FeO and TiO$_2$ concentrations in the South Pole-Aitken basin: Implications for mantle composition and basin formation

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Abstract. We use newly developed spectroscopic methods for determining FeO and TiO$_2$ abundances to study the huge (2500 km diameter) South Pole-Aitken basin. The floor of this basin shows iron and titanium values which are weakly correlated and largely range from 7 to 14 wt% FeO and from 0.5 to 1.5 wt% TiO$_2$. This composition is consistent with an approximate 1:1 mixture of lower crustal material like that found elsewhere on the Moon and mantle rock containing 10-16 wt% FeO and < 0.1 wt% TiO$_2$. Although mantle rock appears to compose about 50% of the floor deposits, models of the formation of large impact basins predict that virtually all the floor materials ought to be derived from the mantle.

1. Introduction

The ejecta deposits of impact craters are largely derived from material which was below the surface at the time of the impact event. Therefore we can use the impact craters that cover the surface of the Moon to probe the composition of the lunar subsurface. The largest of these features are the multiring basins, hundreds to thousands of kilometers across [Wilhelms, 1987]. The largest basin on the Moon is the South Pole-Aitken (SPA) basin, which at 2500 km in diameter and over 12 km deep, is the largest, deepest basin known in the solar system [Spudis et al., 1994]. The existence of a large basin on the southern farside of the Moon was hinted at by the presence of large mountains on the nearside, close to the south pole of the Moon [Hartmann and Kuyper, 1962]. A large topographic low corresponding to the site of SPA basin was first recognized in Soviet Zond 6 photography [Rodionov et al., 1971], and large massifs on the farside were mapped by Wilhelms et al. [1969]. The complete basin was first mapped by Stuart-Alexander [1978]. The basin was detected as a compositional anomaly in the Apollo orbital geochronological data [Metzger et al., 1977] and in spectral data collected during the Galileo Earth-Moon flyby [Belton et al., 1992; Pieters et al., 1993]. Its dimensions and character were clearly revealed by the global topography and compositional data provided by the Clementine mission. Such a large basin must have excavated to substantial depths, perhaps through the entire crustal column and into the mantle [Belton et al., 1992; Pieters et al., 1993]. Thus we can use the SPA basin as a crust and mantle probe.

The Clementine mission provided us with a global multispectral data set for the Moon. Lucey et al. [1995] developed a method for extraction of Fe contents from multispectral imaging of the Moon, and this method was improved upon and further tested by Blewett et al. [1997] and Lucey et al. [1998]. We applied the improved technique to the low-resolution data set of Lucey et al. [1995] to derive an iron abundance map for the SPA region with 35 km spatial resolution and about 1 wt% precision and accuracy. Lucey et al. [1996] and Blewett et al. [1997] presented a similar method to derive titanium abundances for the Moon using Clementine data. We used these methods to create a titanium abundance map of the SPA region, again with about 1 wt% precision and accuracy. These data show that the composition of the interior of the SPA basin provides important constraints on models for the formation of large craters and basins. The SPA basin may allow us to probe the composition of the lunar mantle directly for the first time.

2. Lunar Samples and the Nature of the Lunar Crust

The concentrations of TiO$_2$ and FeO are useful in distinguishing among major lunar rock types (Figure 1), especially when combined with the incompatible element Th, for which we have regional data of limited extent from Apollo 15 and 16 missions and will have global data from the 1998 Lunar Prospector mission. These systems form a reference frame for our analysis of the Clementine data. The characteristics of the main types of lunar rocks are summarized in Table 1 and Figure 1.

