Origin of strong lunar magnetic anomalies: Further mapping and examinations of LROC imagery in regions antipodal to young large impact basins

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The existence of magnetization signatures and landform modification antipodal to young lunar impact basins is investigated further by (a) producing more detailed regional crustal magnetic field maps at low altitudes using Lunar Prospector magnetometer data; and (b) examining Lunar Reconnaissance Orbiter Wide Angle Camera imagery. Of the eight youngest lunar basins, five are found to have concentrations of relatively strong magnetic anomalies centered within 10° of their antipodes. This includes the polar Schrödinger basin, which is one of the three youngest basins and has not previously been investigated in this context. Unusual terrain is also extensively present near the antipodes of the two largest basins (Orientale and Imbrium) while less pronounced manifestations of this terrain may be present near the antipodes of Serenitatis and Schrödinger. The area near the Imbrium antipode is characterized by enhanced surface thorium abundances, which may be a consequence of antipodal deposition of ejecta from Imbrium. The remaining three basins either have antipodal regions that have been heavily modified by later events (Hertzsprung and Bailly) or are not clearly recognized to be a true basin (Sikorsky-Rittenhouse). The most probable source of the Descartes anomaly, which is the strongest isolated magnetic anomaly, is the hilly and furrowed Descartes terrain near the Apollo 16 landing site, which has been inferred to consist of basin ejecta, probably from Imbrium according to one recent sample study. A model for the origin of both the modified landforms and the magnetization signatures near lunar basin antipodes involving shock effects of converging ejecta impacts is discussed.


1. Introduction

Previous work has found evidence that the largest concentrations of strong lunar crustal magnetic fields are in regions antipodal (diametrically opposite) to four young large impact basins: Orientale, Imbrium, Serenitatis, and Crisium [Hood et al., 1979; Hood, 1981; Lin et al., 1988; Richmond and Hood, 2008; Mitchell et al., 2008; Purucker and Nicholas, 2010; Tsunakawa et al., 2010]. In at least two cases (Orientale and Imbrium), unusual “hilly and furrowed” or “grooved and mounded” terrain is found on imagery and geologic maps in the same regions. This terrain has been interpreted as an antipodal consequence of the associated impacts, i.e., shocking by converging ejecta impacts [Moore et al., 1974; Haskin, 1998; Hood and Artemieva, 2008] or converging seismic waves [Schultz and Gault, 1975; Watts et al., 1991]. Models have also been proposed for the production of the antipodal magnetization signatures involving shock remanent magnetization (SRM) of impacting basin ejecta in the presence of a transient magnetic field amplified by the expanding ionized vapor-melt cloud [Hood, 1987; Hood and Huang, 1991; Hood and Artemieva, 2008]. Current evidence indicates that the initial (unamplified) magnetic field during the lunar basin-forming epoch was produced by a core dynamo [Garrick-Bethell et al., 2009; Hood, 2011; Shea et al., 2012; Tikoo et al., 2012; Weiss et al., 2012].

On the other hand, global maps of the distribution of lunar crustal fields also show that the largest group of strong anomalies occurs on the central far side along the northeastern topographic rim of the South Pole-Aitken (SPA) basin, which is the oldest and largest recognizable lunar basin [Purucker et al., 2006; Richmond and Hood, 2008; Mitchell et al., 2008; Purucker and Nicholas, 2010; Tsunakawa et al., 2010]. The location of this group of anomalies near one
edge of the SPA has led to an alternate set of hypotheses for the origin of strong lunar magnetic anomalies. According to one model, they result from magnetized subsurface dike swarms that fed mare basalt patches emplaced within the SPA basin rim [Purucker et al., 2012]. This hypothesis was stimulated in part by the occurrence of west-northwest trending linear magnetic features on both radial component and scalar intensity maps of the region produced using the methods of Purucker and Nicholas [2010]. According to another recent model, these strong anomalies, as well as other more isolated anomalies around the Moon, are a consequence of deposition of iron-rich ejecta from the SPA impactor that is assumed to have impacted the Moon obliquely from the south [Wieczorek et al., 2012]. The latter authors show, on the basis of numerical impact simulations that such ejecta could have been distributed preferentially around the northern edge of the basin where the magnetic anomalies are concentrated.

[4] In this paper, we investigate further the existence of magnetization signatures and landform modification antipodal to young lunar basins by (a) producing more detailed regional crustal magnetic field maps at low altitudes using Lunar Prospector (LP) magnetometer data; and (b) examining Lunar Reconnaissance Orbiter Wide Angle Camera (LROC WAC) imagery. In addition to the four basins identified above, we also consider several other young basins, including the polar Schrödinger basin, which is one of the three youngest lunar basins and which has not previously been considered in this context. In section 2, we first focus on the large group of magnetic anomalies on the south-central far side near and within the SPA basin that has been the subject of several recent studies. The method of field mapping is first summarized and applied to produce a low-altitude regional map of the scalar field magnitude in this region. This map is then compared with topographic maps, partial geologic maps, surface imagery, and compositional (thorium) data to allow a discussion of the various proposed hypotheses for explaining the strong anomalies in this area (antipodal effects of young basins, subsurface dike swarms, and deposition of iron-rich ejecta from the SPA impactor). Although some evidence favoring antipodal effects of the Imbrium and Serenitatis impacts is obtained, the evidence based on this region alone is not conclusive. In sections 3 and 4, we therefore examine the antipodal regions of a series of other young lunar basins to test further the antipodal magnetization hypothesis and to search for evidence of unusual terrain similar to that found near the Imbrium antipode. In section 3, the regions antipodal to Orientale and Crisium are investigated as well as an area on the lunar near side where the strongest isolated anomaly (the Descartes anomaly) is found. In section 4, a map is produced of the scalar field intensity in the northern polar region, including the area near the antipode of Schrödinger. Available imagery in this region is also examined to search for evidence of unusual terrain. Further discussion and conclusions are given in section 5.

2. Magnetic Anomalies on the South-Central Lunar Far Side

[5] In this section, we consider a large area of the south-central far side extending from 140°E to 210°E and from 50°S to 10°S. This area includes part of the SPA basin and adjacent highland terrain and contains the largest group of strong anomalies on the Moon [see, e.g., Richmond and Hood, 2008, Figure 7a; Mitchell et al., 2008, Figure 4; Purucker and Nicholas, 2010, Figure 6; or Tsunakawa et al., 2010, Figure 12]. Most of the mapped anomalies can be interpreted as concentrated near the antipodes of the Imbrium and Serenitatis basins, but they can also be described as concentrated along the northwest topographic rim of the SPA basin.

2.1. Mapping Methods

[6] Because of the relatively weak lunar crustal fields (compared to terrestrial fields) and the presence of external field noise resulting from the solar wind interaction, construction of lunar crustal field maps is not a straightforward task. Several alternate methods for mapping these fields have therefore been developed. First, a direct method has been applied, involving careful selection, editing, and filtering of measurements along a series of orbit tracks, followed by some form of altitude normalization, interpolation, and/or two-dimensional filtering [Hood et al., 1981; Richmond and Hood, 2008; Tsunakawa et al., 2010; Hood, 2011; Hemingway and Garrick-Bethell, 2012]. The second involves the construction of global spherical harmonic models of the crustal field using an extension of techniques that have been applied to map the terrestrial crustal field [Purucker, 2008; Purucker and Nicholas, 2010].

[7] Both mapping methods (direct and spherical harmonic) can be successful in minimizing external field contributions and both methods have advantages and disadvantages. The spherical harmonic approach can be better for producing a global map with homogeneous properties, especially if the effective coverage of relatively noise-free measurements is nearly complete. Also, the resulting vector crustal field B is guaranteed to be a potential field (\(\nabla \cdot \mathbf{B} = 0\)) and it is possible to compute the crustal field model at any altitude from the fitted harmonic coefficients. However, at least in the case of the LP mission, global coverage of high-quality measurements is not entirely complete, which leads to non-negligible errors in some regions and at small spatial scales. The latter errors can be important when attempting to correlate orbital anomalies with surface geology or when modeling anomaly field components to estimate directions of magnetization. For the purpose of constructing regional maps in an area where good quality, low-altitude measurements are available, the direct method can be better, because the resulting map is not affected by lower-quality measurements outside the region of interest and only the best measurements (usually from a single lunation) can be used to construct a given map. For this reason, the relative amplitudes of individual anomalies can usually be accurately determined even if the mapped field is not exactly divergenceless.

