lava exposed to the surface solidifies to form plates; these plates must undergo mainly vertical contraction, since lateral contraction would cause lava to well up and the solidified plates to founder into the liquid beneath, settling toward the base of the pond and probably being partially remelted. Eventually a stable crust will form; evidence from drilling of the Kilauea Iki lava lake pond (Hardee, 1980) confirms that the stable crust forms well before crust will form; evidence from drilling of the Kilauea Iki lava lake pond and probably being partially remelted. Eventually a stable traction would cause lava to well up and the solidified plates to

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In short, we anticipate that nearly all the \(-10\%\) volume decrease upon solidification results in vertical subsidence and that between one-third and all of the \(-4\%\) volume decrease upon cooling results in vertical subsidence. Therefore, we predict that the vertical shrinkage of a lunar melt sheet is between \(-11.3\%\) and \(-14\%\) of its depth. Fig. 7 shows the subsidence predicted as a function of melt sheet depth for these \(11.3\%\) and \(14\%\) limits.

2.7. Thickness and volume of the Orientale melt sheet

The Orientale melt sheet has an observed average vertical relief of \(-1.75\) km (Fig. 5). As stated above, the observed vertical subsidence is a factor of five greater than predicted by Bratt et al. (1985), so subsidence due to solidification and cooling of melt must dwarf subsidence due to other thermal stresses. Therefore, we treat all the observed vertical subsidence as the result of melt solidification and cooling. The \(11.3\%\) and \(14\%\) limits due to cooling and contraction derived above imply that the melt sheet was initially between \(-12.5\) and \(-15.5\) km thick (\(-1.75\) km/\(14\%\) or \(11.3\%\)). Lateral contraction features are observed in the smooth facies, so, anticipating that much of the \(-4\%\) volume decrease upon solidification resulted in lateral shrinkage, we choose the \(-11.3\%\) bound as more realistic and use the corresponding thickness of \(15.5\) km as the thickness of the Orientale melt sheet in all following calculations.

The diameter (\(350\) km) and thickness (\(15.5\) km) of the melt sheet can be used to calculate the volume of the melt sheet. (Note that the thickness of the solidified, cooled melt sheet is not \(15.5\) km, but \(-14\) km – i.e., \(15.5\) km \((-100–11.3\%\); its volume is 84\% of the original melt sheet volume.) Unfortunately, the geometry of the melt sheet is poorly constrained due to the difficulty in predicting the events following the formation of the transient cavity (Fig. 4) leading up to cavity collapse (Head, 2010). The approximation that the melt sheet is shaped like a spherical cap gives a melt sheet volume of \(-8 \times 10^5\) km\(^3\). Inward movement and upward translation during transient cavity collapse could potentially produce a cylindrical melt sheet (Head et al., 1993) with a volume of \(-1.5 \times 10^8\) km\(^3\). The true melt sheet volume is probably somewhere between the extremes of \(-8 \times 10^5\) km\(^3\) and \(-1.5 \times 10^8\) km\(^3\). These bounds are in striking agreement with the \(-0.5\) km\(^3\) estimated from the melt production model of Cintala and Grieve (1998) but are inconsistent with a melt sheet volume of \(-1.9 \times 10^8\) km\(^3\) estimated from the melt production model of Abramov et al. (2012) and our calculations in Section 2.4. Therefore, we take \(-1.5 \times 10^8\) km\(^3\) as the volume of impact melt produced by the Orientale-forming impact, since this volume determined from the melt scaling model of Cintala and Grieve (1998) is most consistent with our estimates of melt sheet volume. We hereafter describe this 15.5 km thick melt sheet with a volume of \(-10^8\) km\(^3\) as a melt sea to emphasize its enormous dimensions (Fig. 1).

In summary, most impact melt (\(-2/3\)) produced by the Orientale-forming impact occurs in a \(-15.5\) km thick (now \(-14\) km thick) impact melt sea \(-350\) km in diameter with a volume of \(-10^6\) km\(^3\). We now model the petrology of this impact melt sea.

3. Cumulate stratigraphy of the Orientale melt sea

3.1. Did the Orientale melt sea differentiate?

Mounting petrologic evidence suggests that impact melt sheets associated with the large terrestrial craters Manicouagan (O’Connell-Cooper and Spray, 2011), Sudbury (Grieve et al., 1991; Thierriault et al., 2002; Zieg and Marsh, 2005), and Morokweng (Hart et al., 2002) have undergone large-scale igneous differentiation.

![Fig. 7. Vertical subsidence in meters as a function of impact melt sheet thickness for the 11.3% and 14% limits derived in this paper and in Wilson and Head (2011).](image-url)
resulting in cumulate stratigraphies similar to those of well-known differentiated plutos such as the Skagaard and the Stillwater. The Sudbury Igneous Complex, the largest of these impact melt sheets, has a thickness of ~2.5–3 km and a volume of ~10^8 km^3 (Thierrault et al., 2002). On the basis of our analysis in Section 2, the solidified Orientale melt sea has a thickness of ~14 km and a volume of ~10^6 km^3. Terrestrial impact melt sheets less than a tenth of the thickness of the Orientale melt sea and a hundredth of its volume have undergone large-scale igneous differentiation. We therefore anticipate that the Orientale melt sea too has undergone large-scale igneous differentiation.

The idea that lunar impact melt sheets have undergone igneous differentiation is by no means new: see, e.g., Basilevsky and Neukum (2010), Cintala and Grieve (1998), Grieve et al. (1991), and Morrison (1998). However, this idea deserves to be revisited now more than ever, as recent and ongoing missions (the Moon Mineralogy Mapper (M3) instrument on India’s Chandrayaan-1 mission and the Lunar Reconnaissance Orbiter (LRO) and Gravity Recovery and Interior Laboratory (GRAIL) missions) collect high-resolution compositional and geophysical data that make it possible to investigate the lunar highlands crust (and massive impact melt sheets) in unprecedented detail.

3.2. Modeling the cumulate stratigraphy of the Orientale melt sea

In the following sections, we develop a model for the cumulate stratigraphy of the Orientale melt sea (indeed, for any impact melt sea formed from melting of the lunar highlands crust and/or upper mantle: see Section 4.2). This model makes predictions about observables: namely, the lithologies present at depth in a solidified melt sea, which may be sampled by central peaks and walls of superposed craters, and the density structure of this solidified melt sea, which may be determined by seismic and/or gravity measurements.