Mare basalts are classified largely on the basis of their TiO$_2$ contents which vary in lunar samples (Apollo, Luna, and meteorites) from several tenths of a weight percent to about 13 wt%. These basalts also contain Th at an average of less than 1 ppm. Mineralogically, mare basalts are characterized by abundant augite. The lunar highlands, however, are composed of a diverse suite of igneous rocks [e.g., Warren, 1985], including anorthosites, trachytes, norites, and basalts. The most abundant lunar anorthosites are called ferroan anorthosites owing to their high Fe/Mg ratios relative to the rest of the lunar highland igneous rocks. They are characterized by extremely low amounts of FeO, TiO$_2$, and Th. The main mafic silicate, present at abundances of less than 10%, is either orthopyroxene or olivine. Trachytes are much richer in mafic silicates (almost all of which is olivine) than ferroan anorthosites, but because they have high...
Figure 1. Lunar rock types can be distinguished by their FeO and TiO2 concentrations (data are chemical analyses of lunar rocks reported by Lindstrom et al. [1990]; Heiken et al. [1991], Ryder and Norman [1980], and Ryder [1985, 1993]). “LKFM” is a rock type thought to compose the lower lunar crust [Ryder and Wood, 1977; Spudis, 1984]. Estimates of mantle compositions are from Table 2. Although some inferred mantle rocks have somewhat elevated TiO2, we argue that such compositions are rare. “FAS” are rocks from the ferroan anorthosite suite, thought to represent cumulates formed during initial lunar differentiation.

MgO/FeO, troctolites contain only 4 wt % FeO on average. They have slightly more TiO2 than ferroan anorthosites. Norites are orthopyroxene-plagioclase rocks. They contain more FeO than the anorthosites and troctolites and slightly more TiO2. Norites and troctolites contain distinctly more Th than ferroan anorthosites, but their Th abundance varies widely (from 0.2 to 4 ppm, averaging slightly less than 1 ppm). KREEP rocks are defined by their enrichments in K, rare earth elements (REE), P, and other incompatible lithophile elements. Although some lunar highland basaltic rocks are pristine, meaning that they are compositionally unaltered since their crystallization from an endogenous igneous melt, most are impact melt breccias. They have FeO contents that overlap those of norites, but significantly higher TiO2 contents; Th contents are highly variable. KREEP basalts and Apollo 14 breccias contain >10 ppm Th. Finally, low-K Fra Mauro basaltic impact melts (nicknamed “LKFM”) contain 0.4 to 7.5 ppm Th, but some samples identified as “KREEP-poor” [Lindstrom et al., 1990] contain 0.4 to 2.2 ppm, the same as pristine norites. LKFM melt rocks have been interpreted as lower crustal rocks, melted and excavated during basin formation.

Table 1. Average concentrations of FeO and TiO2 and Range in Th in Principal Lunar Rock Types

<table>
<thead>
<tr>
<th></th>
<th>FeO, wt %</th>
<th>TiO2, wt %</th>
<th>Th, ppm</th>
<th>Dominant Mafic Mineral</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mare basalts</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Apollo 11 high-Ti</td>
<td>19.2</td>
<td>10.4</td>
<td>0.8 (0.7-3.9)</td>
<td>augite</td>
</tr>
<tr>
<td>Apollo 17 high-Ti</td>
<td>18.8</td>
<td>11.9</td>
<td>0.4 (0.2-0.7)</td>
<td>augite</td>
</tr>
<tr>
<td>Luna 16</td>
<td>18.7</td>
<td>4.8</td>
<td>0.4-2.4</td>
<td>augite</td>
</tr>
<tr>
<td>Apollo 12</td>
<td>20.5</td>
<td>3.6</td>
<td>0.9 (0.7-1.3)</td>
<td>augite</td>
</tr>
<tr>
<td>Apollo 15</td>
<td>19.9</td>
<td>2.0</td>
<td>0.5 (0.4-2.6)</td>
<td>augite</td>
</tr>
<tr>
<td>Luna 24</td>
<td>21.6</td>
<td>1.0</td>
<td>0.2</td>
<td>augite</td>
</tr>
<tr>
<td>Apollo 17 very-low Ti</td>
<td>17.7</td>
<td>0.6</td>
<td>--</td>
<td>augite</td>
</tr>
<tr>
<td>Nonmare rocks</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ferroan anorthosite</td>
<td>1.0</td>
<td>0.08</td>
<td>0.04</td>
<td>orthopyroxene/olivine</td>
</tr>
<tr>
<td>Norites</td>
<td>8.2</td>
<td>0.30</td>
<td>0.9 (0.2-1.7)</td>
<td>orthopyroxene</td>
</tr>
<tr>
<td>Troctolites</td>
<td>4.1</td>
<td>0.2</td>
<td>0.3 (0.2-0.6)</td>
<td>olivine</td>
</tr>
<tr>
<td>Gabbro-norites</td>
<td>11.2</td>
<td>1.3</td>
<td>0.8 (0.5-1.0)</td>
<td>augite, orthopyroxene</td>
</tr>
<tr>
<td>KREEP basalt</td>
<td>9.7</td>
<td>2.0</td>
<td>10</td>
<td>low-Ca pyroxene</td>
</tr>
<tr>
<td>High-K KREEP</td>
<td>10.0</td>
<td>1.7</td>
<td>18</td>
<td>low-Ca pyroxene</td>
</tr>
<tr>
<td>LKFM</td>
<td>8.8</td>
<td>1.4</td>
<td>6.1 (4.7-7.5)</td>
<td>low-Ca pyroxene</td>
</tr>
<tr>
<td>KREEP-poor LKFM</td>
<td>7.8</td>
<td>1.8</td>
<td>1.6 (0.4-2.2)</td>
<td>low-Ca pyroxene</td>
</tr>
<tr>
<td>Aluminous LKFM</td>
<td>7.5</td>
<td>0.8</td>
<td>4.2 (2.7-4.5)</td>
<td>low-Ca pyroxene</td>
</tr>
</tbody>
</table>