[8] Since regional maps are sufficient for the present work, we apply here the simplified direct method described by Hood [2011]. Unlike the approach of Richmond and Hood [2008], which sought to maximize global coverage, the measurements used here have been selected for highest available quality over the region of interest at the lowest possible altitudes. Although some measurements were obtained at altitudes as low as 15–20 km, the horizontal resolution of the resulting field maps is usually limited by the separation...
between consecutive orbit tracks, which is near 30 km (\(\sim 1^\circ\)) at the equator. As in previous work, periods that are heavily affected by short-wavelength external field noise are identified by the absence of a good cross-correlation between anomalies measured on adjacent orbit tracks. These periods are eliminated either by deleting the entire orbit or by partial editing. External fields with longer wavelengths (comparable to the lunar radius) in remaining orbit segments are minimized by fitting and removing a second-order polynomial from the data (quadratic detrending). If gaps exceeding 30 km between adjacent remaining orbits are present, the gap is filled by adding one or more orbit segments at a comparable altitude from a different lunation or, as a last resort, by linear interpolation between the remaining orbit tracks.

In order to correct approximately for the dependence on altitude above the mean lunar radius \(z\) of lunar magnetic anomaly amplitudes, we assume a power law dependence \(B \propto z^{\alpha}\), where \(\alpha\) is estimated empirically using data at a series of altitudes above a large number of anomalies. A value of \(\alpha \approx 2.5\) has been found to be approximately valid over a range of altitudes less than 50 km [Richmond and Hood, 2008]. A large majority of the measurements selected here are obtained at altitudes less than 30 km, and the field maps are normalized to a constant altitude of 25 km. Most measurements are obtained at altitudes less than 25 km so the resulting upward continuation has the effect of suppressing short-wavelength noise.

Finally, the selected, edited, detrended, and altitude-normalized field measurements are two-dimensionally filtered to linearly interpolate between orbit segments and produce an equally spaced grid suitable for contour mapping. This step reduces the horizontal resolution of the final map to about 2\(^\circ\) of latitude or longitude (\(\sim 60\) km at the equator). However, the relative amplitudes of anomalies are not strongly affected so the resulting map is suitable for comparative studies with surface geology.

To illustrate the method of analysis, Table 1 lists 20 orbit segments selected from a period in June 1999 when the LP spacecraft altitude was relatively low and external field fluctuations were minimal. For all selected orbit segments, equatorial crossing longitudes are located on the central far side and only measurements obtained at latitudes between 60°S and 60°N are considered. Figure 1 plots the three vector field components (B-Radial, B-East, and B-North) plus the field magnitude (B-Total) for these orbit segments after editing to eliminate periods of short-wavelength external field fluctuations and after quadratic detrending. Especially broad regions of anomalies are evident on the south-central far side near the top of the figure (longitudes of 190°E to 195°E) and near the bottom of the figure (longitudes of 175°E to 180°E).

A careful examination of Figure 1 shows that the radial component of the crustal field is least affected by external field variations. This is partly because this component is typically somewhat stronger than the tangential components for an isolated anomaly, but it is mainly because external field variations normal to the surface are suppressed by opposing induced currents in the electrically conducting subsurface (Lenz’s Law) while field variations tangent to the surface are amplified. The tendency for the normal field component to be continuous across the surface of the conducting interior follows from the integral equivalents of Maxwell’s equations [e.g., Jackson, 1975]. Empirically, it is known that this is true near the Moon based on direct comparisons of vector field measurements far from the Moon with those obtained simultaneously at or near the surface [see, e.g., Hood et al., 1982, Figure 4].

Figure 2 shows a contour map of the two-dimensionally filtered scalar field magnitude across the study area. The contour map is superposed upon a color-coded shaded relief map of Clementine LIDAR topography (U.S.G.S.; http://www.mapaplanet.org). To produce the map, the magnetometer data (all from ascending node passes during June of 1999) were first sorted into 0.5° bins and were then two-dimensionally smoothed twice using a 5 bin \(\times\) 5 bin moving boxcar filter. This filtering method effectively interpolates across minor gaps between passes and also smooths the peak anomaly amplitudes such that the maximum amplitudes shown on the map are somewhat less than the maximum amplitudes observed on some orbit passes. For example, the largest single anomaly shown in Figure 2 has a peak amplitude at 25 km of \(\sim 12.5\) nT and is located at approximately 35°S, 163°E. An examination of the corresponding detrended data segments (similar to those shown in Figure 1) shows that the peak amplitude of this anomaly on one orbit pass was about 30 nT at an altitude of about 19 km, which yields an estimated amplitude at 25 km of \(30 \times (19/25)^{2.5} \approx 15\) nT. Thus, the smoothed peak amplitude is about 17% lower than the true (altitude-adjusted) peak amplitude on one orbit pass for this anomaly.

### Table 1. Selected LP 1999 MAG Data Intervals for Figure 1

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<tr>
<th>Month</th>
<th>Start Time (day/h)</th>
<th>End Time (day/h)</th>
<th>Longitude(^{a}) (°E)</th>
<th>Altitude(^{a}) (km)</th>
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\(^{a}\)Value at equator crossing.
Figure 1. Field magnitude and vector field components (nT) during a selected series of orbit passes from June of 1999 (see Table 1 for details). Contributions of external magnetic fields to the data were minimized by editing and quadratic detrending (see the text).

On the other hand, some anomalies occur within the basin rim (blue) while areas of higher elevation (red) have weak anomalies.

Figure 3 shows a superposition of the same field magnitude contour map onto a partial geologic map of the region, modified from Hood and Williams [1989] and based on the original geologic map of Stuart-Alexander [1978]. (Note that U.S.G.S. lunar geologic maps can be obtained at http://www.lpi.usra.edu/resources/mapcatalog/usgs/.) As seen on this figure, the strongest anomalies also tend to occur in a region around the pre-Nectarian Ingenii basin near the Imbrium antipode (cross). This region is distinguished by the occurrence of unusual terrain, “material of grooves and mounds, Imbrian in age” (blue unit). As noted by Stuart-Alexander, this unit (hereafter referred to as GM terrain) is most probably a consequence of antipodal effects of the Imbrium basin-forming event (convergence of ejecta and/or seismic waves). Its existence was first reported during detailed examinations of Lunar Orbiter and Apollo metric camera photography [e.g., Schultz, 1972].

Figure 4a is a composite of Wide Angle Camera (WAC) images of the Imbrium antipode region (cross) obtained with the Lunar Reconnaissance Orbiter Camera.
Figure 2. Two-dimensionally filtered field magnitude (nT) at 25 km altitude on the south-central lunar far side, constructed using Lunar Prospector orbit passes from June of 1999. The magnetic field contour map (contour interval, 1 nT) is superposed on color-coded Clementine topography overlaid on a U.S.G.S. shaded relief map of the Moon (Mercator projection; http://www.mapaplanet.org).

(LROC) [Robinson et al., 2010; http://www.lroc.asu.edu]. Within Mare Ingenii, swirl-like albedo markings are visible on the mare surface, marking the location of the strongest single anomaly in Figures 2 and 3. Recent analyses of both Clementine and LRO data [Neish et al., 2011; Kramer et al., 2011a, b] have supported a model for the origin of the swirls involving magnetic deflection of the solar wind ion bombardment, which contributes to the darkening with time of the lunar surface [Hood and Schubert, 1980; Hood and Williams, 1989; Blewett et al., 2011]. The swirls therefore represent a secondary consequence of the existence of strong anomalies rather than source materials for the anomalies. The locations of prominent swirls visible on imagery at high sun illumination angles are also indicated on Figure 3 with small black markings.

