Exposure of spectrally distinct material by impact craters on Mercury: Implications for global stratigraphy

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\textbf{Abstract}

MESSENGER’s Mercury Dual Imaging System (MDIS) obtained multispectral images for more than 80\% of the surface of Mercury during its first two flybys. Those images have confirmed that the surface of Mercury exhibits subtle color variations, some of which can be attributed to compositional differences. In many areas, impact craters are associated with material that is spectrally distinct from the surrounding surface. These deposits can be located on the crater floor, rim, wall, or central peak and in the ejecta deposit, and represent material that originally resided at depth and was subsequently excavated during the cratering process. The resulting craters make it possible to investigate the stratigraphy of Mercury’s upper crust. Studies of laboratory, terrestrial, and lunar craters provide a means to bound the depth of origin of spectrally distinct ejecta and central peak structures. Excavated red material (RM), with comparatively steep (red) spectral slope, and low-reflectance material (LRM) stand out prominently from the surrounding terrain in enhanced-color images because they are spectral end-members in Mercury’s compositional continuum. Newly imaged examples of RM were found to be spectrally similar to the relatively red, high-reflectance plains (HRP), suggesting that they may represent deposits of HRP-like material that were subsequently covered by a thin layer (~1 km thick) of intermediate plains. In one area, craters with diameters ranging from 30 km to 130 km have excavated and incorporated RM into their rims, suggesting that the underlying RM layer may be several kilometers thick. LRM deposits are useful as stratigraphic markers, due to their unique spectral properties. Some RM and LRM were excavated by pre-Tolstojan basins, indicating a relatively old age (>4.0 Ga) for the original emplacement of these deposits. Detailed examination of several small areas on Mercury reveals the complex nature of the local stratigraphy, including the possible presence of buried volcanic plains, and supports sequential buildup of most of the upper ~5 km of crust by volcanic flows with compositions spanning the range of material now visible on the surface, distributed heterogeneously across the planet. This emerging picture strongly suggests that the crust of Mercury is characterized by a much more substantial component of early volcanism than represented by the phase of mare emplacement on Earth’s Moon.

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1. Introduction

Mercury is the least visited of the terrestrial planets, having been encountered by only two spacecraft. More than half of the planet remained unseen after three flybys by Mariner 10 in 1974–1975. Mercury then remained unvisited until 2008, when the MESSENGER spacecraft completed two of its three planned flybys (the first on 14 January 2008 and the second on 6 October 2008) on its way to insertion into orbit about the innermost planet (Solomon et al., 2007). MESSENGER’s Mercury Dual Imaging System (MDIS), consisting of a monochrome, narrow-angle camera (NAC) and an 11-band (400–1040 nm), wide-angle camera (WAC) (Hawkins et al., 2007), has imaged over 80\% of the surface to date, including approximately 50\% seen for the first time by a spacecraft. The Mariner 10 and MESSENGER images reveal that Mercury has subtle yet distinct color variations across its surface, many of which can be attributed to compositional heterogeneity (Hapke et al., 1975; Dzurisin, 1977; Schultz, 1977; Rava and Hapke, 1987; Robinson and Lucey, 1997; Robinson and Taylor, 2001; Blewett et al., 2007, 2009; Robinson et al., 2008; Murchie et al., 2008).
Fundamental questions concerning the nature of Mercury's crust remained after the Mariner 10 flybys. To first order, Mercury's heavily cratered surface resembles that of the Moon, though it lacks a strong albedo contrast like that between the lunar maria and highlands. There was no consensus, however, on whether Mercury's smooth plains originated by impact (e.g., Oberbeck, 1975; Wilhelms, 1976) or volcanism (e.g., Murray et al., 1974; Trask and Strom, 1976; Strom, 1977; Dzurisin, 1978; Kiefer and Murray, 1987). Color-ratio maps made from the Mariner 10 ultraviolet and orange color channels (355 and 575 nm) enabled the identification of distinct surface units on the basis of relative color (Hapke et al., 1975, 1980; Dzurisin, 1977; Schultz, 1977; Rava and Hapke, 1987). Later recalibration and reanalysis of the Mariner 10 data allowed the correlation of many color boundaries with geomorphic boundaries, suggesting that a substantial number of the smooth plains are volcanically in origin (Robinson and Lucey, 1997; Robinson and Taylor, 2001). Mercury's albedo variations are subtle compared to those of the Moon but exhibit a range more extensive than that of the lunar highlands and similar to variations among the maria (Denevi and Robinson, 2008).