Data from Lindstrom et al. [1990], Heiken et al. [1991], Ryder and Norman [1980], and Ryder [1985, 1993].

*Ranges are given in parentheses.*
Table 2. Estimates of the FeO, TiO$_2$, and Th Contents and Mineralogy of the Lunar Mantle

<table>
<thead>
<tr>
<th>Estimating from bulk moon</th>
<th>FeO, wt %</th>
<th>TiO$_2$, wt %</th>
<th>Th, ppm</th>
<th>Mineralogy$^\dagger$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Bulk moon crust</td>
<td>17.5</td>
<td>0.1</td>
<td>0.03</td>
<td>ol, opx</td>
</tr>
<tr>
<td>2. Bulk moon crust</td>
<td>13.8</td>
<td>0.1</td>
<td>0.03</td>
<td>ol, opx</td>
</tr>
<tr>
<td>3. Bulk moon crust</td>
<td>11.4</td>
<td>0.3</td>
<td>0.03</td>
<td>ol, opx</td>
</tr>
<tr>
<td>Estimates from mare basalt</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4. Very-low Ti basalt source-1</td>
<td>17.7</td>
<td>0.13</td>
<td>&lt; 0.1</td>
<td>ol, opx</td>
</tr>
<tr>
<td>5. Very-low Ti basalt source-2</td>
<td>19.7</td>
<td>0.09</td>
<td>&lt; 0.1</td>
<td>ol, opx</td>
</tr>
<tr>
<td>6. Apollo 15 green glass</td>
<td>12.3</td>
<td>0.10</td>
<td>&lt; 0.05</td>
<td>ol, opx</td>
</tr>
<tr>
<td>7. Low-Ti basalt source-1</td>
<td>18.0</td>
<td>0.34</td>
<td>0.1</td>
<td>ol, opx</td>
</tr>
<tr>
<td>8. Low-Ti basalt source-2</td>
<td>20.0</td>
<td>0.28</td>
<td>0.1</td>
<td>ol, opx</td>
</tr>
<tr>
<td>9. High-Ti source</td>
<td>17.0</td>
<td>1.5</td>
<td>0.1</td>
<td>ol, aug, ilmenite</td>
</tr>
</tbody>
</table>


$^\dagger$Th in 4–9 estimated from bulk Th in basalt, assuming 10% partial melting and no assimilation.

Abbreviations are ol, olivine; opx, orthopyroxene; aug, augite; ilm, ilmenite.