A general model of cumulate stratigraphy can be subdivided into three component models: (1) a mixing model to recover the composition of the differentiating liquid, if unknown; (2) a thermodynamic model to determine the crystallization sequence of this liquid composition; and (3) a fluid mechanical model to convert this crystallization sequence into a stratigraphic sequence. Our model of the cumulate stratigraphy of the Orientale melt sea (1) determines the composition of the differentiating liquid by mixing lunar crust and upper mantle compositions weighted by their mass fractions in a modeled melt cavity, (2) finds the crystallization sequence of this liquid on the Fo–An–Qz pseudoternary phase diagram, and (3) sinks or floats crystals according to crystal–liquid density contrasts in order to convert the crystallization sequence into a stratigraphic sequence.

3.3. Composition of liquid(s) in the Orientale melt sea (bulk composition)

We calculate the bulk composition of the Orientale melt sea by mixing lunar crust and upper mantle compositions weighted by their mass fractions in a modeled melt cavity (Fig. 4a).

The geometry of the melt cavity (Fig. 4a) has been approximated as a truncated sphere (Barr and Citron, 2011; Pierazzo et al., 1997). For simplicity, we treat the melt cavity as a sphere tangent to the lunar surface. The volume of this spherical melt cavity must be the volume of impact melt produced, which was calculated in Section 2.3 to be to be ~1.5 × 10^8 km^3. Therefore, the radius of the melt cavity is ~70 km (i.e., the cube root of 1.5 × 10^8 km^3/(4/3π)), rounded to an even 70 km for the following calculations. The diameter of the melt cavity is approximately twice the average crustal thickness of the lunar farside (Wieczorek et al., 2006; Zuber et al., 1994), implying that the Orientale impact melted the lunar upper mantle. The Orientale impact apparently did not excavate the lunar upper mantle—no mafic mantle material has been detected in Orientale basin deposits (Head, 2010; Head et al., 2010a; Pieters et al., 2009; Yamamoto et al., 2010)—but this is as expected, since in Orientale-size lunar basins with a transient cavity 620 km in diameter the depth of melting is modeled to exceed the depth of excavation by a factor of two (see Fig. 22 of Cintala and Grieve (1998)).

We model the lunar highlands crust and upper mantle as composed of entirely anorthite, enstatite, and forsterite in varying proportions (Shearer et al., 2006). This is a good approximation for the crust—lunar highland rocks generally have modal plagioclase + pyroxene + olivine > 99 (Shearer et al., 2006; Wieczorek et al., 2006)—and a reasonable approximation for the upper mantle from the base of the crust to 200 km, which has been modeled as orthopyroxenite by thermodynamic models matching measured and modeled seismic velocities (Khan et al., 2006; Kuskov, 1997). Ti-rich phases are neglected, but these are probably more important in the nearside crust and upper mantle (Wieczorek et al., 2006). We take the farside crustal thickness to be 70 km (Zuber et al., 1994). This is somewhat thicker than suggested by recent modeling (Wieczorek et al., 2006), but using the number of 70 km affords two conveniences: first, in this case the modeled depth of melting is twice the thickness of the lunar crust, so that exactly half the volume of the melt cavity can be taken to be crustal material; second, approximating the melt cavity geometry as a sphere rather than as a truncated sphere (Barr and Citron, 2011) somewhat decreases the volume of crustal material in the melt sea, which is counteracted by increasing the modeled crustal thickness. The lunar crust is compositionally stratified: the upper crust is anorthosite and the lower crust, exposed in basin ejecta, is probably anorthositic norite (Pieters et al., 2009; Wieczorek et al., 2006). Therefore, we separate the crust into two planar layers of equal 35 km thickness: the thicknesses of these crustal layers are equal, as is approximately the case in the dual-layered crustal thickness model of Wieczorek et al. (2006). The densities and compositions of these layers are based on the dual-layered crustal thickness model of Wieczorek et al. (2006) with the additional constraint that enstatite and forsterite are present in the proportion 3:1 in both these layers, justified by the modal mineralogy of anorthosite and anorthositic norite samples (some lunar sample modes are tabulated in Wieczorek et al. (2006)). The upper mantle is taken to have a density of 3.3 g/cm^3 and a composition of 75 wt.% enstatite and 25 wt.% forsterite (Khan et al., 2006). All bulk composition model parameters are presented in Table 1.

These parameters (Table 1) capture in the An–Fo–Qz system the salient features of the lunar crust and upper mantle—an anorthosite upper crust, an anorthositic norite lower crust, and a pyroxene upper mantle. Minor changes in these parameters do not change the crystallization sequences in Section 3.5.

The mass fraction of each layer melted by a 140 km diameter spherical melt cavity tangent to the lunar surface is calculated from the volume of the spherical melt cavity intersecting the layer and the density of the layer (Table 2). The bulk composition of the melt sea is then given by the compositions of each layer weighted by their mass fractions in the melt (Table 2). The resulting melt sea is fairly mafic, with an olivine noritic bulk composition (Stoffler et al., 1980). This modeled mafic bulk composition is not necessarily inconsistent with

<table>
<thead>
<tr>
<th>Layer</th>
<th>Thickness (km)</th>
<th>Density (g/cm^3)</th>
<th>An (wt.%)</th>
<th>En (wt.%)</th>
<th>Fo (wt.%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper crust</td>
<td>35</td>
<td>2.8</td>
<td>86</td>
<td>10.5</td>
<td>3.5</td>
</tr>
<tr>
<td>Lower crust</td>
<td>35</td>
<td>3.0</td>
<td>60</td>
<td>30</td>
<td>10</td>
</tr>
<tr>
<td>Upper mantle&gt;70</td>
<td>3.3</td>
<td>0</td>
<td>75</td>
<td>25</td>
<td></td>
</tr>
</tbody>
</table>
remote sensing observations suggesting that surface or near-surface portions of the melt sea are anorthositic or noritic (see Section 4.1), since melt sea differentiates need not have the bulk composition of the melt sea (see the discussion of South Pole-Aitken basin in Section 4.2.2).

### 3.4. Initial composition of liquid(s) in the Orientale melt sea (melt mixing)

We now investigate how the bulk composition of the Orientale melt sea is related to the initial composition of differentiating liquid(s) in the sea.

Vigorous convection, driven by the thermal gradient between hot melt and a cold free surface, is anticipated in a large melt sea (Morrison, 1998). Convection may mix and homogenize the melt sea so that the composition of any small volume of melt is identical to the bulk composition of the melt. In this case, the initial composition of the differentiating liquid in the Orientale melt sea is simply the bulk composition of the Orientale melt sea given in Table 2 above.