Exposures of grooved and mounded terrain can be seen in Figures 4a and 4b, respectively. Grooved terrain is especially pronounced in the walls of craters and basins in the general area around the Imbrium antipode. The most extreme example is found lining the southern wall of the Ingenii basin (arrow in Figure 4a). However, it is also found lining the rest of the basin wall as well as within a number of craters surrounding Ingenii (see also Figure 16b below). Grooved terrain in basin and crater walls is usually oriented roughly perpendicular to the basin or crater rim; this is in contrast to normal crater terraces, which are oriented mainly parallel to crater rims. Figure 4b is a higher-resolution view of the area southeast of Birkeland (white rectangle at lower right in Figure 4a), which lies within the material of grooves and mounds (blue unit in Figure 3). This unit is characterized by undulating topography on a scale of 2 to 10 km (solid arrows) and the presence of lineations with widths of order 1 km and lengths of 5–15 km (dashed arrow). It has a surface texture that is distinctly rougher than that of normal pre-Nectarian highland terrain or younger crater ejecta mantles, which is much smoother and has few, if any, lineated features. To illustrate the difference, Figure 4b includes part of the Birkeland ejecta mantle, which is noticeably smoother than the preexisting GM terrain.

There are several reasons why the GM terrain around Mare Ingenii is probably an antipodal consequence of the Imbrium impact. First, as discussed further in section 3 below, morphologically related terrain is also known to exist near the Orientale antipode [Wilhelms and El-Baz, 1977], which is the youngest large basin on the Moon. Second, comparable terrain is also found near the antipode of the ~1550 km diameter Caloris basin on Mercury [see, e.g., Murray et al., 1974; Denevi and Robinson, 2008; Blewett et al., 2010, Figure 4] and opposite to the 9 km diameter crater Stickney on Phobos [Fujiiwara, 1991]. There is therefore little doubt that antipodal terrain modification can occur following a massive basin-forming impact.

As noted in section 1, possible mechanisms for producing modified antipodal terrain include convergence of ejecta and convergence of seismic waves. Although two-dimensional numerical simulations suggested that seismic
production of disrupted antipodal terrain is possible on the Moon [Watts et al., 1991], later, three-dimensional simulations yielded much weaker evidence for antipodal focusing of seismic energy, at least for bodies with icy mantles [Bruesch and Asphaug, 2004]. Because of a thick layer of unconsolidated material in the upper crust (the “megaregolith”), surface waves attenuate rapidly with distance on the Moon [Oberst and Nakamura, 1992] so that any antipodal effects would need to be produced mainly by reflected and focused compressional body waves. The latter would arrive at the antipode in less than an hour after the impact. It is possible, for example, that the grooved terrain in the walls of Ingenii and other craters near the Imbrium antipode was produced by massive landslides and disruption induced by the seismic mechanism [Schultz and Gault, 1975].

Numerical simulations indicate that antipodal ejecta deposition occurs beginning several hours after the impact, especially for oblique impact angles [see, e.g., Hood and Artemieva, 2008, Figures 6 and 7]. Possible observational evidence for antipodal ejecta deposits from a relatively young large crater (Tycho, 86 km in diameter) has recently been reported in the form of unusual young impact melts and rocky exposures [Robinson et al., 2011; Bandfield et al., 2013]. As discussed in Hood and Artemieva [2008], numerical simulations indicate that most antipodal ejecta from a basin-scale event consist of impact melt originating from the lower crust or upper mantle. Individual molten ejecta parcels larger than a few meters in size and ejected at angles less than about 35° will remain mostly molten (with a solidified outer shell) until they impact and disintegrate in the antipodal region [Artemieva, 2013].

The remainder of the ejecta will consist of smaller particles, solidified during the flight from the primary impact location. Extensive secondary impact cratering in the antipodal region is therefore not expected. Rather, a mixture of impact melt and rocky material will be deposited. It is possible that the mounded and lineated terrain outside of the Ingenii basin (e.g., Figure 4b) represents the remnants of this material. At the time of impact at speeds of several kilometers per second, shocking of the material to peak pressures of ~ 20 GPa will occur for an Imbrium-scale event. Such shocks are sufficient to produce strong and stable SRM [Cisowski et al., 1973; Gattacceca et al., 2010] in a transient magnetic field. Some of the deposited impact melt may also have acquired thermoremanent magnetization (TRM), if it cooled slowly enough in the steady lunar dynamo field.

Previous statistical studies have found that the GM terrain near the Imbrium antipode is at least moderately correlated with crustal magnetic fields as measured using both Apollo subsatellite and LP magnetometer and/or electron reflectometer (ER) data [Hood and Williams, 1989; Richmond et al., 2005]. Specifically, normalized occurrence
Figure 4. (a) Region near the Imbrium antipode (cross) on the south-central lunar far side. “Grooved” terrain is present in the walls of the pre-Nectarian Ingenii basin (arrow). (b) Higher-resolution view of “material of grooves and mounds” (white rectangle at lower right in Figure 14a) as identified on the geologic map of Stuart-Alexander [1978]. Images were adapted from LROC WAC Global Morphological Map product. See the text for further explanation.

Figure 5. Possible modified terrain near the Serenitatis antipode (cross). The visible extent of this terrain is estimated in Figure 3 near 20°S, 160°W (blue unit). (LROC WAC).

rates of strong magnetic fields mapped using Apollo ER data and LP magnetometer data are larger, on average, over the GM terrain than over other geologic units in this region. However, as can be seen in Figure 3, the correlation is not high, with some anomalies occurring over ordinary pre-Nectarian highland terrain and some areas of the GM terrain having relatively weak magnetic fields. One reason for the reduced correlation may be that parts of the GM terrain have been overlain by later mare basalt flooding episodes and by ejecta from younger impacts. For example, the strongest single anomaly in the region (smoothed amplitude 12 nT at 25 km altitude) is located within Mare Ingenii, which is covered with basalt (red unit). Also, ejecta from the younger crater Birkeland northeast of Ingenii covers a substantial part of the GM terrain in an area of strong magnetic fields.

As shown in Figure 5 (also an LROC WAC composite), unusual hilly and lineated terrain that could be related to the GM terrain near the Imbrium antipode is present within 5 to 10° of latitude of the Serenitatis antipode (cross). This terrain is not shown on the map of Stuart-Alexander [1978], but its existence was communicated to one of us (LLH) by D. Wilhelms in 1978. The resemblance of this terrain to the GM terrain was noted also in the paper by Hood and Williams [1989] who included a Lunar Orbiter photograph of the same area. The area occupied by this terrain is therefore added to the partial geologic map in Figure 3 (blue areas in upper right part of the figure) even though it is not present on standard U.S.G.S. maps. Its proximity to the geometric antipode and its morphological similarity to the GM terrain favor a genetic relationship to the Serenitatis impact. However, alternate explanations cannot be ruled out. For example, Wilhelms [1987] notes that this terrain could alternatively be interpreted as associated with secondary cratering from nearby basins such as Korolev and Apollo.

Figure 6 is a superposition of a contour map of LP Gamma Ray Spectrometer thorium data [Lawrence et al., 2000] onto the same partial geologic map used in Figure 3. Specifically, absolute thorium (Th) abundances (in units of microgram per gram) derived from data obtained at low altitudes (Release 2) were obtained from http://www.lunarpenalty\@M.lanl.gov/pages/GRSthorium.html. These data have 2° × 2° resolution at the equator, which is comparable to that of the 2D filtered magnetometer data shown in Figure 3.

As discussed by Lawrence et al. [2000], the Th distribution within SPA appears to have two components. First, there is a low-abundance, diffuse component that correlates negatively with topography and is centered near the center of SPA. The latter probably reflects an increasing concentration of Th and mafic elements with depth as expected for a crust that formed from a large-scale differentiation event. Second, there is a higher-abundance component that is present mainly in the northwest sector of SPA and probably has a different origin from that of the diffuse component. As seen in the figure, the higher-abundance Th component is distributed not far from the Imbrium antipode. One early
model for the origin of the latter component was suggested by Hawke and Spudis [1980]. They proposed that cryptomare (hidden mare) deposits that are enhanced in Th may have been extruded from depth near the Imbrium antipode via crustal fissures caused by converging seismic waves. However, Lawrence et al. [2000] have argued that this explanation is unlikely because other (visible) mare surfaces in the region are not enhanced in Th.

As seen in Figure 6, the enhanced Th abundances also correlate moderately with the occurrence of the GM terrain, and the correlation is somewhat better than for the crustal magnetic field in Figure 3. The possible association of the enhanced Th with this terrain has been discussed by a number of authors, including Hawke and Spudis [1980] who argued that a genetic connection is unlikely because the GM terrain is absent where the largest concentration of Th known at that time is found (centered near the crater Birkeland). However, a genetic connection was considered more favorably by L. Haskin [e.g., Haskin, 1998] and was discussed in more detail by Wieczorek and Zuber [2001].

