Short Communication

Impact melt differentiation in the South Pole-Aitken basin: Some observations and speculations

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ABSTRACT

The stratigraphy of the South-Pole Aitken basin (SPA) interior is consistent with that of a massive impact melt sheet that differentiated to form cumulates. Spectroscopic and geophysical constraints on the stratigraphy of SPA suggest a ~12.5 km thick layer of norite above ultramafic pyroxenite and dunite layers. A similar stratigraphy is produced from differentiation by crystal settling of a ~50 km thick impact melt sheet (lunar impact melt sheets > 10 km thick likely undergo differentiation by crystal settling) formed by an oblique impact (and thus containing ~20 vol. % crustal material). We propose that impact melt differentiation can account for geophysical (nonzero crustal thickness) and geochemical (~2 ppm Th) anomalies in SPA.

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1. Introduction

The interior of the lunar South Pole-Aitken basin (SPA) is more mafic than typical lunar highlands crust (e.g., Metzger et al., 1974; Head et al., 1993; Pieters et al., 2001): nonmare basin floor materials are generally noritic (Pieters et al., 2001; Uemoto et al., 2011) and contain an average of 10 wt. % FeO (Jolliff et al., 2000). Although the SPA interior is a mafic anomaly, it is not anomalously mafic. After all, SPA is the largest (undisputed) lunar impact basin (Wilhelms, 1987). The extent of SPA-associated gravity anomalies indicates that the SPA transient cavity measured ~800 km in diameter (Wieczorek and Phillips, 1999; Potter et al., 2012). The depth of crater excavation is about 1/10th the diameter of the crater transient cavity for complex craters and small basins (Croft, 1985); extrapolation of this crater scaling relationship (which seems to apply to basins < 1000 km in diameter (Wieczorek and Phillips, 1999)) indicates that the SPA-forming impact ought to have excavated ~100 km into the Moon, through the <50 km thick plagioclase-rich crust (Wieczorek et al., 2013) into the plagioclase-poor, olivine- and pyroxene-rich (Khan et al., 2007) upper mantle. Yet the floor of SPA is (lower crustal?) norite rather than (mantle) olivine pyroxenite, and this norite is evidently not just a superficial veneer—the magnitude of gravity anomalies in the SPA interior suggests a ~12.5 km thick layer of norite (Wieczorek et al., 2013). The SPA interior is anomalously feldspathic.

Why is the SPA interior so feldspathic? Perhaps this is the result of post-SPA modification. SPA seems to be the oldest (recognizable) lunar impact basin (Wilhelms, 1987); perhaps crust-derived ejecta from later lunar basins buried (and diluted) mafic materials in the SPA interior. The ejecta thickness and mixing calculations of Petro and Pieters (2004) indicate that probably <1 km (and certainly <1.5 km) of basin ejecta has been emplaced on SPA; the mixing fraction of this foreign material with primary floor material could approach ~50% in SPA interior deposits. Although mixing of plagioclase-rich ejecta with mantle pyroxenite accounts for the noritic composition of the SPA interior, it does not account for its crustal thickness: the 1 km of plagioclase-rich material emplaced by post-SPA impacts is an order of magnitude less than the inferred norite layer thickness of ~12.5 km (Wieczorek et al., 2013). It also seems unlikely that the noritic floor of SPA formed from post-SPA mare (or cryptomare) infill—noritic floor materials (Pieters et al., 2001) display hummocky, non-volcanic morphology; clinopyroxene-poor norite is an unlikely melt composition; and accumulated lava flow thicknesses in other lunar basins are generally <1.5 km (DeHon and Waskom, 1976), again an order of magnitude less than the inferred norite layer thickness of ~12.5 km.