However, even a well-mixed melt sea may not be homogenous: Zieg and Marsh (2005) suggest that the relatively homogenous granophyre and norite layers of the Sudbury Igneous Complex, a terrestrial impact melt sheet, formed due to physical liquid immiscibility. In their scenario, mafic and felsic liquids (melts of different target lithologies) were prevented from rapidly mixing by shear due to their order-of-magnitude different viscosities. Before these liquids could mix by diffusion, they separated according to their viscosity and density, a process taking only weeks to months (Zieg and Marsh, 2005) after impact, much longer than the processes that drive melt ejection, transient cavity excavation and subsequent crater modification (Melosh, 1989; Hawke and Head, 1977). On this basis, we take the melt to have been homogeneous during melt cavity collapse: in this case, even though ejection decreases the volume of the melt sea by one-third, it does not affect its bulk composition.

### 3.5. Crystallization sequences in the Orientale melt sea

We now model the crystallization sequences of homogenous and density-stratified melt seas. Both equilibrium crystallization (crystals always in equilibrium with liquid) and perfect fractional crystallization (crystals only instantaneously in equilibrium with liquid) are modeled in order to bound melt sea lithologies, although vigorous convection is likely to have kept crystals in equilibrium with liquid in the Orientale melt sea.

Crystallization sequences are determined using the An–Fo–Qz pseudoternary phase diagram (Fig. 8a). The 1 atm phase diagram is used since pressures are <1 kbar even at the bottom of the melt sea, 15.5 km deep in the crust (Mizutani and Osako, 1974). The melt has an initial temperature >2000 °C (Melosh, 1996), well above the liquidus of the homogenous melt sea at ~1550 °C. (The liquidus of the mafic liquid in the density-stratified melt sea is similar, ~1650 °C.)

#### 3.5.1. Equilibrium crystallization of a homogenous melt sea

Olivine-saturated liquid crystallizes olivine until the liquid becomes sufficiently silicic to react with previously-crystallized olivine to form orthopyroxene. The liquid evolves along the olivine–orthopyroxene boundary curve to the peritectic, where plagioclase comes on the liquidus; the liquid is consumed before all olivine has been resorbed. The final crystallization sequence is olivine → orthopyroxene + plagioclase + orthopyroxene + olivine (Fig. 8b).

#### 3.5.2. Fractional crystallization of a homogenous melt sea

Olivine-saturated liquid crystallizes olivine until the liquid becomes sufficiently silicic to crystallize orthopyroxene. No olivine solids are in equilibrium with the liquid, which leaves the olivine–orthopyroxene boundary curve, crystallizing orthopyroxene until the liquid becomes plagioclase-saturated. The liquid evolves along the plagioclase–orthopyroxene boundary curve to the eutectic, where plagioclase, orthopyroxene, and quartz co-crystallize until the liquid is consumed. The final crystallization sequence is olivine → orthopyroxene + plagioclase + orthopyroxene + plagioclase + orthopyroxene + quartz (Fig. 8c).

#### 3.5.3. Equilibrium crystallization of a density-stratified melt sea

The pure plagioclase-composition liquid does not undergo differentiation. The crystallization sequence of the mafic liquid (a Fo + En mixture) follows from the well-known Fo–Qz binary phase diagram: olivine-saturated liquid crystallizes olivine until the liquid evolves to the peritectic, where olivine reacts with liquid to form orthopyroxene; the liquid is consumed before all olivine has been consumed.

### Table 2
Bulk composition of the Orientale melt sea.

<table>
<thead>
<tr>
<th>Proportion of melt sea</th>
<th>Layer</th>
<th>(vol.%)</th>
<th>(wt.%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper crust</td>
<td>15.6</td>
<td>14.0</td>
<td></td>
</tr>
<tr>
<td>Lower crust</td>
<td>34.4</td>
<td>33.1</td>
<td></td>
</tr>
<tr>
<td>Upper mantle</td>
<td>50.0</td>
<td>52.9</td>
<td></td>
</tr>
<tr>
<td>An (wt.%)</td>
<td>En (wt.%)</td>
<td>Fo (wt.%)</td>
<td></td>
</tr>
<tr>
<td>Bulk composition of melt sea</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>31.9</td>
<td>51.1</td>
<td>17.0</td>
<td></td>
</tr>
</tbody>
</table>

Table 3
Density-stratified melt sea parameters. Layer thicknesses are for a cylindrical melt sea 15.5 km thick.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Density (g/cm³)</th>
<th>Proportion of melt sea (wt.%)</th>
<th>(vol.%)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anorthositic liquid</td>
<td>2.61</td>
<td>31.9</td>
<td>34.4</td>
<td>5.3</td>
</tr>
<tr>
<td>Mafic liquid</td>
<td>2.92</td>
<td>68.1</td>
<td>65.6</td>
<td>10.2</td>
</tr>
</tbody>
</table>
3.5.4. Fractional crystallization of a density-stratified melt sea

The pure plagioclase-composition liquid does not undergo differentiation. The crystallization sequence of the mafic liquid (a Fo + En mixture) follows from the well-known (see, e.g., Winter (2010)) Fo–Qz binary phase diagram: olivine-saturated liquid crystallizes olivine until the liquid evolves to the peritectic. All olivine has been removed from the liquid, so the liquid composition leaves the peritectic, crystallizing orthopyroxene and finally orthopyroxene and quartz at the eutectic. The final crystallization sequence of the mafic liquid is olivine $\rightarrow$ orthopyroxene $\rightarrow$ orthopyroxene + quartz.

The crystallization sequences above are probably a good first approximation to the true crystallization sequence of the Orientale melt sea. However, certain phases are not modeled, such as clinopyroxene and Ti-bearing phases. Also, the composition of solids is assumed to be constant, whereas in reality plagioclase, olivine, and orthopyroxene are all solid solutions. The Mg or An number of the first olivine/orthopyroxene or plagioclase solids is almost certainly greater than the Mg or An number of the last olivine/orthopyroxene or plagioclase solids. Since the appearance of clinopyroxene or changes in Mg or An number may provide important constraints on melt sheet differentiation, we plan to use more sophisticated thermodynamic models to investigate the crystallization sequences of lunar impact melt sheets in greater detail.

3.6. Crystal–liquid density contrasts in the Orientale melt sea

We compare crystal and liquid densities to determine whether crystallizing solids will sink or float in order to convert the crystallization sequences in Section 3.5 to stratigraphic sequences.