3. The Composition of the Lunar Mantle

No samples of the lunar mantle are known, so its composition must be inferred indirectly. We present several estimates here to use as guidelines for comparison with our spectroscopic measurements of the composition of the floor of the SPA basin. Formation of the lunar crust involved removing crystals from melt with the composition of the bulk Moon. This differentiation left behind a mantle depleted in the crustal components, so that estimates of average mantle composition can be derived from mass balance calculations using compositions of the bulk Moon and the crust. Our calculations assume a feldspathic crust 71 km thick [Taylor, 1982] (10.5% of the lunar volume) and use three different models of the Moon’s bulk composition [Taylor, 1982; Delano, 1984; Wenke et al., 1977]. These calculations yield an estimated bulk mantle FeO content of between 11.4 and 17.5 wt %, and TiO$_2$ content of < 0.3 wt %, and a concentration of only 0.03 ppm Th (Table 2).

Experiments at high temperatures and pressures can be used to infer the FeO and TiO$_2$ contents of the minerals composing the mare basalt source regions in the lunar mantle. A key assumption in these estimates is that we have unmodified samples of magmas formed by partial melting; that is, the magmas were not modified by fractional crystallization or assimilation as they migrated from the mantle to the surface. The experiments involved in the studies went to great lengths to maximize this possibility by choosing samples least likely to be altered (high Mg/Fe, rocks at the high-Mg end of fractionation trends). The second important assumption is that the pressure and temperature at which a lava composition becomes saturated with two or more phases represent the pressure and temperature of its origin. This assumption is reasonable but almost impossible to test, and it is possible that melting to produce some lunar basalt took place over a range of pressures [Longhi, 1992]. Mantle composition estimates 4 through 9 in Table 2 were chosen to reflect the full range of mare basalt types.

The results from the high-T, high-P experiments indicate that lunar mantle olivines are magnesian (Fo$_{75}$–Fo$_{94}$), most containing about 21 wt % FeO and no TiO$_2$. Pyroxenes (low-Ca in all but the high-Ti basalt sources) have lower FeO (9-14 wt %) and contain 0.1 to 2.2 wt % TiO$_2$. In order to use these mineral data to constrain the mantle compositions we must assume the relative abundances of olivine and pyroxene. We bracketed the abundances of these phases by assuming either 50% of each or 75% olivine and 25% pyroxene. This is consistent with the expected abundances of olivine and pyroxene deposited in a lunar magma ocean [e.g., Taylor, 1982]. We also added the appropriate mare basalt magma composition back into the mantle source region, assuming 5% partial melting. With the exception of the very magnesian Apollo 15 green glass source, our estimated mantle composition based on the experimental petrology contains between 17 and 20 wt % FeO, not very different from the mare basaltos (Figure 1). The mantle sources that gave rise to high-Ti basalt contain >1 wt % TiO$_2$. We can estimate the Th abundance in the mare basalt source areas from the Th concentrations in mare basalts, by assuming 10% partial melting, and essentially completely partitioning of Th into the magmas. In estimating Th from the compositions of mare basalts, we must also assume that no Th was acquired by mare magmas by assimilation of crustal materials. Nevertheless, this simple calculation allows us to place an upper limit on the Th concentration in the mantle. The results suggest that the mantle contains no more than about 0.1 ppm Th (Table 2), roughly consistent with our estimate from bulk Moon compositions (within a factor of 3).