We conclude that the feldspathic SPA interior is primary, a direct outcome of the SPA-forming impact. What impact processes might result in an anomalously feldspathic basin interior? Perhaps SPA did not excavate as deeply as proportional scaling models predict (Wieczorek and Phillips, 1999); perhaps the SPA-forming impact was very oblique, as suggested by the elliptical outer rim structure (Garrick-Bethell and Zuber, 2009), thus excavating only shallowly into the lunar crust (Schultz, 1997). Shallow excavation is also favored by feldspathic SPA ejecta (Wieczorek and Phillips, 1999), although Yamamoto et al. (2012) have found (mantle-derived?) olivine in craters superposing the margins of SPA. Although it is certainly possible that the feldspathic SPA interior is the result of shallow excavation, we do not consider this explanation further. Instead, we
investigate whether the relatively feldspathic interior of SPA can be attributed to impact melting, a shock process as volumetrically important as excavation in large basins (Cintala and Grieve, 1998).

Impact melting alone is not an explanation for the relatively feldspathic interior of SPA: the depth of melting considerably exceeds the depth of excavation for SPA-scale impacts (Cintala and Grieve, 1998), so the bulk composition of SPA impact melt is more mafic than the bulk composition of SPA ejecta. But impact melt can undergo igneous differentiation (Grieve et al., 1991), concentrating feldspathic components: Morrison (1998) hypothesized that the SPA impact melt sheet had differentiated by crystal settling to form ultramafic cumulates overlaid by a plagioclase-enriched residuum. According to this interpretation, the noritic interior of SPA is a (late-stage) impact melt sheet; (2) the critical thickness above which lunar impact melt sheets undergo differentiation by crystal settling; and (3) the crystallization sequence and cumulate stratigraphy of the differentiated SPA melt sheet, which closely matches our previously established constraints on the SPA subsurface. Finally, we discuss the implications for the geophysics and geochemistry of the SPA interior as well as SPA sample return missions.

2. Spectroscopic and geophysical constraints on the SPA subsurface

Surface nonmare lithologies in the SPA interior are generally noritic (Pieters et al., 2001; Uemoto et al., 2011). What lithologies exist at depth in the SPA subsurface?

2.1. Constraints from central peak mineralogy

The central peaks of complex craters superposing the SPA interior exhume material from tens of kilometers deep in the SPA subsurface. Many authors (e.g., Tompkins and Pieters (1999), Nakamura et al. (2009), Yamamoto et al. (2012)) have characterized the mineralogy of such central peaks. In order to profile the mineralogy of the SPA subsurface, we select twelve complex craters (Fig. 1a) with previously characterized central peaks according to several criteria: selected complex craters (1) should superpose the SPA interior as defined by low topography and high FeO abundance (Garrick-Bethell and Zuber, 2009); (2) should have a well-defined central peak not buried by mare infill (Yingst and Head, 1999); (3) should not display unusual morphology which could confuse determination of central peak sampling depth; (4) should not superpose large basins such as Schrödinger or Apollo which probably confuse the primary stratigraphy of the SPA subsurface. Next, we calculate the sampling depths of these twelve central peaks according to the formula $u = 0.02D_1^{0.85}$, where $u$ is the stratigraphic uplift of a central peak of a complex crater with (present) rim diameter $D_1$ (Cintala and Grieve, 1998). The stratigraphic uplift is taken to be the sampling depth of the central peak below the unmodified, primary floor of SPA, as the depth of these twelve central peaks below the present floor of SPA is on the same order of magnitude ($\sim 1$ km) as the amount of foreign material introduced into SPA by large post-SPA basin-forming impacts (Petro and Pieters, 2004). Fig. 1b shows an SPA stratigraphic column color-coded according to central peak mineralogy. Note that we assume radial continuity and symmetry in subsurface strata (i.e., stratigraphic horizons run parallel to the present SPA surface), likely the case for lithologies derived from a melt sheet (or a magma ocean) that underwent differentiation by crystal settling.