An–Fo–Qz liquids are substantially less dense (always >0.1 g/cm³, calculated from Nelson and Carmichael (1979)) than anorthite crystals ($\rho = 2.75$ g/cm³). Even in an iron-rich liquid with a constant Mg number of 40 (Fig. 9a), more ferroan than even the interstitial mafic minerals in ferroan anorthosites (Shearer et al., 2006), anorthite never floats in a plagioclase-saturated liquid (Fig. 9b). Our conclusion is that anorthite crystals do not float in (Ti-poor) impact melts of the crust and upper mantle. The critical implications of this idea for the cumulate stratigraphies and density structures in impact melts of the crust will be considered in Section 4.2.1. Since plagioclase crystals are substantially less dense than orthopyroxene and olivine crystals ($\rho \sim 3.2$ g/cm³), we conclude that all crystallizing solids sink in their coexisting liquids. Therefore, the crystallization sequences in Section 3.5 above are effectively stratigraphic sequences: for example, co-crystallizing plagioclase and orthopyroxene both sink to form norite.

3.7. Normative calculations and stratigraphic columns for the Orientale melt sea

Finally, we calculate the weight proportions and thicknesses of each co-crystallizing assemblage to create stratigraphic columns for solidified melt seas. (Although crystal settling is likely to be
kinetically incompatible with perfect equilibrium crystallization, we nonetheless convert equilibrium crystallization sequences to stratigraphic sequences in order to bound melt sea stratigraphies. Real stratigraphic sequences may approach sequences based on equilibrium crystallization. The following sections briefly describe our normative calculations, which are based on the An–Fo–Qz and Fo–Qz phase diagrams.

3.7.1. Equilibrium crystallization of a homogenous melt sea

The liquid crystallizes to give monomineralic olivine (dunite) which then reacts with liquid to form an overlying layer of pyroxenite and finally norite as plagioclase comes on the liquidus at the peritectic. The weight proportion of the dunite layer is given simply by the initial wt.% Fo of the liquid; the norite layer has an An/En ratio of ~3/2 as measured from the phase diagram and, accordingly, its weight proportion is given by the sum of the initial wt.% An and 2/3 that wt.% of En; the wt.% of the pyroxenite layer is equal to the remaining wt.% En (Table 4).

3.7.2. Fractional crystallization of a homogenous melt sea

The weight proportions of the four crystallizing lithologies—monomineralic olivine (dunite), monomineralic orthopyroxene (pyroxenite), orthopyroxene + anorthite (norite), and orthopyroxene + anorthite + quartz (quartz diorite)—are constrained by four equations, the first three relating the compositions of each lithology (measured from the An–Fo–Qz ternary) to the bulk composition and the fourth relating the relative proportions of dunite and pyroxenite as determined from olivine and orthopyroxene crystallization path length on the An–Fo–Qz ternary (Fig. 8c). The resulting system of linear equations can be written as the following matrix equation:

\[
\begin{pmatrix}
1 & 0.7 & 0.32 & 0.16 \\
0 & 0 & 0.55 & 0.51 \\
0 & 0.3 & 0.13 & 0.33 \\
-1 & 1.5 & 0 & 0
\end{pmatrix}
\begin{pmatrix}
X_{Du} \\
X_{Px} \\
X_{Nr} \\
X_{Qz}
\end{pmatrix}
= 
\begin{pmatrix}
0.53 \\
0.32 \\
0.15 \\
0
\end{pmatrix}
\]

We solve this matrix equation for the weight proportions of these lithologies (Table 4).

3.7.3. Equilibrium crystallization of a density-stratified melt sea

The overlying anorthositic liquid crystallizes to anorthosite. The mafic liquid crystallizes to form monomineralic olivine (dunite) and monomineralic orthopyroxene (pyroxenite) layers. The weight proportions (Table 4) of all three layers are given simply by the initial wt.% An, Fo, and En of the liquid.

3.7.4. Fractional crystallization of a density-stratified melt sea

The weight proportions of the three crystallizing lithologies—monomineralic olivine (dunite), monomineralic orthopyroxene (pyroxenite), and orthopyroxene + quartz (quartz pyroxenite)—are constrained by three equations, the first two relating the composition of each lithology (measured from the Fo–Qz binary) to the bulk composition and the third relating the relative proportions of dunite and pyroxenite as determined from olivine and orthopyroxene crystallization path length on the Fo–Qz binary. The resulting system of linear equations can be written as the following matrix equation:

\[
\begin{pmatrix}
1 & 0.7 & 0.55 \\
0 & 0.3 & 0.45 \\
-1 & 1.5 & 0
\end{pmatrix}
\begin{pmatrix}
X_{Du} \\
X_{Px} \\
X_{Qz}
\end{pmatrix}
= 
\begin{pmatrix}
0.78 \\
0.22 \\
0
\end{pmatrix}
\]

We solve this matrix equation for the weight proportions of these lithologies (Table 4).

![Figure 10](image-url) Model Orientale cumulative stratigraphies and density profiles (to scale) produced by equilibrium and fractional crystallization of homogenous and density-stratified melt seas.

<table>
<thead>
<tr>
<th>Layer Composition</th>
<th>Density (g/cm³)</th>
<th>Proportion of melt sea (wt.%) (vol.%)</th>
<th>Thickness (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Equilibrium crystallization, homogenous melt sea</td>
<td>Norite 61.6 wt.% An, 38.4 wt.% En</td>
<td>2.95</td>
<td>51.8</td>
</tr>
<tr>
<td></td>
<td>Pyroxenite 100 wt.% En</td>
<td>3.20</td>
<td>31.2</td>
</tr>
<tr>
<td></td>
<td>Dunite 100 wt.% Fo</td>
<td>3.25</td>
<td>17.0</td>
</tr>
<tr>
<td>Fractional crystallization, homogenous melt sea</td>
<td>Quartz diorite 51.0 wt.% An, 26.1 wt.% Qz</td>
<td>2.85</td>
<td>13.7</td>
</tr>
<tr>
<td></td>
<td>Anorthositic norite 55.0 wt.% An, 45.0 wt.% En</td>
<td>2.95</td>
<td>45.3</td>
</tr>
<tr>
<td></td>
<td>Pyroxenite 100 wt.% En</td>
<td>3.20</td>
<td>16.4</td>
</tr>
<tr>
<td></td>
<td>Dunite 100 wt.% Fo</td>
<td>3.25</td>
<td>24.6</td>
</tr>
<tr>
<td>Equilibrium crystallization, density-stratified melt sea</td>
<td>Anorthosite 100 wt.% An</td>
<td>2.75</td>
<td>31.9</td>
</tr>
<tr>
<td></td>
<td>Pyroxenite 100 wt.% En</td>
<td>3.20</td>
<td>51.1</td>
</tr>
<tr>
<td></td>
<td>Dunite 100 wt.% Fo</td>
<td>3.25</td>
<td>17.0</td>
</tr>
<tr>
<td>Fractional crystallization, density-stratified melt sea</td>
<td>Anorthosite 100 wt.% An, 78.6 wt.% En</td>
<td>2.75</td>
<td>31.9</td>
</tr>
<tr>
<td></td>
<td>Quartz pyroxenite 21.4 wt.% Qz</td>
<td>3.10</td>
<td>27.8</td>
</tr>
<tr>
<td></td>
<td>Pyroxenite 100 wt.% En</td>
<td>3.20</td>
<td>9.4</td>
</tr>
<tr>
<td></td>
<td>Dunite 100 wt.% Fo</td>
<td>3.25</td>
<td>30.9</td>
</tr>
</tbody>
</table>
Finally, the weight proportions are converted to volume proportions based on layer density (a weighted average of component mineral density); these volume proportions are directly proportional to the thicknesses in a cylindrical melt sea (Table 4). We choose to model melt seas as cylindrical since this geometry has been predicted as the consequence of inward movement and upward translation during transient cavity collapse (Head, 2010, 2012; Head et al., 1993). The effect of a melt sea with a curved base, which has more of its volume near the surface, is to increase the thickness of deep layers and decrease the thickness of shallow layers.) The cumulative stratigraphies and density structures of these modeled melt seas are shown in Fig. 10.