While the nature of the mare basalt source regions can be inferred from the basalts themselves, other data are required to estimate the volumetric importance of the different source regions. This is particularly important because the large variation of TiO$_2$ in the basalts reflects a significant variation in TiO$_2$ content in the mantle. Areal abundances of different basalt types derived from remote sensing is a key parameter in this estimate. The surface distribution of basalt Ti contents likely reflect a combination of the Ti content of the lunar mantle, the case of melting of mantle sources with various Ti contents, and the ease of magma transport from the mantle. Mantle sources with the lowest and highest concentrations of radioactive elements would be most prone to melt. Such sources would be richest in TiO$_2$. On the other hand, high-Ti basalt magmas have higher densities than low-Ti magmas, which would inhibit their migration through the low-density crust, leading to fewer eruptions of high-Ti mare magmas than of other types. These two effects may balance one another.
Plate 1. Images of FeO and TiO₂ for South Pole-Aitken basin. Images are in orthographic projection and centered on 45°S and 180°.
Using the methods of Lucey et al. [1996], Taylor et al. [1996] showed that on an areal basis high-Ti basalts are rare relative to medium- and low-Ti basalts. Only 10% of mare surface area contains >6 wt % TiO₂, and only 5% of the mare surfaces contain >7.5 wt % TiO₂. We conclude that to first order the areal abundance of high-Ti basalts roughly reflects the volumetric abundance of the high-Ti source regions, and so they are rare and can be neglected as a component of the lunar mantle in our analysis of the floor deposits of SPA basin. Furthermore, it is unlikely that high-Ti mantle rocks remains in the upper mantle tapped by SPA, as they would have been much denser than the underlying mantle and sunk before the formation of SPA basin [Hess and Parmentier, 1995].

4. Composition of South Pole-Aitken Basin

Using the Clementine data of Lucey et al. [1994,1995] and the methods of Lucey et al. [1996,1998] and Blewett et al. [1997], we derived images of the concentration of FeO and TiO₂ at 35 km resolution. These data are shown centered on SPA basin in Plate 1. Figure 2 plots SPA FeO versus TiO₂ with the major lunar rock typc fields and our model mantle compositional superimposed. Inspection of Figure 2 shows that the composition of the interior of SPA is unlike any major lunar rock type or model lunar mantle composition, although its ejecta (the highlands terrain immediately surrounding the basin rim) is typical of the lunar highlands (FeO of about 3-4 wt %, TiO₂ = 0.3 wt %, noritic anorthosite mineralogy [Pieters et al., 1997]). (We note that at extremely low TiO₂ contents, less than that of the landing sites, the calibration of Blewett et al. [1997] may be problematic, however, the interior of SPA is within the range of the calibration.) The elevated FeO content of the basin floor may result from exposure of rocks from the mantle of the Moon. However, the TiO₂ abundances are significantly higher than our expected mantle compositions so the basin floor cannot be pure mantle composition. These data suggest that SPA could be a mixture of approximately equal parts of typical LKFM (lower crustal composition) and low-Ti mantle rocks with 10-20 wt % FeO. This FeO content is in the range of likely mantle compositions we estimated above (Table 2). This would result in a Th concentration of about half the level in LKFM, or about -4 ppm, as is observed in the northern floor materials of SPA basin [Metzger et al., 1977].

It is possible that the correlation of TiO₂ with FeO SPA results from differentiation of the large melt sheet, as has been proposed for the Sudbury impact structure in Canada [Grieve et al., 1991]. If so, the slope of the trend suggests fractionation of olivine with a composition of >Fo₉₅, which corresponds to a molar MgO/(MgO+FeO) in the magma of >0.84, as magnesian (or more) than the source of the Apollo 15 green glasses. If this is the case, then the floor of SPA is a mixture of very magnesian mantle with lower crustal materials richer in FeO and TiO₂. Fractional crystallization of the melt sheet should have decreasing FeO and TiO₂ with depth, so detailed examination of the deposits surrounding craters and basins within SPA can test the fractional crystallization hypothesis. We have determined the composition of the ejecta from Apollo basin (480 km in diameter [Spudis, 1993] that is superposed on the floor of SPA to be 8.6 wt % FeO and 0.4 wt % TiO₂ at the low FeO and TiO₂ end of the SPA cluster in Figure 2, consistent with fractionation.

Large expanses of mantle material would mean high density material crops out on the surface, an occurrence not compatible with the slight negative free-air anomaly the basin displays in the global gravity map [Zuber et al., 1994]. However, if the melt sheet and underlying material are substantially brecciated and the melt is half mantle and half lower-density LKFM basaltic material, the effective density would be much lower than that expected for solid mantle rock.