What can we infer about the subsurface stratigraphy of SPA from Fig. 1b (bearing in mind the caveat that spectral data incompletely constrain rock type)? (1) Olivine-bearing lithologies (probably olivine-rich lithologies (Yamamoto et al., 2012)) are present only at depth in the SPA subsurface, as evidenced by the central peak of the 184 km diameter Zeeman crater and the olivine-bearing peak rings of the Schrödinger basin (Yamamoto et al., 2012). (2) Orthopyroxene-bearing lithologies overlie olivine-bearing lithologies, and the proportion of orthopyroxene decreases with decreasing depth. However, exceptions to this trend exist

![Fig. 1. Spectroscopic and geophysical constraints on SPA subsurface stratigraphy: (a) twelve complex craters with previously characterized central peaks (Tompkins and Pieters, 1999; Nakamura et al., 2009; Moriarty et al., 2011; Yamamoto et al., 2012) mapped on LOLA-derived topography; (b) constraints on subsurface stratigraphy from central peak mineralogy; the dashed red line at 12.5 km depth corresponds to the crust-mantle boundary inferred from (c) the average (GRAIL-derived) crustal thickness (Wieczorek et al., 2013) in a relatively unmolested region of the SPA interior. SPA rim outlines in (a) and (c) are from Garrick-Bethell and Zuber (2009). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)]
(anorthosite in the central peak of Alder? (Moriarty et al., 2011)), and the literature is equivocal on the proportion of orthopyroxene in the most mafic central peaks: e.g., does the central peak of Lyman contain norite (Tompkins and Pieters, 1999) or pyroxenite (Nakamura et al., 2009)? We favor more recent measurements indicating ultramafic compositions (Nakamura et al., 2009; Klima et al., 2011), since, as Nakamura et al. (2009) point out, derived central peak FeO concentrations of ~15 wt. % are too ferroan for norite. What about clinopyroxene-bearing lithologies identified in central peaks (Tompkins and Pieters, 1999)? We take these to be secondary, derived from scattered mare basalts plutons or dikes (Purucker et al., 2012) not pictured in the primary stratigraphic column in Fig. 1b. Mare basalts extrusives and cryptomare in SPA (Yingst and Head, 1999; Whitten and Head, 2013) indicate local mare magmatism, and floor-fractured craters distributed throughout the SPA interior point to local mare plutonism (Jozwiak et al., 2012).

2.2. Constraints from gravity data

Plagioclase-rich norite apparently transitions to orthopyroxene-rich pyroxenite at some depth in the SPA subsurface: floor materials are norritic (Pieters et al., 2001; Uemoto et al., 2011), whereas the central peaks of large complex craters contain pyroxenite (Nakamura et al., 2009). This transition depth is poorly constrained by VNIR spectroscopy (Fig. 1b) but well-constrained by gravity data: the depth at which low density norite transitions to mantle-density pyroxenite is simply the crustal thickness (assuming a noritic crustal composition). Wieczorek and Phillips (1999) determined that the noritic crust in the SPA interior is ~40 km thick using low-degree gravity data. This crustal thickness is inconsistent with pyroxenite exhumed from <40 km depth by widely distributed complex craters (Nakamura et al., 2009). We reassess the crustal thickness of the SPA interior (Fig. 1c) using a recent GRAIL-derived crustal thickness model (model 1 of Wieczorek et al. (2013)). The average (mean) crustal thickness in a relatively unmodified (large complex crater-free) region of the SPA interior (outlined in Fig. 1c) is 13.7 km (with a standard deviation of 1.05 km). Post-SPA basin ejecta must comprise some of this crustal thickness: Petro and Pieters estimate that post-SPA basin ejecta amounts to at most ~1.2 km (in an SPA interior region north of our outlined region). Therefore, we take the transition depth from norite to pyroxenite in the SPA interior to be 12.5 km (the dashed red line in Fig. 1b).

3. Modeling impact melt differentiation in SPA

Having outlined the stratigraphy of the SPA subsurface (Fig. 1b), we proceed to model the cumulative stratigraphy produced by impact melt differentiation.