4. Discussion

We now assess our modeled cumulative stratigraphies (Fig. 10) and consider more generally the implications of impact melt differentiation for the lunar highlands crust.

4.1. Constraints on Orientale melt sea cumulative stratigraphy: implications for processes once operating in the Orientale melt sea

The modeled cumulative stratigraphies (Fig. 10) produced by melt homogenization/density stratification and equilibrium crystallization/fractional crystallization have observable differences. Constraints on the lithologies present at depth in the solidified Orientale melt sea make it possible to rule out some of these modeled stratigraphies and determine what processes operated in the mixing and crystallization of the Orientale melt sea.

There are currently two important constraints on the cumulative stratigraphy of the Orientale melt sea: (1) The top ~1 km of the solidified impact melt sea is anorthosite. M² spectral observations of the Mauder Formation (Head, 2010; Head et al., 2010a; Pieters et al., 2009) indicate that the upper surface of the smooth facies is anorthosite. Moreover, small (~10 km diameter) craters superposing the impact melt sheet, which depth/diameter ratios predict excavate material from a depth of up to ~1 km, uncover more anorthosite. (2) Norite occurs at ~4 km depth. The central peak of Mauder, a 55 km diameter crater superposed on the Orientale melt sheet, is predicted by depth/diameter ratios to have sampled the melt sea at a depth of ~4 km (Whitten et al., 2011). This peak has a noritic composition (Whitten et al., 2011). These two constraints can themselves be interpreted as evidence for the differentiation of the Orientale melt sea: near-surface lithologies are not the same as those present at depth.

Which modeled cumulative stratigraphies (Fig. 10) are consistent with these constraints? Fractional crystallization of a homogenous melt sea produces a ~2 km thick layer of quartz diorite at the top of the melt sea, clearly inconsistent with constraint (1). Density stratification produces a ~5 km thick layer of pure anorthosite at the top of the melt sea, inconsistent with constraint (2). Equilibrium crystallization of a homogenous melt sea produces a ~7.5 km thick layer of norite at the top of the melt sea, consistent with constraint (2). This is not necessarily inconsistent with constraint (1): the anorthositic roof of the Orientale melt sheet could be an early-formed quench crust; a fallback breccia, as in Sudbury (Thierrault et al., 2002); or the result of modal separation of plagioclase from pyroxene in late-crystallizing norite.

Therefore, we conclude that the Orientale melt sea was homogenized by melt mixing and underwent equilibrium crystallization. Indeed, both homogenization and equilibrium crystallization suggest vigorous convection, consistent with a cooling lava lake (Worster et al., 1990) and particularly with a cooling impact melt sea (Morrison, 1998).

It is possible that high-resolution gravity measurements from the ongoing GRAIL mission could provide additional constraints on Orientale melt sea cumulative stratigraphy. Certainly, gravity measurements from the GRAIL mission can provide constraints on melt sea thickness. We note in passing that, in light of our evidence for a thick sea of impact melt at the center of the Orientale basin, crustal thickness models for the Orientale basin need to be reassessed (Fig. 11). We predict that the “crust” occurring at the center of the Orientale basin is actually a solidified melt sea with a density of ~3.05 g/cm³, not anorthosite or brecciated anorthosite with a density of ~2.75 g/cm³, as has been traditionally assumed (Andrews-Hanna and Stewart, 2011; Hikida and Wieczorek, 2007). The ~10 km thick anorthositic crust modeled to occur at the center of the Orientale basin (Andrews-Hanna and Stewart, 2011; Hikida and Wieczorek, 2007) becomes a ~20 km thick impact melt sheet (Fig. 11) using this revised density, roughly consistent with (~although ~>1.3 times greater than) our melt sheet thickness estimates in Section 2.

4.1.1. Is crystal settling an operative differentiation mechanism in impact melt seas?

In this paper, we interpret the differentiation of the Orientale impact melt sea (and other lunar impact melt seas; see Section 4.2) in terms of crystal settling: we transform crystallization sequences into stratigraphic sequences under the assumption that gravity efficiently segregates crystals from melt to form igneous strata. We are by no means the first authors to make this assumption: Morrison (1998) and Nakamura et al. (2009) used crystal settling to interpret the differentiation of the South Pole-Aitken Basin impact melt sheet. However, field studies and theoretical analyses of terrestrial impact melt sheets and magma chambers (Marsh, 1989; Marsh, 1996; Warren et al., 1996; Zieg and Marsh, 2005) may indicate that crystal settling does not operate in the differentiation of impact melt. Field studies of the Sudbury Igneous Complex, a terrestrial impact melt sheet, show that massive granophyre and norite layers in the Sudbury Igneous Complex are homogenous; crystal settling did not take place in their parent magmas (Zieg and Marsh, 2005). This is perhaps unsurprising: crystal settling operates on isolated phenocrysts in a magma body, not the dense mat of crystals along the cooling margin of a magma body (Marsh, 1989; Marsh, 1996); impact melts, however, are superheated (>2000 °C) (Melosh, 1996), and contain no phenocrysts to settle. Marsh (1996) writes, “...initially crystal-free magmas [such as impact melts] cannot undergo strong chemical differentiation by fractional crystallization of singly settling crystals.”