The latter authors note that the two Th maxima (centered at about 35°S, 175°E and at 41°S, 166°E) are associated with two young Eratosthenian-aged craters (Birkeland and an unnamed crater), suggesting that these craters excavated thorium-rich upper crust and redistributed it at the surface. The same ejecta mantles would have covered the GM terrain, partially explaining its absence at those locations. One of these high-thorium regions lies within the GM terrain while the other lies just outside of it. Wieczorek and Zuber therefore consider it to be possible that the GM unit is the main source of the Th enhancement in this part of SPA. A comparison of Figure 6 with Figure 2 also suggests that the two Th maxima (4.5 µg/g) may be related to the two main magnetic field maxima over the main GM terrain unit. Although the two Th maxima are displaced to the south of the magnetic field maxima, this could be explained by excavation and exposure of Th-rich material by the two Eratosthenian-aged craters mentioned above.

If part or all of the GM terrain is a consequence of deposition and shocking of converging Imbrium ejecta, then the enhanced Th shown in Figure 6 could have originated in the lower crust and upper mantle beneath the Imbrium basin [Haskin, 1998]. The Imbrium impact event is known to have excavated material enriched in incompatible and heat-producing elements (i.e., KREEP) from the Procellarum-KREEP Terrane (PKT) [e.g., Jolliff et al., 2000]. Numerical impact simulations [e.g., Hood and Artemieva, 2008] show that all distant ejecta for an Imbrium-scale event come from the crust and a substantial part of the antipodal ejecta comes from the lower crust, which could have been enriched in KREEP beneath Imbrium.

Finally, it should be noted that Wieczorek and Zuber [2001] discounted an Imbrium origin for the GM terrain in favor of a Serenitatis origin on the basis of analytic ballistic ejecta trajectory calculations for a rotating Moon. Using vertical impact-scaling relationships and neglecting any interaction of the ejected material, they found that the distribution of antipodal ejecta deposits should have a crescent-like shape [see Wieczorek and Zuber, 2001, Figure 3 and Plate 3], unlike the irregular distribution that is observed for the GM terrain. For the most probable values of the adopted impact-scaling constants, large westward offsets of the antipodal deposits from the geometric antipode were obtained, ranging from ~20° for the present-day lunar rotation rate to ~30° for a rotation rate of 12 days (equivalent to an Earth-Moon distance of 35 Earth radii, assuming synchronous rotation). The large offsets were due to relatively long flight times (>24 h) of material ejected at large angles (approaching 45°) during the early excavation stage.

However, later explicit three-dimensional simulations for a vertical Imbrium-scale lunar impact also yielded flight times to the antipode of >24 h but found that only a negligible amount of material was ejected at angles that would lead to antipodal ejecta deposition [see Hood and Artemieva, 2008, Figure 6a]. Since it is unlikely that a lunar basin-forming impact was nearly vertical, explicit simulations were also carried out for an oblique impact (angle from the horizontal of 30°). For this case, it was found that substantial antipodal ejecta deposition does occur and initial downrange ejection angles were relatively low (<35°) so that most ejecta arrive at the antipode within 4 to 8 h [see Hood and Artemieva, 2008, Figure 6b]. For plausible early lunar rotation rates, the westward displacement was therefore estimated to be no more than 5° to 10°, which is much less than the angular extent of the GM terrain shown in Figures 3 and 6. These numerical results, together with the possible presence of a separate occurrence of unusual GM-like terrain near the Serenitatis antipode (Figure 5), is more consistent with an Imbrium origin for the GM terrain, whether or not this terrain is the source of the enhanced Th anomalies.

2.3. Interpretation of the South-Central Farside Anomalies

The comparisons of the distribution of crustal magnetic fields to surface geology and Th abundance on the south-central far side given in the previous subsection (Figures 2–6) strongly suggest but do not prove that these
anomalies are a consequence of antipodal effects of the Imbrium and Serenitatis impacts. Although it might be more probable that the GM terrain around the Ingenii basin is a result of deposition of Imbrium-produced melt and rocky material rather than modification of the pre-existing surface by seismic waves, the correlation of this terrain with magnetic field amplitudes shown in Figure 3 is only moderate. Without more information, it is therefore not possible to conclude that the GM terrain is the main source material for the observed magnetic anomalies in and around the Ingenii basin. Also, as discussed above, it is not certain that the unusual terrain north of the Serenitatis geometric antipode shown in Figure 5 is genetically related to the Serenitatis impact. Finally, while the enhanced Th concentrations in the northwest part of SPA appear to correlate somewhat better with the GM terrain than do the magnetic fields and can be plausibly interpreted as a consequence of antipodal ejecta deposition from the Imbrium event [Haskin, 1998], alternate explanations for these enhanced concentrations [e.g., Hawke and Spudis, 1980] cannot be completely ruled out.

[30] Alternate models for the sources of the central far-side magnetic anomalies [Purucker et al., 2012; Wieczorek et al., 2012] are difficult to eliminate on the basis of data for this region alone. Although the WNW trending magnetic lineations found on the maps of Purucker and Nicholas [2010] are not replicated on independent maps (e.g., Figure 2), it can be argued that the advanced processing techniques (sequential dipole and coestimation approaches) applied in their analysis are able to extract weaker signals than can be detected using simpler direct mapping methods. As reviewed by Purucker et al. [2012], previous work has shown that lava ponds are especially voluminous and numerous in the northwest part of SPA [Yingst and Head, 1999]. The largest single lava pond is within the Ingenii basin where the strongest single magnetic anomaly is found (Figure 3). Feeder dikes (sheet-like magmatic intrusions) undoubtedly existed during the period when the lava ponds formed as they did elsewhere on the Moon during the formation of the lunar maria [e.g., Head and Wilson, 1992]. It is possible that the dikes near the SPA basin solidified when a strong core dynamo field existed, while dikes that led to the formation of the nearside maria solidified at a later stage when no significant magnetization would have been acquired. In view of the known importance of volcanic sources of terrestrial magnetic anomalies, which also formed in the presence of a long-lived dynamo magnetic field, a dike-related origin for the anomalies on the south-central far side is difficult to eliminate a priori.

[31] One possible problem for the subsurface dike source hypothesis for the origin of the magnetic anomalies in this region is that high-resolution gravity data from the Gravity Recovery and Interior Laboratory (GRAIL) mission have so far not confirmed the presence of subsurface dikes in the areas where lineated magnetic anomalies are found. Some dike-like structures have been identified in the GRAIL data and are believed to have formed by magmatism in the presence of an expanding lithosphere during pre-Nectarian and Nectarian times [Andrews-Hanna et al., 2013]. However, the ancient dikes identified in GRAIL data are found in a variety of locations around the Moon whereas the largest group of magnetic anomalies is on the south-central far side. Also, while a few dikes are identified near the SPA basin, these are found in a different area (120°E to 150°E) than where the lineated magnetic features are found (160°E to 210°E). Finally, the GRAIL-identified dikes near SPA have a northwest orientation rather than a west-northwest orientation.

[32] Another possible problem for the subsurface dike source hypothesis is that prior work has strongly suggested that basin ejecta materials are the most probable sources of lunar orbital magnetic anomalies [Strangway et al., 1973b; Hood et al., 1979; 2001; Hood, 1981; Halekas et al., 2001]. The dominant ferromagnetic carriers in lunar samples are microscopic Fe-Ni alloy (kamacite) particles that are most commonly present in impact-produced materials such as breccias and are much less abundant (by a factor of roughly 4) in volcanic materials such as mare basalt [e.g., Strangway et al., 1973a; Fuller and Cisowski, 1987]. This contrasts with the terrestrial (and martian) case where more oxidizing conditions led to mainly iron oxide remanence carriers (e.g., magnetite) that tend to be prevalent in volcanic materials. Consistently, of the four Apollo missions that carried surface magnetometers, the two that landed in highland areas (Apollo 14 on the Fra Mauro Formation and Apollo 16 on the Cayley Plains near the Descartes Formation) found stronger fields (43 and 103 nT at Apollo 14 and up to 327 nT at Apollo 16) [Dyal et al., 1974]. These early results led to the original suggestion of Strangway et al. [1973b] that basin ejecta materials such as the Cayley Formation are likely sources of magnetic anomalies detected from orbit.