3.1. Bulk composition of SPA impact melt

Morrison (1998) postulated that noritic lithologies in the SPA interior were impact melt differentiates. Was the SPA impact melt sheet sufficiently aluminous to form 12.5 km of norite? Not by the estimate of Warren et al. (1996), who calculated that ~99% of the SPA impact melt sheet was mantle-derived. In this case, the SPA melt sheet (like the lunar mantle) contains <1 wt. % Al2O3; forming 12.5 km of norite (with ~20 wt. % Al2O3) would require a parent melt sheet several hundred km in thickness with a volume of perhaps 5 × 108 km3 (~2.5% of the Moon). This inferred melt sheet volume is considerably in excess of estimates of total SPA melt volume ranging from ~7 × 108 km3 to ~1 × 109 km3 (Morrison, 1998; Kring et al., 2012) based on melt scaling and hydrocode models (Cintala and Grieve, 1998; Abramov et al., 2012; Potter et al., 2012). Rather than conclude from this gross inconsistency that noritic intra-SPA lithologies are not impact melt differentiates, we suggest that the oblique SPA-forming impact (Schultz, 1997) produced impact melt containing a greater proportion of crustal material c (the ratio of crust vol. % to crust vol. % + mantle vol. %) than suggested by Warren et al. (1996). Hurwitz and Kring (2013) suggest alternatively that the SPA-forming impact melted aluminous, pre-cumulate overturn upper mantle (Hess and Parmentier, 1995). One difficulty with this hypothesis is that crust and upper mantle isotherms uplifted by the cumulative overturn process could exert total viscous relaxation of SPA topography (Bratt et al., 1985).

We calculate c for impact angles of 90° (vertical) and 45° (oblique) by determining the proportion of a 40 km thick crustal layer (Wieczorek et al., 2013) incorporated in the SPA melt cavity. What is the shape of this melt cavity? In vertical impacts, impact melt is generated in a buried sphere (depth of burial/radius ~0.8) (Barr and Citron, 2011). In oblique impacts with impact angles < ~45°, impact melt is generated in an approximately hemispherical region (Pierazzo and Melosh, 2000). The radius of this buried sphere or hemisphere can be determined from the fact that its volume is (by definition) equal to the volume of melt produced by the SPA-forming impact (~1.0 × 108 km3 and ~6.8 × 107 km3 for vertical and oblique impacts respectively, calculated using the scaling model of Abramov et al. (2012) with the parameters of Kring et al. (2012)). For an impact angle of 90°, c ~0.05; for an impact angle of 45°, c ~0.2, four times greater. (The curvature of the Moon increases c by about 10% for SPA-scale impacts.) The oblique SPA-forming impact generated a hemispherical melt cavity that incorporated a greater proportion of crustal material than would be the case in a vertical impact. Voluminous feldspathic SPA impact melt differentiates provide additional evidence that the SPA-forming impact was oblique.

Basin excavation and modification processes alter the bulk composition of SPA impact melt. Warren et al. (1996); Hurwitz and Kring (2013) take the basin excavation flow to skim almost all crustal material from the melt cavity, decreasing c to nearly 0. Yet pre-excitation convective melt mixing, melt mixing due to the excavation flow, trapping of crustal material beneath the projectile (Stewart, 2011), ejected impact melt flowing back into the melt sheet (Ivanov et al., 2010), and an oblique (Schultz, 1997) or anomalously shallow excavation flow (Wieczorek and Phillips, 1999) may all conspire to decrease c only slightly. See also the discussion in the supplementary information of Nakamura et al. (2012). In the following, we take c to be unchanged by the excavation flow to provide an upper bound on the thickness of noritic melt differentiates. We take the thickness of the SPA melt sheet to be 50 km (approximately the thickness of an elliptical prism with the dimensions of the SPA interior from Garrison-Bethell and Zuber (2009) holding 108 km3 of melt).