What are the implications for solidified melt sea stratigraphy if crystal settling indeed does not operate in the differentiation of impact melt seas? Impact melt bodies are then either (1) homogenous or (2) heterogeneous but differentiated by another mechanism, such as viscous emulsion differentiation (Zieg and Marsh, 2005). However, the constraints on Orientale melt sea cumulative stratigraphy described in Section 4.1 are inconsistent with large-scale homogeneity and not entirely consistent with viscous emulsion differentiation (Fig. 10, lower left). Perhaps our model of viscous emulsion differentiation is colored by our assumption that the crater excavation flow does not change the bulk composition of the impact melt sea: if the excavation flow removed only impact melt of upper crustal composition, an uppermost anorthosite layer (Fig. 10, lower left) would be thinned, bringing noritic lithologies towards the surface. In this case, the noritic central peak of Mauder represents recrystallized lower crustal material.

If viscous emulsion differentiation is indeed the operative differentiation mechanism in impact melt seas, solidified impact melt seas recapitulate the structure of the lunar crust and upper mantle, just as the Sudbury impact melt sheet recapitulates the preexisting
crustal structure of the Canadian shield (Zieg and Marsh, 2005). Moreover, since impact melt seas have a very different aspect ratio from the initial melt cavity (Fig. 4), solidified impact melt bodies may bring lower crust and upper mantle materials to the near surface where they can be excavated by complex craters of moderate size (such as Maunder in Orientale). Further investigation of lunar impact melt sea stratigraphy will clarify the differentiation of lunar impact melts and the roles of crystal settling and viscous emulsion differentiation.

4.2. Cumulate stratigraphy of other lunar impact melt seas

There are at least 30 lunar basins (craters >300 km diameter) besides Orientale, as determined from the database of Head et al. (2010b). Impact melt scaling relations (Cintala and Grieve, 1998) predict that the impacts that formed these basins each produced >10^5 km^3 of melt. Very probably, these basins, like Orientale, contain solidified impact melt seas. What is the cumulate stratigraphy of these melt seas and how might it differ from the cumulate stratigraphy of Orientale? Since target crust/upper mantle structure is not thought to change substantially across the farside highlands (Wieczorek et al., 2006), the important variable controlling melt sea bulk composition is depth of melting, which is a function of basin size. The compositional endmembers are (1) relatively small basin-forming impacts that melt only the lunar crust and (2) giant basin-forming impacts that melt mainly the lunar mantle (Fig. 12). The compositional watershed is the plagioclase-olivine cotectic (Fig. 12): small basin-forming impacts form melt seas that become plagioclase-saturated early, whereas large basin-forming impacts form melt seas that become plagioclase-saturated late in crystallization. Having considered the cumulate stratigraphy of a melt sea in a moderately-sized basin, Orientale, we now proceed to model the cumulate stratigraphy of melt seas occurring in a small lunar basin and the very large South-Pole Aitken basin.

4.2.1. Cumulate stratigraphy of a melt sea in a small lunar basin

We consider a hypothetical basin-forming impact that melts the entire thickness of the lunar crust and none of the upper mantle, corresponding to a spherical melt cavity 70 km in diameter, which we take to be the thickness of the farside crust (Wieczorek et al., 2006; Zuber et al., 1994). A spherical melt cavity 70 km in diameter contains ~1.7 x 10^5 km^3 of melt, which, according to the scaling relations of Cintala and Grieve (1998), corresponds to a basin transient cavity ~250 km in diameter and a final basin diameter of ~415 km (Croft, 1985), almost exactly the diameter of the basin Korolev (Baker et al., 2011a). We reuse our modeled bulk composition parameters (Table 1) to calculate the bulk composition of this crustal melt sea: 72.5 wt.% An, 20.5 wt.% En, and 7.0 wt.% Fo. We model equilibrium crystallization of a homogenized crustal melt sea since these processes best reproduced the observed cumulate stratigraphy of the Orientale melt sea (Section 4.1).

This crustal melt is initially plagioclase-saturated (Fig. 13a). Plagioclase crystallizes until the liquid becomes olivine-saturated; the liquid then evolves along the plagioclase Olivine cotectic to the
peritectic, where orthopyroxene and plagioclase co-crystallize; the liquid is consumed before all olivine has been resorbed. The crystallization sequence of this liquid (Fig. 13b) is plagioclase → plagioclase + olivine → plagioclase + orthopyroxene – olivine. Plagioclase sinks in impact melts of the crust (Fig. 9), so this crystallization sequence is effectively the cumulate stratigraphy. We calculate the weight proportions of these co-crystallizing assemblages using methods similar to those in presented in Section 3.7. The cumulate stratigraphy of a crustal melt sea occurring in a Korolev-sized basin is compared with the likely cumulate stratigraphy of the solidified Orientale melt sea in Fig. 14.

The lithologies exposed at or near the surface or in the central peaks of complex craters are modeled to be norite in both small and large melt seas. The critical observable difference between these two melt seas is their density structure. The crustal melt sea has an inverted density profile—the result of early-crystallizing plagioclase solids sinking to form an anorthosite layer beneath denser troctolite and norite layers—whereas the Orientale and other mantle melt seas have a regular density profile with a dense dunite layer underlying less dense pyroxenite and norite. Future seismic measurements on solidified melt sea surfaces could potentially resolve these different density structures.

4.2.2. South Pole-Aitken basin melt sea cumulate stratigraphy

The enormous 2500 km diameter South Pole-Aitken (SPA) basin (Wilhelms et al., 1979) almost certainly melts huge volumes of ultramafic mantle material (Morrison, 1998). However, the SPA floor, possibly the surface of a melt sea, is noritic, not ultramafic, and craters superposing this floor excavate pyroxene, not mantle olivine (Borst et al., 2012; Moriarty et al., 2011, 2012; Nakamura et al., 2009). Morrison (1998) suggested that differentiation of the SPA melt sheet could reconcile these apparently conflicting observations: a basal ultramafic layer could sequester mantle olivine and an upper noritic layer could concentrate crustal plagioclase. We estimate the thickness and model the cumulate stratigraphy of the South Pole-Aitken basin melt sea to investigate this idea in greater detail.