[33] Early studies of Apollo 15 and 16 subsatellite magnetometer data [Coleman et al., 1972; Russell et al., 1975] found anomalies over areas dominated by the Fra Mauro and Cayley Formations [e.g., Hood et al., 1979]. The Fra Mauro Formation is a knobby unit distributed mainly around the Imbrium basin and is interpreted as primary Imbrium ejecta [Head and Hawke, 1975]. The Cayley Formation is a flat light plains unit that overlies the Fra Mauro; it may represent fluidized ejecta from Imbrium (and possibly Orientale) and local material pulverized and scattered by secondary basin ejecta impacts [e.g., Wilhelms, 1987]. A later statistical analysis of LP ER data over the near side also found that the Cayley Formation correlates well with crustal field intensity [Halekas et al., 2001]. Finally, the LP ER and magnetometer data showed that the strongest single anomaly on the lunar near side occurs over a high-albedo area of the Descartes Formation near the Apollo 16 landing site where the strongest surface fields were measured [Halekas et al., 2001; Richmond et al., 2003]. As discussed further in section 3.2 below, the Descartes Formation is accepted by most workers to consist of impact basin ejecta.

[34] Although one magnetic anomaly associated with a nearside extensional graben, the Rima Sirsalis rille, was reported using Apollo subsatellite ER data [Anderson et al., 1977] and a model for its origin involving a magnetized dike swarm was proposed [Srlnka et al., 1979], later detailed mapping of LP magnetometer data showed that the main magnetic anomaly is offset from the rille and correlates instead with a smooth plains unit interpreted as primary or secondary basin ejecta [Hood et al., 2001]. The Rima
Sirsalis anomaly and the stronger Reiner Gamma anomaly are both elongated in a direction that is approximately radial to the Imbrium basin [Halekas et al., 2001; Hood et al., 2001]. It was therefore suggested that the sources of these anomalies consist of unusually magnetic Imbrium basin ejecta lying in many cases beneath the mare surface. The latter interpretation was supported by modeling of the Reiner Gamma anomaly at different altitudes, which showed that the anomaly source was probably shallow, between the surface and 1 km depth [Nicholas et al., 2007]. As discussed by Hood and Artemieva [2008] and Gattacceca et al. [2010], these ejecta materials were shocked during emplacement and therefore could have acquired shock remanent magnetization (SRM) in a transient magnetic field generated by interaction of the impact-produced vapor-melt cloud with the ambient lunar dynamo magnetic field.

The recent model of Wieczorek et al. [2012] is more consistent with the basin ejecta hypothesis for the sources of strong magnetic anomalies but proposes that the SPA ejecta acquired unusually strong magnetization because it was enriched in iron from the impactor. It also hypothesizes that thermoremanent magnetization (TRM) was acquired by thick ejecta deposits as they cooled in the presence of a steady core dynamo magnetic field. In addition, the model is consistent with evidence that the SPA basin is elliptical in shape and was therefore probably produced by an oblique impact in a northward or southward direction [Garrick-Bethell and Zuber, 2009]. One minor criticism of the model is that it may not be fully consistent with returned sample studies, which show that the most strongly magnetized breccias are those that contain metallic iron remanence carriers produced in situ by impact reduction of preexisting iron silicates [Stringway et al., 1973a; Nagata et al., 1974; Housley et al., 1973; Housley, 1977; Fuller and Cisowski, 1987]. The main carriers of strong remanence in lunar samples are native Fe particles in the single domain size range (tens to hundreds of nanometers) that are produced by the latter process [e.g., Fuller and Cisowski, 1987]. Metallic Fe-Ni of meteoritic origin is also found in lunar rocks [e.g., Rochette et al., 2010] but is usually coarse-grained (> 100 µm in size), which is in the more weakly magnetic multidomain size range. Nevertheless, the addition of a large quantity of multidomain iron may still have produced strong anomalies, and this material may not have been sampled or transferred to Earth in lunar meteorites. So, it is not possible to eliminate the model on this basis alone.

### 3. Other Antipodal Anomaly Concentrations and the Descartes Anomaly

In view of the lack of conclusive evidence favoring the hypothesis that antipodal effects of young basin-forming impacts were mainly responsible for producing strong anomalies on the south-central far side, it is necessary to consider strong anomalies elsewhere on the Moon, including both large concentrations and isolated anomalies. As noted by Wieczorek et al. [2012], many strong anomalies are not associated with basin antipodes (e.g., the Descartes, Reiner Gamma, and Rima Sirsalis anomalies) and most basins do not have magnetic anomalies concentrated near their antipodes. An absence of anomalies antipodal to older (early Nectarian and pre-Nectarian) basins could be ascribed to the effects of later impact gardening and mare basalt flooding. However, if the antipodal magnetization hypothesis is valid, most young large basins should show antipodal magnetization signatures. The same basins should also show some evidence for unusual terrain similar to the GM terrain if such terrain is closely associated with the source materials of the magnetic anomalies. Finally, studies of isolated strong anomalies on the geologically less complex near side may provide further tests of hypotheses for the sources of anomalies near SPA on the far side.

Table 2 lists lunar basins younger than Nectaris in approximate order of increasing relative age [after Wilhelms, 1984]. This ordering is only approximate and may contain significant errors (see, e.g., Spudis et al. [2011] for a recent study of uncertainties in the age of the Serenitatis basin). Nevertheless, to provide an objective means of selecting relatively young basins for analysis, this ordering is adopted here. As mentioned in the Introduction (section 1), previous work has found evidence for concentrations of magnetic anomalies near the antipodes of Orientale, Imbrium, Serenitatis, and Crisium. In this section, we reexamine the antipodal regions of Orientale and Crisium as well as the area on the near side that includes the strongest isolated anomaly on the Moon, the Descartes anomaly. In section 4, regions antipodal to the remaining four basins younger than Crisium are examined.

#### 3.1. Orientale Antipode Region

Figure 7 is a superposition of a scalar field magnitude contour plot onto a partial geologic map of the region centered on the eastern limb that includes the Orientale antipode [again after Hood and Williams, 1989]. The magnetic field map was produced from a combination of LP ascending node orbit passes during April and May of 1999 using the same methods described in section 2.1 and normalized to 25 km altitude. A series of two-dimensionally filtered anomaly maxima are distributed over a wide area centered roughly 10° east of the geometric antipode. The largest smoothed anomaly peak is 9 nT, somewhat less than the maximum anomaly amplitude near the Imbrium antipode in Figure 3. Unusual terrain, characterized as “furrowed and pitted material, Imbrian or Nectarian in age” on the geologic map of
Figure 7. Same format as Figure 3 but for the region centered on the eastern limb including the Orientale antipode (cross). The partial geologic map is after Hood and Williams [1989] and is based on the U.S.G.S. geologic map of Wilhelms and El-Baz [1977].

Wilhelms and El-Baz [1977], is shown in blue. The largest single occurrence of this terrain (hereafter referred to as FP terrain) is centered near and to the east of the antipode. Like the GM terrain, the FP terrain correlates only moderately with overhead magnetic field intensity; although the field is generally larger-than-average over the unit, much of the anomaly concentration (maxima of 9 and 7 nT) is located just southeast of the mapped terrain.

[39] Figure 8a is in the same format as Figure 4a and is an LROC WAC composite showing much of the area around the Orientale antipode (cross). The largest single anomaly in Figure 7 (smoothed amplitude 9 nT) is centered near the small fresh crater Goddard A, located north-northeast of Goddard. Swirl-like albedo markings are visible near this location despite the low sun illumination angle. Locations of prominent swirls are also shown in Figure 7 as small dark markings. Figure 8b is a higher-resolution view of part of the FP terrain located west of Hubble crater. Grooved lineations can be seen (arrows) in the southern wall of Hubble and in the walls of a smaller crater southwest of Hubble. Like the Imbrium antipode grooves, these lineations are oriented approximately perpendicular to the crater rims. The FP terrain west of Hubble resembles somewhat the GM terrain shown in Figure 4b in that it has a rougher texture than ordinary pre-Nectarian terrain and contains numerous lineations with widths of order 1–2 km and lengths of 2–10 km. But the undulations (mounds) seen in Figure 8b are smaller in scale and less prominent than those in Figure 4b.