If the interpretation of approximately 50:50 mantle-LKFM explanation is correct, the floor of SPA should have a high abundance of olivine, as was reported by Pieters et al., 1993. More recently, however, Pieters et al. [1997] showed that for a 35 km

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**Figure 2.** FeO versus TiO₂ for the interior of South Pole-Aitken basin (dots). Fields are for major rock types shown in Figure 1. The field for mantle rocks does not include the mantle source estimated for high-Ti mare basalts (see text). The SPA data lie midway between fields defined by LKFM (probably lower crustal impact melts) and mantle rocks, suggesting a roughly 1:1 mixture of those two materials. The trend of the data might define a olivine fractionation trend, but if so, the olivine was exceptionally magnesian, >Fo₉₅.
wide longitudinal strip through the northern portion of the basin, no exposures of fresh material were dominated by olivine in the manner of the central peaks of Copernicus or Tsiolkovsky. The nature of the method used in that study puts an upper limit on olivine abundance at these locations of about 30 wt%. If this upper limit is representative of the basin as a whole, then our assumptions of 75:25 olivine/pyroxene or 50:50 olivine/pyroxene in the mantle leads to a maximum mantle component of 45-60% in the basin, entirely consistent with the measurements of Pieters et al. [1997].

There are alternatives to our interpretation that mantle rock is exposed in the floor of SPA basin [Pieters et al., 1997]. One possibility is that the floor is a mixture of average highlandnorites and troctolites with low-Ti mare basalts. Ancient mare deposits are common in many highland regions of the Moon [Schultz and Spudis, 1979; Head et al., 1993] in which fragments of mare basalt are intimately mixed into the highlands regolith. Such deposits would increase FeO and TiO2, producing a trend like that shown in Figure 2. However this mixture would have Th content of < 1 ppm, substantially below the 2-4 ppm of SPA measured by the Apollo gamma-ray experiment. If the SPA floor is a mixture of mare basalts and highland rocks, it will contain significant clinopyroxene (augite) and be poor in, or free of, olivine. Pieters et al. [1997] also showed that clinopyroxene is not and abundant in the portion of the basin they examined and also noted that no ancient, partially obscured mare deposits (cryptomare) are present anywhere within the basin. These results strongly suggest that the source of the Ti and Th enrichments at SPA are not due to mare basalt. Another possibility is that the floor consists of lower crustal impact melts that are similar in origin to LKFM found on the nearside but having lower TiO2 and Th contents. This is possible in principle: LKFM is quite variable in its chemical characteristics (Table 1) and the lower crust of the Moon on the farside may not be similar to that sampled by the basins which gave rise to LKFM on the nearside [Lucey et al., 1994]. Thus our new compositional data are most consistent with the presence of substantial mantle material, or an unusual type of lower crustal material.

5. Models of Impact Basin Formation and Composition of SPA Basin

Our measurements of FeO and TiO2 of the SPA basin allow us to test models for basin excavation. The mechanics of the gigantic basin-forming events must be pieced together from the indirect evidence provided by terrestrial impact structures [e.g., Grieve et al., 1981], theoretical understanding of the impact process [e.g., O'Keefe and Ahrens, 1994], and extrapolation from much smaller experimental craters [e.g., Holsapple, 1993].

The observed basin diameter of 2500 km [Spudis et al., 1994] suggests that the diameter of the transient crater was of the order of 1400 km [Grieve et al., 1981; Spudis, 1993], assuming a near-normal incidence hypervelocity impact. Such a crater would excavate material from as deep as 150 km within the Moon and would have a total excavated volume of about 1011 x 106 km3 of material. Assuming an average crustal thickness of 100 km (a value common on the lunar farside [Zuber et al., 1994]), about 95 x 106 km3 would be derived from the crust (95% of all ejecta) while almost 8 x 106 km3 (5% of ejecta) would come from the mantle of the Moon. Thus, even though the SPA basin is large enough to have completely excavated the crust from its target site, according to the proportional growth model [Grieve et al., 1981; Spudis, 1993], the bulk of ejecta would come from the crust of the Moon (Figure 3).