How thick is the SPA melt sea? We take the SPA transient cavity diameter to be 2099 km (Wicenzek and Phillips, 1999). Using the impact melt volume scaling relationship derived by Abramov et al. (2012) and the model parameters described in Kring et al. (2012), we calculate that the South Pole-Aitken basin-forming impact produced \( \sim 6.8 \times 10^7 \) km\(^3\) of impact melt. About 25 vol.% of this melt was excavated from the basin transient cavity, leaving \( 5.1 \times 10^7 \) km\(^3\) of impact melt inside the inner basin ring to form a melt sea. The thickness of this melt sea is constrained by its areal extent. Morrison (1998) calculated the areal extent of the SPA melt sea by taking the inner ring of SPA to be a circle \( \sim 1800 \) km in diameter (Wilhelms et al., 1979). We calculate the areal extent of the SPA melt sea using the recent measurements by Garrick-Bethell and Zuber (2009) suggesting that the inner ring of SPA is an ellipse with a semimajor axis of 970 km and a semiminor axis of 720 km. A prismatic melt sea with this areal extent containing \( \sim 5.1 \times 10^7 \) km\(^3\) of impact melt is \( \sim 25 \) km thick. However, the Orientale melt sea probably occurs in a central depression with a diameter \( \sim 70\% \) of that of the Outer Rook Ring (350 km vs. 480 km). If this ratio holds for SPA, so that the semimajor and semiminor axes of the surface of the SPA melt sea are \( \sim 680 \) km (70% of 970 km) and \( \sim 500 \) km (70% of 720 km), then the SPA melt sea may be \( \sim 50 \) km thick.

What is the cumulate stratigraphy of a hypothetical solidified SPA melt sea? A spherical melt cavity containing the estimated \( \sim 6.8 \times 10^7 \) km\(^3\) SPA melt volume has a radius of \( \sim 250 \) km, corresponding to a depth of melting of \( \sim 500 \) km. The bulk composition of the SPA melt sea, reusing the bulk composition model parameters in Table 1, is nearly that of the upper mantle, with 4 wt.% An, 72 wt.% En, and 24 wt.% Fo. The crystallization sequence of this melt and the modal calculations are the same as for the Orientale melt lake described in Section 3. The solidified South Pole-Aitken basin melt sea, if \( \sim 25 \) km thick, is thus predicted to consist of a thin \( \sim 2.6 \) km layer of norite overlaying a massive \( \sim 16.5 \) km layer of pyroxenite and a basal 5.9 km thick dunite cumulate (Fig. 14).

What are the implications of this cumulate stratigraphy? First, the noritic floor of SPA could very well be the roof of a massive differentiated impact melt sea. Second, observations of crater central peaks that determine the depth at which norite transitions to pyroxenite can be used to constrain the thickness of the SPA melt sea. Third, Nakamura et al. (2009) argue that the SPA impact melt sea has not undergone differentiation on the basis that all central peaks observed in their study excavate pyroxenite. We suggest instead that the SPA impact melt sea has undergone differentiation and that all central peaks observed in their study (which excavate material from depths of \( \sim 6-10 \) km) excavate a 16.5 km thick homogenous pyroxenite cumulate (Fig. 14).

4.3. Impact melt differentiates in the lunar sample suite?

Huge volumes of impact melt are produced by basin-forming impacts. Much of this impact melt may have differentiated: the Orientale-forming impact alone may have given rise to \( \sim 10^9 \) km\(^3\)
of impact melt differentiates. For comparison, the total volume of lunar mare basalts is estimated to be $\sim 10^7 \text{ km}^3$ (Head and Wilson, 1992). Could impact melt differentiates excavated from massive impact melt seas be represented in the lunar sample suite?

4.3.1. Can impact melt differentiates pass as pristine?

Pristine lunar rocks (Warren, 1993), which have coarse-grained textures and low siderophile element concentrations, are thought to represent endogenous lunar igneous rocks largely unaltered by impact processes. We suggest that impact melt differentiates formed in massive impact melt seas also have coarse-grained textures and low siderophile element concentrations and so may be able to pass for pristine lunar rocks (Basilevsky and Neukum, 2010).

Massive impact melt seas cool slowly. The diameter of the Orientale melt sea ($\sim 175 \text{ km}$) exceeds its probable depth ($\sim 15.5 \text{ km}$) by an order of magnitude, so we approximate a (non-convecting) melt sea as a cooling slab similar to an intruded sill. The solidification time $t$ of a sill cooling by conduction can be approximated by the formula

$$t = \frac{0.694 (L/\alpha)^2}{K}$$  \hspace{1cm} (3)

where $L$ is the half-thickness of the sill and $K$ is its thermal diffusivity, $\sim 10^{-2} \text{ cm}^2/\text{s}$ for rock (Jaeger, 1968). From (3), we estimate that the Orientale melt sea took on the order of $10^5$ years to solidify, long enough for impact melt differentiates formed in that sea to develop coarse-grained textures (Zieg and Marsh, 2002). These calculations are supported by the observation that impact melt differentiates formed at depth in the $\sim 3 \text{ km}$ thick terrestrial Sudbury impact melt sheet are also coarse-grained (Zieg and Marsh, 2002) and were long mistaken for endogenous plutonic rocks (Thierrault et al., 2002).

Also, although impact melt seas may have a relatively high bulk concentration of meteoritic siderophile elements, these siderophile elements may fractionate into basalt deposits of metal or sulfides unmixed from metal- or sulfur-saturated impact melt, so that silicates from these impact melt seas may have nearly lunar, not meteoritic, siderophile element concentrations.

4.3.2. Excavating impact melt differentiates

Most impact melt differentiates form at depth. Certain impact melt differentiates, such as dunites and troctolites, may form at depths below 10 km. For these lithologies to be exposed at the surface and introduced into the lunar sample suite, they must be excavated by the impact cratering process. Very deep lithologies can only be excavated by very large craters; we can see that deep differentiates cannot possibly have been excavated at Orientale (no craters $>65 \text{ km}$ superpose the Orientale melt sheet, so dunite possibly occurring below 12 km in the Orientale melt sea has never been excavated and cannot occur in the lunar sample suite). At other basin melt sheet locations, shallow impact melt differentiates, probably noritic in composition, are the most likely to be excavated by impacts and introduced into the lunar sample suite. Old basins (Fig. 15) with many superposed large craters and basins, such as the South Pole-Aitken basin (Head, 2010; Head et al., 1993, 2010a; Pieters et al., 1997) are the most likely source regions for deep impact melt differentiates, and certain norites or pyroxenites could come from SPA.

Sections 4.3.1 and 4.3.2 suggest that impact melt differentiates may exist in the lunar sample suite and may be able to pass for highland plutonic rocks. We now consider three highland plutonic rock suites—Mg-suite lithologies, young ferroan anorthosites, and Mg-spinel lithologies—to determine whether these puzzling lithologies could potentially be impact melt differentiates.