Figure 8. (a) Region near the Orientale antipode (cross). The strongest magnetic anomaly shown in Figure 7 is centered northeast of the 93 km diameter crater Goddard. (b) Higher-resolution view of a small area northwest of the antipode (white rectangle in Figure 8a). Note the grooved terrain in the walls of several craters (arrows) and the “furrowed and pitted” material outside the craters identified on the geologic map of Wilhelms and El-Baz [1977].
Figure 9. Superposition of the two-dimensionally filtered magnetic field magnitude at ~25 km altitude (contour interval, 1 nT) onto an LROC WAC photomosaic of a section of the southeast-central near side.

[42] Figure 10 shows two higher-resolution views in the near vicinity of the anomaly. Figure 10a is an LROC WAC composite while Figure 10b is an Apollo metric camera photo with an opposite sun illumination direction. The hilly and furrowed terrain referred to by Wilhelms and El-Baz [1977] is especially visible in Figure 10b. The location of the strong anomaly peak is indicated by the value of 12 nT in Figure 10a and is centered over an area of higher albedo [Richmond et al., 2003]. Since this area is located near the Apollo 16 landing site, it has been studied extensively. Prior to the landing, it was thought that the Descartes and Cayley Formations were volcanic in origin [e.g., Milton, 1972]. Following the mission, which included geologic studies on surface traverses and the return of samples that had all experienced impact processing, it was generally agreed that both of these units (Cayley and Descartes) had an impact-related origin. By the early 1980s, it had been established that the Descartes Formation consists of basin ejecta materials resulting mainly from either the Imbrium event [Muehlberger et al., 1980; Hodges et al., 1973; Ulrich, 1973; Hodges and Muehlberger, 1981] or the Nectaris event [Head, 1974; Spudis, 1984; 1993].

[43] As seen in Figure 10a, the area has been partly “sculptured” by Imbrium secondaries, leading to linear features radial to Imbrium (dashed arrows). According to Muehlberger et al. [1980] and Hodges and Muehlberger [1981], some Imbrium-related sculpturing is also found in the Descartes terrain while similarly sculptured terrain is not found around Nectaris. They also argue that the rugged appearance of the Descartes terrain and possible interbedding with Cayley plains units suggest that it is younger than Nectaris. Finally, as seen in the figure, the Descartes material near Apollo 16 overlies the northern part of the crater Descartes, suggesting a southward emplacement direction.

[44] On the other hand, the area shown in Figure 10a is within a few hundred kilometers of the outermost rim of the Nectaris basin (centered about 300 km southeast of Theophilus). It is therefore very likely that Nectaris ejecta existed in the area prior to the Imbrium impact [e.g., Spudis, 1984]. According to some geologic interpretations, the Descartes Formation extends well south of Abulfeda [Wilhelms and McCauley, 1971], which would also favor a Nectaris origin. Muehlberger et al. [1980] also agree that Nectaris ejecta could be present at depth underlying the Imbrium ejecta. Photogeologic analyses alone are therefore not definitive in establishing the basin of origin of the ejecta materials that make up the Descartes Formation. A number of observations seem to favor an Imbrium origin but it can also be argued that a Nectaris origin is more likely.

[45] An alternate approach toward establishing the origin of the Descartes terrain near Apollo 16 has more recently been discussed by Norman et al. [2010]. They analyzed $^{40}$Ar-$^{39}$Ar ages and trace element compositions of proba-
Figure 11. Same format as Figures 3 and 7 but for the region centered on the southeastern far side including the Crisium antipode (cross partly covered by contour lines). The partial geologic map is after Hood and Williams [1989], which is based on the full geologic map of Scott et al. [1977].

3.3. Crisium Antipode Region

[46] Figure 11 is a contour map of the magnetic field magnitude normalized to 25 km altitude superposed onto a partial geologic map of the southwestern far side and western limb region [after Hood and Williams, 1989, which is based on Scott et al., 1977]. It includes most of the Orientale basin and the area centered on the Crisium antipode (cross). The anomaly map was produced using a combination of LP magnetometer passes from April, May, and June of 1999. As seen in the figure, anomalies are generally weak over most of the region with the exception of the group of anomalies centered south-southwest of the antipode. Although some weak anomalies appear to be present within Orientale, these are caused by external field noise that was incompletely removed by the editing and detrending mapping technique (see Hood [2011], Figure 16 for an alternate map at a higher altitude over part of the same region). Detailed examinations of individual LP orbit passes show that crustal fields are essentially undetectable within the inner Orientale basin.

[47] This group of strong anomalies in Figure 11 includes the strongest single maximum on the Moon (smoothed amplitude 16 nT at 25 km). It is located on the outer fringes of the Orientale ejecta blanket (green). It is also located within ~ 500 km of the outer topographic boundary of the SPA basin (compare with Figure 2). The origin of this anomaly group is uncertain. On the one hand, according to the model of Hood and Artemieva [2008], they may have originated from the antipodal effects of the Crisium impact (i.e., deposition of converging antipodal ejecta in the presence of an amplified magnetic field). On the other hand, it has been proposed that the sources of these anomalies, like those of the anomalies on the south-central far side, consist of iron-rich ejecta from the SPA impactor that cooled in a steady core dynamo magnetic field [Wieczorek et al., 2012]. In the latter case, the location of the anomaly group near the Crisium antipode would be fortuitous.

[48] Figure 12a is in the same format as Figures 4a and 8a; it is an LROC WAC composite showing the area around the Crisium antipode, including part of the Orientale basin and the crater Gerasimovich, which is just south of the main anomaly maximum (16 nT). Most of the pre-Nectarian topography is overlain and obscured by Orientale ejecta. It is therefore not possible to investigate whether any unusual (e.g., hilly and furrowed) terrain was produced at the time of the older Crisium impact. Figure 12b shows several degraded craters just northwest of the antipode that could retain some evidence of grooved terrain in their walls (arrows). However, this terrain is only visible in their eastern walls and at least some of the linear features are oriented radial to Orientale. So, it could alternatively be interpreted as an area that has been “sculptured” by Orientale secondaries. We conclude that there is no convincing evidence for unusual grooved and mounded or hilly and furrowed terrain near the Crisium antipode but that this is ascribable to obscuration by Orientale ejecta.

4. Anomalies in the North Polar Region

[49] As shown in previous sections, concentrations of strong magnetic anomalies and possible modified landforms are present near the antipodes of four of the eight youngest basins listed in Table 2 (Orientale, Imbrium, Serenitatis, and Crisium). In this section, we consider the remaining four basins that are younger than Crisium according to the relative age list estimated by Wilhelms [1984]: Schrödinger, Sikorsky-Rittenhouse, Bailly, and Hertzsprung. Of these, Schrödinger is the youngest, although it is not especially large (~ 320 km in diameter) while Hertzsprung is the largest (570 km in diameter).

[50] The antipodal region of Hertzsprung is located on the east-central near side within Mare Facunditatis in the pre-Nectarian Facunditatis basin. It is overlain by mare basalt flows of late Imbrian age, and there are no strong magnetic anomalies in this area. As demonstrated by the Reiner Gamma anomaly, the presence of thin overlying mare basalt flows does not necessarily preclude the survival of a strong anomaly source. However, if the subsurface magmatism...
that produced the surficial basalt flows penetrated near the anomaly sources, then these sources could have been thermally demagnetized. Therefore, the absence of significant anomalies in this antipodal region is not a decisive contradiction of the antipodal magnetization hypothesis. Also, as reviewed in section 2.2, numerical simulations show that antipodal ejecta deposition depends sensitively on impact angle and may not occur when an impact is nearly vertical. Thus, the small fraction of lunar basins produced by nearly vertical impacts would not have antipodal ejecta deposition or SRM signatures.