3.2. Does the SPA melt sheet differentiate?

Will a ~50 km thick SPA impact melt sheet undergo igneous differentiation by crystal settling? If so, the timescale of melt sheet cooling and solidification must exceed the timescale on which crystals settle through the melt sheet. Since the timescale of cooling, proportional to the square of the melt sheet thickness (L2), grows faster than the timescale of settling, which is proportional to the melt sheet thickness L, there must be some critical melt sheet thickness (Lcr) for which these two timescales are equal. All melt sheets thicker than Lcr will undergo igneous differentiation by crystal settling. What is this thickness, and is it exceeded in impact melt sheets much thinner than the lunar magma ocean?

Although it is in principle possible to calculate Lcr, it is difficult to determine appropriate average values for parameters such as melt viscosity and crystal diameter. However, Lcr itself is constrained by
terrestrial impact structures. Recent drilling reveals that the impact melt sheet of the 90 km diameter Manicouagan impact crater in Canada, long thought to be uniformly ~0.4 km thick and undifferentiated, in fact contains markedly thicker sections (~1.5 km thick) hosted in a fault-bounded trough (Spray and Thompson, 2008), and, moreover, that this thicker impact melt has undergone differentiation by crystal settling (Spray and Thompson, 2008). Our interpretation of these facts is that the (terrestrial) $L_{cr}$ is between 0.4 km and 1.5 km. Indeed, the ~3 km thick impact melt sheet of the ~250 km (outermost rim) diameter Sudbury impact crater in Canada has also undergone igneous differentiation (Grieve et al., 1991). The lunar $L_{cr}$ is about six times (i.e., the ratio of lunar to terrestrial gravitational acceleration) the terrestrial $L_{cr}$; lunar impact melt sheets > ~10 km thick will differentiate, all other things being equal between the Earth and the Moon. Not all other things are equal, in particular melt viscosity: SPA melt, being considerably more mafic and less aluminous than Manicouagan melt (Spray and Thompson, 2008) is substantially less viscous, which will decrease the lunar $L_{cr}$ (by the viscosity ratio). In any case, the lunar $L_{cr}$ is less than ~10 km and the SPA impact melt sheet is ~50 km thick. If the Manicouagan melt sheet differentiated by crystal settling, the SPA melt sheet did as well. Alternatively, Zieg and Marsh (2005) have argued that impact melt sheets differentiate by a different mechanism; see also the discussion in Section 4.1.1 of Vaughan et al. (2013).

### 3.3. Crystallization sequence and final megastratigraphy of the SPA melt sheet

We now determine the crystallization sequence and cumulate stratigraphy of the SPA melt sheet from the proportion of crustal material in impact melt $c$. We take the lunar crust and mantle to be composed entirely of plagioclase, olivine, and orthopyroxene, as the lunar minerals ilmenite and clinopyroxene were probably minor phases in target post-cumulate overturn mantle: mare basalts in SPA are low-Ti (Yingst and Head, 1999), indicating ilmenite-poor mantle source regions; inversions of seismic velocity data indicate that the lunar mantle generally contains < 10 vol. % clinopyroxene (Khan et al., 2007). SPA melt (c likely less than ~0.2) is olivine- and orthopyroxene-rich. The equilibrium crystallization sequence of an olivine- and orthopyroxene-rich mixture of olivine, orthopyroxene, and plagioclase (SPA melt) determined from the low-pressure An-Fo-Qz phase diagram is as follows: olivine; orthopyroxene with resorption of olivine; plagioclase and orthopyroxene. Since the SPA melt sheet likely differentiated by crystal settling, and olivine and orthopyroxene are denser than coexisting liquid, this crystallization sequence is effectively a stratigraphic sequence: melt differentiation produces dunite overlain by pyroxenite and norite. The proportions of norite, pyroxenite, and dunite layers can be calculated by mass balance from assumed crust and mantle compositions (see Vaughan et al. (2013) for details). Briefly, the initial vol. % of olivine (the vol. % of olivine in the mantle, possibly ~25 vol. % (Wieczorek et al., 2006), multiplied by 1-c) corresponds to the final vol. % of dunite and the initial wt. % alumina (~25 wt. % Al$_2$O$_3$ (Warren, 1990) multiplied by the weight fraction of crust) must correspond to the final wt. % alumina (~20 wt. % Al$_2$O$_3$ multiplied by the weight fraction of norite).