According to current cratering models [Cintala and Grieve, 1994], the provenance of the impact melt of this large-scale event is entirely different than that of the ejecta. Scaling relations [Holsapple, 1993; Cintala and Grieve, 1994] suggest that SPA basin must have been formed by a projectile at least a few hundred kilometers in diameter. We have modeled the impact of a 250-km diameter projectile impacting the Moon at 20 km s^-1. Such an impact would generate about 180 x 106 km3 of melt, almost twice the estimated volume of material excavated from the cavity [Cintala and Grieve, 1994]. Moreover, this impact melt was generated largely in the upper mantle; crustal rocks make up less than a few percent of the melt zone. A large fraction of this melt would be ejected from the basin cavity, but more than half would remain in and below the basin floor. Using a somewhat different approach, Warren et al. [1996] also conclude that the SPA impact melt sheet would consist virtually entirely of mantle rock.

Recently, Schultz [1997] suggested that the size, geometric asymmetries, and geological relations of SPA basin indicate that its formation must have been unusual. Schultz [1997] postulates the creation of SPA basin by a very low velocity, oblique impactor. Such an impact would have several consequences, including nonproportional cavity growth and crater excavation from extremely shallow depths [Schultz, 1997]. Moreover, the near-grazing collision of a large, slow projectile might entail the capture of a ring of debris in Earth-Moon space, a source of late-heavy-bombardment cataclysm debris [Ryder, 1990]. Schultz [1997] contends that any "normal" impact the size of SPA should have shattered the Moon completely according to normal crater scaling and is troubled by the fact that the "collision did not initiate significant mantle uplift" [Schultz, 1997, p. 1260].

However, Zuber et al. [1994] showed that the floor of SPA basin does show anomalously thin crust (20 km) compared to the average farside crustal thickness (68 km). The global gravity map shows a negative free-air anomaly, but of much lower magnitude than could be reasonably expected from the astonishingly high apparent crater volume (23 million km3 [Spudis and Aitkins, 1996]). This suggests that the basin must be partly compensated, as shown in the Bouger anomaly map of Zuber et al. [1994]. These data indicate a large positive anomaly. As the associated crustal thickness map shows that the depth to the mantle is much less here than elsewhere and the surface relief is much less than the amount of thinning, the mantle must be uplifted.

Finally, we note that before SPA basin had been described, it had been suggested that Imbrium basin, less than half the diameter of SPA, was almost large enough to shatter the Moon [Gault, 1974]. Thus, unlike some of the assumptions of Schultz [1997] SPA basin does display many of the attributes of allegedly "normal" basins, including a plug of uplifted mantle and a high apparent crater volume.

6. Conclusions

The geochemical characteristics of the interior of South Pole-Aitken basin are consistent with a mixture of mantle and lower crustal material. The data also permit the presence of a pure, but unusual lower crustal composition. Either possibility requires rethinking of standard models for large basin formation. However, the latter possibility requires that the floor of SPA basin contains virtually no mantle material, suggesting that existing models of impact melt generation are entirely wrong or that the
assumptions of a hypervelocity, nonoblique impact are incorrect. Considering the enormous size of the basin, we believe that SPA would have excavated into the mantle. We conclude that the floor of SPA basin consists of a mixture of lower crust composed of LKF mafic rock and an upper mantle which is constrained to contain 10-20 wt % FeO and ~0.1 wt % TiO₂. If true, our conclusion does not necessarily call into question current models for large basin formation, nor the assumptions of a nonoblique hypervelocity impact. The modification stage of formation of SPA basin must have been an extremely violent event with ample opportunity for mixing of melt, fallback ejecta, and brecciated basin floor material.

Full coverage of the basin with Lunar Prospector data may greatly improve our confidence in these results, but our analysis indicates that use of FeO, TiO₂, and Th alone probably cannot resolve this issue. Continued examination of Clementine data for the presence of olivine, with special attention paid to improving the detection limit, will aid in distinguishing between these possibilities. It is likely, however, that the ultimate resolution to the problem will require a sample from the interior of South Pole-Aitken basin. Regardless of the outcome of the remote sensing studies, South Pole-Aitken remains one of the largest and, owing to the significance of the mantle or lower crustal rock types which compose its floor, doubtless most important geochemical provinces on the Moon.

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