[51] The remaining three basins (Sikorsky-Rittenhouse, Bailly, and Schrödinger) are all located in the southern polar region (Figure 13). As seen in the figure, Schrödinger is the most pristine, consistent with its very young relative age assignment in Table 2 (second only to Orientale). Bailly is comparable in size to Schrödinger but is overlain by ejecta from the younger Orientale basin and the nearby Eratosthenian-aged crater Hausen [Wilhelms et al., 1979]. For this reason, its age ranking (fifth youngest) may be uncertain. In any case, the Bailly geometric antipode (67°N, 112°E) lies within ~ 60 km of the rim of the 212 km diameter Schwarzschild crater (Figure 14). The latter crater is Nectarian in age [Lucchitta, 1978] but is well preserved and could be younger than Bailly. If so, then any antipodal magnetic anomalies resulting from the Bailly impact would have been greatly weakened by shock demagnetization effects of the subsequent Schwarzschild impact [Halekas et al., 2002; 2003; Lillis et al., 2010]. Sikorsky-Rittenhouse is located adjacent to Schrödinger, is nearly covered with ejecta from that basin [Wilhelms et al., 1979], and is only weakly recognizable on topographic maps (Figure 13). It is only listed as a “possible” basin by Wilhelms [1984] while Frey [2011] found that the topographic quasi-circular depression (QCD) associated with this basin was actually characterized by a circular thick area. It is therefore very questionable whether Sikorsky-Rittenhouse is a basin at all, or, if it is, that it has a late Nectarian relative age. Unlike Hertzsprung and Bailly, the antipodal regions of Schrödinger and Sikorsky-Rittenhouse are in an older highland area, are not overlain by mare basalt, and are not near younger basins or large craters [Lucchitta, 1978]. Their antipodal regions should therefore preserve magnetization and/or landform modification signatures if they were produced at the respective times of impact.

[52] Figure 14 is a contour plot of the two-dimensionally filtered field magnitude at a normalized altitude of 25 km over the north polar region (60°N to the pole). This map was produced using 286 LP orbit passes at relatively low altitudes over these high northern latitudes selected from 6 months of measurements (February through July of 1999). The contour map is overlaid onto Lunar Orbiter Laser Altimeter (LOLA) topography. The smoothed anomaly

**Figure 13.** Lunar Reconnaissance Orbiter Laser Altimeter (LOLA) topographic map of the lunar south polar region (60°S to the pole). Inner circles are at latitudes of 70°S, 80°S, and 85°S. The locations of three basins with Imbrian or upper Nectarian ages according to Wilhelms [1984] are indicated.
amplitudes on the map (contour interval, 0.5 nT) are accurate to ±1 nT, but some minor artifacts are present near the pole due to incomplete removal of long-wavelength external fields (meridionally elongated contours). Most of the north polar region is characterized by relatively weak fields. Easily, the largest concentration of strong anomalies is centered near 294°E, 79°N, or approximately 8° of arc from the Schrödinger geometric antipode (314°E, 75°N, indicated by the white arrow). Smoothed anomaly amplitudes in this group are as large as 6.5 nT but are somewhat weaker than those near other basin antipoles considered above (e.g., 9 nT for the Orientale antipode group). Although the Sikorsky-Rittenhouse geometric antipode (69°N, 291°E) is also within ~10° of the center of this group of anomalies, it is questionable whether this is a true lunar basin, as discussed above. We therefore propose that this group of anomalies is an antipodal signature of the Schrödinger impact. Although it is offset somewhat in location from the geometric antipode, the offset (~8°) is comparable to that found for the Orientale antipode group, for example. Some offset is expected due mainly to asymmetries imposed in an oblique impact [Hood and Artemieva, 2008] and to the finite lunar rotation rate [Wieczorek and Zuber, 2001]. Several isolated anomalies are also present; the strongest one has a smoothed amplitude of ~2.5 nT and is located near 240°E, 78°N.

Figure 14. Two-dimensionally filtered field magnitude (nT) at ~25 km altitude in the north polar region (60°N to the pole). The contour map (contour interval: 1 nT) is superposed on LOLA topography.

Figure 15. (a) Section of the north polar region including the Schrödinger antipode (cross). The locations of magnetic anomaly maxima (2.5, 5.5, and 6.5 nT) shown in Figure 14 are indicated. (b) Possible hilly and furrowed terrain (arrows) south of Sylvester (area of lower white rectangle in Figure 15a).
basins on basin. This would increase to five the number of young antipodal magnetic and structural signature of a young lunar rain in the same area provide suggestive evidence of another for Imbrium).

Figure 16. (a) Part of Figure 15a (upper white rectangle) showing possible grooved terrain in the walls of Froelich (arrow). (b) Similar grooved terrain (arrows) in crater walls north of Mare Ingenii near the Imbrium antipode (upper white rectangle in Figure 4a).

antipode of Schrödinger. As indicated by the arrow, possible grooved terrain is present in the western wall of Froelich with an orientation perpendicular to the crater rim. For comparison, Figure 16b is a WAC composite image at the same resolution showing several craters north of the Ingenii basin near the Imbrium antipode (upper white rectangle in Figure 4a). While the Froelich wall “grooves” are not identical to those in craters near Ingenii, they are similar enough to be considered as possible grooved terrain associated with the Schrödinger impact. The more subdued appearance of the Froelich grooves is attributable to the smaller size of the Schrödinger basin (320 km versus 1200 km for Imbrium).

5. Discussion and Conclusions

[55] Overall, the group of strong anomalies centered within 8° of the Schrödinger antipode and the unusual terrain in the same area provide suggestive evidence of another antipodal magnetic and structural signature of a young lunar basin. This would increase to five the number of young basins on Wilhelms’ [1984] list with probable antipodal sig-

natures. As discussed in the previous subsection, the three remaining basins that are younger than Crisium have antipodal regions that have either been modified by later impacts and magmatic eruptions (Hertzsprung and Bailly) or are not clearly recognized as a true basin (Sikorsky-Rittenhouse). It can therefore be rationalized why antipodal signatures are not found for these basins.

[56] While none of the five positive cases is individually convincing, the combination represents fairly compelling evidence that most strong lunar magnetic anomalies, including those near the northwest rim of SPA, originated as a consequence of antipodal effects of lunar basin-forming impacts. Consistent with these results, Mitchell et al. [2008] carried out a Monte Carlo simulation and found that the probability of finding antipodal magnetic enhancements only for the largest post-Nectarian basins was less than 0.1%.

[57] In view of evidence that many other lunar anomalies (including the Descartes anomaly) are associated with surface geologic units consisting of basin ejecta, the most probable mechanism for producing the antipodal anomalies is the deposition of converging ejecta in the presence of a magnetic field. It is possible that the antipodally deposited impact melt and rocky material was especially susceptible to strong magnetization because of shock-induced production of more Fe-Ni remanence carriers (or more carriers in the single-domain size range). Numerical simulations for an Imbrium-scale event show that most ejecta impacting in the antipodal zone originate in the lower crust or upper mantle beneath the basin [Hood and Artemieva, 2008]. This material would have been more mafic, which may have been a factor in the production of more remanence carriers during the impact. The same numerical simulations together with laboratory studies of shock effects in returned samples strongly suggest that shock remanent magnetization (SRM) was a likely magnetization process and that the ambient core dynamo magnetic field was transiently distorted and amplified at the time of SRM acquisition [Hood and Artemieva, 2008; Gattacceca et al., 2010]. The derivation of antipodal ejecta mainly from the lower crust should result in antipodal compositional anomalies as well as unusual terrain resulting from shock effects on the impacting ejecta. This expectation may be consistent with enhanced Th abundances near the Imbrium antipode, as argued by Haskin [1998], since the lower crust and upper mantle beneath Imbrium are known to be enriched in KREEP [e.g., Jolliff et al., 2000].

[58] As discussed in section 3.2, the isolated Descartes anomaly on the lunar near side (which is not antipodal to any basin) may provide useful additional insights into the identity of the strong anomaly sources. This anomaly is unique in that (a) it is the strongest isolated anomaly on the Moon; (b) it occurs over exposed terrain that is morphologically similar (furrowed and pitted) to that found near the Orientale antipode [Schultz and Gault, 1975; Wilhelms and El-Baz, 1977]; and (c) its probable source materials (the Descartes Formation near the Apollo 16 landing site) have been at least partly sampled. Previous geologic studies have established that the Descartes Formation consists of basin ejecta materials, either Imbrium ejecta or Nectaris ejecta reworked by the Imbrium impact [e.g., Head, 1974; Muehlberger et al., 1980; Spudis, 1984]. The Descartes terrain near the Apollo 16 landing site, where the strong magnetic anomaly is found,
may therefore be a good candidate for a prototypical strong lunar magnetic anomaly source.