4.3.3. Mg-suite lithologies

Mg-suite rocks—dunites, norites, and troctolites characterized by high Mg numbers (70–95) and high rare earth element (REE) content (up to 400 times chondritic) (Hess, 1994; Shearer et al., 2006)—are traditionally thought to have formed from intrusions of mafic magmas into ferroan anorthosite crust, since dunites, norites, and troctolites can be formed from fractional crystallization of a Fo-rich liquid assimilating plagioclase (see Fig. 8a for reference). The Mg-suite chemical signature is harder to explain: simple mixing of a high Mg number liquid with a low Mg number, REE-enriched (KREEP) liquid produces a liquid with insufficiently high Mg number to be the Mg-suite parent magma; Hess (1994) argued that the Mg-suite parent magma is not a mixture of melts, but a melt of a mixture: a mixture of KREEP, olivine-rich dunite, and anorthosite produces a partial melt with high Mg number and REE content. The process giving rise to this mixture is unclear; processes such as cumulate overturn have been invoked (Shearer et al., 2006).

Could differentiation of impact melt seas produce Mg-suite lithologies? Certainly, dunites, troctolites, and norites can form in impact melt seas: the Orientale melt sea contains norite and probably dunite, and smaller impacts that melt mainly the lunar crust (Section 4.2) will produce troctolite. But these lithologies are mixtures of melts and for the reasons described above cannot have the full Mg-suite chemical signature. Also, thermobarometry of Mg-suite rocks suggests that these rocks formed 40–50 km deep in the crust (McCallum and Schwartz, 2001). It is possible that remotely-sensed Mg-suite lithologies are really unrelated Mg-suite-like dunites, troctolites, and norites lacking the Mg-suite chemical signature. These Mg-suite-like lithologies could be impact melt differentiates; however, until Mg-suite-like lithologies are identified in

**Fig. 14.** Model cumulate stratigraphies and density profiles for melt seas in basins of three different sizes: a “crustal” melt sea occurring in a small basin approximately the size of Korolev, the Orientale melt sea, and the SPA melt sea. Layers in each melt sea are to scale, although melt seas are not to the same scale.
the lunar sample suite, the existence of these counterfeits must be regarded as speculation.

4.3.4. Young ferroan anorthosites

Certain ferroan anorthosites with ages <4.4 Ga (Borg et al., 2011) are much too young to have crystallized in the lunar magma ocean. These ferroan anorthosites are thought to be cumulates formed during early plutonism (Shearer et al., 2006). Since impact melting generally resets geochronometers, young ferroan anorthosites could have formed in massive impact melt sheets (formation in thin, flash-frozen impact melt sheets is probably incompatible with the large grain sizes of these anorthosites). One difficulty with this idea is that anorthite is negatively buoyant in impact melts of the crust and mantle (Fig. 9), so anorthosite generally lines the bottoms of melt seas and may be difficult to excavate in quantity. If the anorthositic roof of the Orientale melt sheet indeed formed due to igneous differentiation, young anorthosites from impact melt sheets may be more common, and easier to excavate, than has been imagined.

4.3.5. Mg-spinel lithologies

Recently discovered Mg-spinel lithologies (Pieters et al., 2011), rocks with ~30 wt% spinel containing <5 wt% Fe, have been explained as the product of melt-wallrock reactions of Mg-suite magmas with anorthosite crust (Prissel et al., 2012). Potentially, spinel could form from reaction of anorthite liquids formed by impact melting with mafic minerals or melts. In density-stratified melt seas, anorthosite liquids overlie mafic liquids; large volumes of spinel could potentially be produced at this interface. One difficulty with this idea is that impact melts of the ferroan crust and mantle are relatively ferroan and could produce ferroan spinels.

4.4. Volumetric importance of impact melt in the lunar highlands crust

How much impact melt is incorporated in the lunar highlands crust? The database of lunar craters ≥20 km in diameter compiled by Head et al. (2010b) includes 30 basins >300 km in diameter. Using the impact melt volume scaling laws of Cintala and Grieve (1998), we estimate that collectively these 30 basins contain on the order of 10^6 km^3 of impact melt, 10 times the total volume of mare basalts (Head and Wilson, 1992) and about 1/20th the volume of the lunar crust. These basins, and the massive impact melt sheets that they contain, are distributed widely across the Moon (Fig. 15). Understanding the geology and the petrology of the 10^8 km^3 of massive impact melt deposits occurring in these basins—work we have just begun to undertake here—is critically important for understanding the lunar highlands crust.

5. Conclusions

Basin-forming impacts forming craters >300 km in diameter produce enormous volumes of impact melt (~10^9 km^3). We have investigated the geology and petrology of these huge volumes of impact melt, focusing especially on the relatively young lunar basin Orientale, and conclude that:

1. Models of impact melt production combined with recent geologic and topographic data suggest that most impact melt (~2/3) produced by the Orientale-forming impact occurs in a ~15.5 km thick (solidified to ~14 km thick) impact melt sea ~350 km in diameter with a volume of ~10^6 km^3.
2. The Orientale basin impact melt sea has probably undergone large-scale igneous differentiation, since terrestrial impact melt sheets (such as Manicouagan, Sudbury, and Morokweng) a tenth of its thickness and a hundredth of its volume are known to have differentiated.
3. A modeled cumulate stratigraphy (occurring below a quench crust and anorthositic fallback breccia, with an ~8 km thick layer of norite overlying a ~4 km thick pyroxenite with a basal ~2 km thick layer of dunite) produced by equilibrium crystallization of a homogenized melt sea is consistent with remotely-sensed norite excavated from ~4 km depth in the central peak of Maunder crater.
4. Small basins melting mainly the lunar crust produce initially plagioclase-saturated melt seas that first crystallize negatively buoyant anorthosite and therefore develop inverted density structures. The noritic floor of the South-Pole Aitken basin may be the roof of a differentiated impact melt lake; crater central peaks superposing SPA may excavate a massive layer of pyroxenite.
5. Thirty lunar basins >300 km in diameter together contain on the order of 10^8 km^3 of impact melt, one-twentieth the total volume of the lunar crust. Impact melt differentiates excavated from these solidified impact melt seas may well be represented in the lunar sample suite, mimicking pristine highland plutonic lithologies.
6. High-resolution compositional and gravity data collected by the M3 instrument on India’s Chandrayaan-1 mission and the ongoing LRO and GRAIL missions make it possible to probe the structure of the lunar highlands crust in unprecedented detail; analyses based on these datasets, such as constraining the cumulative stratigraphy of melt seas from careful compositional analyses of crater central peaks, will further our understanding of the geology and petrology of massive impact melts.

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References