Fig. 2a shows the cumulative stratigraphy produced by impact melt differentiation if the proportion of crustal material in SPA melt c is unchanged by excavation. We conclude from the remarkable resemblance of Fig. 2a and Fig. 2b that, as Morrison (1998) postulated, the stratigraphy of the SPA interior is consistent with the cumulative stratigraphy of a massive differentiated impact melt sheet.

### 4. Discussion

#### 4.1. Geophysical implications

How should the ~12.5 km crustal thickness of the SPA interior be interpreted? We have suggested in this paper that this crustal thickness corresponds not to lunar primary crust, but rather to the noritic residuum formed by impact melt differentiation (tertiary crust; Taylor, 1989). We outline our scenario for the genesis of intra-SPA crust: (1) a basin-forming impact melts preexisting primary crust and mantle; (2) some crustal melt is mixed with mantle melt and escapes excavation; (3) impact melt differentiation reconstitutes this crustal material to form tertiary crust. Nonzero crustal thickness does not mean that basin-forming impacts did not totally excavate the local primary crust. Models of basin excavation (Wieczorek and Phillips, 1999) should be reconsidered in light of this possibility. The thickness of (primary) crust in the interiors of Serenitatis, Imbrium, and SPA may well be 0 km, preserving proportional scaling for the largest lunar impact basins.

#### 4.2. Geochemical implications

Impact melt differentiation should concentrate incompatible trace elements (ITEs) in the noritic residuum of SPA melt. This may explain relatively high Th concentrations of ~2 ppm in the SPA interior (Jolliff et al., 2000). If Th is perfectly incompatible, the top ~1 km of primary SPA material (~1/50th the volume of the SPA melt sheet), the SPA melt residuum, contains all the bulk melt Th (approximated as the bulk Moon Th, 125 ppb (Taylor, 1982)). This top ~1 km of primary SPA material is mixed with ~50% of nearly Th-free upper crustal basin ejecta (Petro and Pieters, 2004). Therefore, average Th concentrations in the SPA interior ought to be ~3 ppm, similar to the ~2 ppm Th observed. We conclude that the chemistry of the SPA interior is as consistent with an ITE-enriched noritic residuum formed by impact melt differentiation as noritic, KREEPy “LKFM” lower crust (Jolliff et al., 2000).
4.3. Implications for sampling SPA

Sample return from the SPA interior is a top lunar science priority (Jolliff et al., 2010), as SPA impact melts are ideal materials for constraining the age of the SPA-forming impact and testing the late lunar cataclysm hypothesis. If the noritic SPA floor is an impact melt differentiate, then samples from nearly any nonmare location in the SPA interior can be used to date the SPA-forming impact. The landing ellipse for an SPA sample return mission may be 1940 × 1440 km, the dimensions of the SPA interior (Garrick-Bethell and Zuber, 2009).

5. Conclusions

1. Spectroscopic and geophysical constraints on the stratigraphy of the SPA interior suggest a ∼12.5 km thick layer of norite above ultramafic pyroxenite and dunite layers.
2. A similar stratigraphy is produced from igneous differentiation by crystal settling of a ∼50 km thick impact melt sheet (lunar impact melt sheets > 10 km likely undergo differentiation by crystal settling) formed by an oblique impact (and thus containing ∼20 vol. % crustal material).
3. The nonzero crustal thickness in the SPA interior may represent the thickness of the noritic residuum formed by impact melt differentiation, not that of primary crust.
4. The chemistry of the SPA interior is consistent with an ITE-enriched noritic residuum formed by impact melt differentiation.

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