[69] One recent laboratory study of Descartes breccias has yielded a probable age (3.866 ± 0.009 Gyr) that is close to that of the Imbrium basin and also infers that KREEP-bearing rocks contributed to their composition [Norman et al., 2010]. The latter authors have therefore interpreted these properties as implying that the Descartes terrain consists of Imbrium basin ejecta. If their interpretations are correct, then this terrain may be similar in origin to that which occurs antipodal to Imbrium and is suspected to be the source material for strong magnetic anomalies and enhanced Th anomalies. This component of Imbrium basin ejecta may have been especially susceptible to strong magnetization (due to enhanced shock-induced abundances of single domain Fe-Ni remanence carriers, for example) and may have originated in the lower crust beneath Imbrium where KREEP is enriched. Although most such ejecta would have been preferentially deposited near the antipode [Hood and Artemieva, 2008], some of these ejected materials may have been deposited in isolated locations nearer the basin (e.g., due to mutual collisions of ejecta in an actual impact), as needed to explain anomalies such as Descartes, Reiner Gamma, and Rima Sirsalis. Acquisition of SRM may have occurred in a moderately amplified magnetic field in a turbulent boundary layer upstream of the vapor-melt cloud as it expanded from the impact point [see Hood and Artemieva, 2008, Figure 9b].

[60] The hypothesis that most sources of relatively strong lunar magnetic anomalies consist of antipodal basin ejecta deposits must meet the constraint that required magnetization intensities are not unreasonably large. Modeling of the largest single anomaly near the Imbrium antipode (smoothed amplitude 12 nT in Figures 2 and 3) indicates that a source consisting of a near-surface disk with radius ~70 km and dipole moment $5 \times 10^{13}$ A m$^2$ is sufficient to explain the anomaly. Numerical simulations for an Imbrium-scale basin-forming impact yield estimated mean antipodal ejecta thicknesses ranging from ~300 m for a 100 km diameter impactor to ~1100 m for a 200 km diameter impactor [see Hood and Artemieva, 2008, Figure 7b]. Antipodal ejecta thicknesses for a real impact would not be constant but would vary by factors of at least two or three; thicknesses of up to 2 or 3 km could therefore be present locally. The Descartes terrain at the location of the Descartes magnetic anomaly must be at least several kilometers thick since it covers the northern edge of the crater Descartes (Figure 10a). For a 3 km thickness of the model disk for the Imbrium antipode anomaly, a mean magnetization intensity of order $\sim 1$ A m$^{-1}$ is needed. A similar mean intensity is needed to explain the amplitude of the Descartes anomaly if the ejecta deposit source model is valid.

[61] A mean magnetization intensity of order 1 A m$^{-1}$ could have been acquired in a thick layer of antipodal ejecta material if (a) magnetization was acquired rapidly via SRM in a transiently amplified magnetic field; or (b) magnetization was acquired by slow cooling via TRM in the lunar dynamo magnetic field provided that magnetization susceptibilities of this material were considerably larger than those of returned Apollo impact-melt breccias. As reviewed in Wieczorek et al. [2012], the expected thermoneremant magnetization intensity $M_s$ can be related to the applied field $B$ (in units of Tesla) by $M_s = BM_s/a$, where $M_s$ is the saturation remanent magnetization intensity and $a$ is a proportionality constant equal to about $3 \times 10^{13}$ for equant grains of multidomain metallic iron. $M_s \simeq 12$ A m$^{-1}$ for returned Apollo mafic impact-melt breccias [see Wieczorek et al., 2012, Table S2], while lunar dynamo surface field intensities during the impact basin-forming era may have been as large as ~50 $\mu$T [Garrick-Bethell et al., 2009; Shea et al., 2012]. Thus, if antipodal magnetic sources formed by slow cooling in the steady lunar dynamo field, values of $M_s$ not much larger than 0.2 A m$^{-1}$ are to be expected if magnetization susceptibilities of antipodal ejecta material are comparable to those of returned samples from the lunar near side. As noted above, it is possible that these susceptibilities were in fact larger than those of nearside impact-melt breccias because of derivation of this material from the more mafic lower crust and upper mantle and because of high applied shock pressures during the impact. Thus, a TRM origin remains marginally possible if the most favorable parameters are considered. But, as discussed by Gattacceca et al. [2010], an SRM origin for the strong antipodal magnetization enhancements is more easily understood, especially if acquisition occurred in a transiently amplified magnetic field. On the basis of laser shock experiments on mare basalt samples in a controlled magnetic field, it was found that magnetization intensities as large as 1 A m$^{-1}$ could have been acquired via SRM in an applied field of several hundred microteslas.Transient magnetic field amplitudes of this order could have been produced in basin antipole zones and/or in a turbulent boundary layer produced as the impact vapor-melt cloud expanded around the Moon [Hood and Artemieva, 2008].

[62] It is worth noting that significant magnetic anomalies are not expected to be produced antipodal to a relatively young (post-dynamo) lunar impact crater. This question is raised by recent evidence mentioned in section 2.2 for unusual young impact melts and rocky exposures antipodal to the 86 km diameter Tycho crater [Robinson et al., 2011; Bandfield et al., 2013] that could be a consequence of antipodal ejecta deposition [Artemieva, 2013]. First, even if a younger crater-forming impact produced a vapor-melt cloud that was sufficiently massive to remain a collisional gas and expand thermally around the entire Moon as occurs for a basin-scale event [Hood and Artemieva, 2008], the ambient magnetic field (the present-day solar wind field) would have had an amplitude of no more than 10 nT, which is a factor of 1000 to 5000 less than the former lunar dynamo surface field. Transient fields in the antipodal zone would therefore have been relatively small and/or the region of compressed field amplification would have been much smaller in scale, leading to relatively weak or inhomogeneous SRM intensities. Second, impact melt and solid rocky ejecta in the antipodal zone of an impact crater are much less massive than that antipodal to a lunar impact basin. Consistently, an examination of LP magnetometer data from April of 1999 (when external field variations were relatively minimal during passes over the Tycho antipode region) yields no evidence for a discernible anomaly larger than 1 nT at 33 km altitude.

[63] Finally, the probable importance of basin ejecta sources of strong lunar magnetic anomalies, which likely acquired their magnetization via SRM in a transient...
magnetic field, does not preclude the existence of some anomaly sources that acquired thermoremanent magnetization (TRM) in a steady core dynamo magnetic field. Such sources are of special interest because they can potentially provide information about the history of the dynamo and the position of the rotational pole (true polar wander).

First, many anomalies within large lunar basins most probably have sources consisting of impact melt that acquired TRM in the core dynamo magnetic field [Hood, 2011]. This is because numerical simulations demonstrate that the interiors of such basins would have been heated to temperatures exceeding the solidus to a large depth, thereby thermally erasing any preexisting shock magnetization and requiring long time periods (several Myr) to completely cool through the Curie blocking spectrum. Anomalies within the Crisium basin, which are moderately strong, fall into this category, for example. Modeling of one of these anomalies yielded a paleomagnetic pole position not far from the present rotational pole, supporting a core dynamo magnetizing field [Hood, 2011]. Only if ejecta from a later basin-forming impact is deposited within an older basin would sources in the older basin have an SRM origin.

Second, if a basin ejecta deposit is sufficiently massive and remains partially molten, then it may require a long time interval to cool through the Curie temperature, allowing TRM acquisition in the core dynamo field. However, modeling of strong magnetic anomalies suggests that the latter situation is not common. For example, modeling of the Reiner Gamma anomalies, which consist of a strong southwest main anomaly and a weaker northeast anomaly, yields quite different directions of magnetization for the two anomaly sources [Kurata et al., 2005]. This is consistent with previous studies of magnetic anomalies over the Fra Mauro Formation, which found evidence for different magnetization directions for two nearby exposures [Hood, 1980]. Therefore, caution should be exercised when modeling most strong lunar magnetic anomalies to estimate directions of magnetization for the purpose of constraining true polar wander. It is very possible that many sources of these anomalies acquired their magnetizations in transient magnetic fields that were not aligned with the core dynamo field.

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