The higher resolution, higher signal-to-noise ratio, near-global coverage, and expanded wavelength range of the MESSENGER images permit a more detailed analysis of Mercury's surface properties. The MDIS images (Robinson et al., 2008), as well as Mercury Atmospheric and Surface Composition Spectrometer data (McClintock et al., 2008) are consistent with early ground-based observations of relatively featureless (no obvious absorption bands) reflectance spectra having overall positive ("red") slopes with increasing wavelength (e.g., McCord and Adams, 1972; Vilas, 1985, 1988). Despite its relatively featureless spectra, the surface of Mercury does exhibit subtle spectral variations, which are due primarily to changes in albedo and the steepness of the spectral slope over the MDIS wavelength range (Denevi et al., 2009). These spectral differences have been used to identify several major and minor color units (Robinson et al., 2008; Murchie et al., 2008; Denevi et al., 2009; Blewett et al., 2009). Although the first-order color variations are associated with maturity effects, some color variations can be attributed to compositional differences (Robinson and Lucey, 1997; Blewett et al., 2007; Robinson et al., 2008; Murchie et al., 2008).

MESSENGER images confirm the Mariner-10-based hypothesis that widespread volcanism occurred on the surface of Mercury and played an important role in the planet's surface evolution (Head et al., 2008, 2009a,b; Murchie et al., 2008; Denevi et al., 2009). Little is known about the three-dimensional structure of the crust, yet knowledge of Mercury's crustal stratigraphy is necessary to understand the history and evolution of the planet. Without in situ measurements, there are very few ways to observe what lies beneath the surface. Impacts excavate material from depth, making it possible to assess vertical heterogeneities in the crust using the resulting craters. Large basins in particular can excavate material from depths of tens of kilometers.

In this paper, we investigate the stratigraphy of Mercury's upper crust by means of craters that excavated spectrally distinct material. Several areas of Mercury are examined in detail using WAC and NAC images. Studies of laboratory, terrestrial, and lunar craters, along with theoretical models of the cratering process, allow us to estimate the depths of origin of the spectrally distinct ejecta. From these results we construct stratigraphies of the upper crust in these areas to assess how Mercury's crust varies with depth, to make preliminary assessments about the distribution of color units, and to progress toward an understanding of how the crust evolved.

2. The colors of Mercury

The major and minor color units of Mercury have been newly redefined using data from the first two MESSENGER flybys (Robinson et al., 2008; Murchie et al., 2008; Denevi et al., 2009; Blewett et al., 2009). Here, we briefly summarize these units using the definitions of Denevi et al. (2009) and Blewett et al. (2009). We refer to units as having relatively “red” (steeper) or “blue” (shallower but still positive) spectral slopes through the MDIS wavelength range. It is important to note that these terms refer to color relative to Mercury's widespread intermediate terrain (Robinson et al., 2008; Denevi et al., 2009), and that all materials observed on Mercury exhibit positive spectral slopes.

Three major terrain classes are defined on the basis of spectral slope, reflectance, and morphology: smooth plains, spectrally intermediate terrain (IT), and low-reflectance material (LRM). The smooth plains themselves are subdivided into three types: high-reflectance red plains (HRP), intermediate plains (IP), and low-reflectance blue plains (LBP). Representative MDIS spectra of these color units are shown in Fig. 1. The IT generally corresponds to areas with higher crater density than the IP (Denevi et al., 2009). The IT and IP have similar spectral properties, suggesting they may have formed in a similar manner at different times (Denevi et al., 2009).

HRP represent the brightest and reddest of the smooth plains, with average reflectance values as much as 20% higher than the global mean at 750 nm. IP appear spectrally similar to IT but have lower crater densities. These intermediate units exhibit spectral properties similar to the global mean. The LBP are intermediate between IP and LRM in both reflectance and spectral slope. Typical LRM deposits have reflectance values as much as 30% below the global mean and are bluer than most other materials on Mercury (spectral slopes ~5% lower than HRP). The LRM likely constitutes a compositional end-member (Denevi et al., 2009).

The observed color variations are primarily associated with subtle changes in the steepness of the spectral slope, which differs by no more than ~5% between the reddest and bluest of the major
color units. The general spectral trend observed on the surface is decreasing reflectance with the decreasing (bluer) spectral slope. This trend is opposite to that expected from space weathering, where materials darken and redden over time (e.g., Fischer and Pieters, 1994). Therefore, the spectral variations among the major color units can be attributed to compositional differences (Robinson et al., 2008; Blewett et al., 2009). Most likely, the units HRP, IP/IT, LBP, and LRM can be considered to represent a compositional continuum.

There are three additional, areally minor color units found on Mercury’s surface (e.g., Robinson et al., 2008). These units stand out prominently in color composites and tend to be associated with impact craters. One minor unit consists of fresh (immature) crater ejecta, which have been less affected by space weathering than typical surface units on Mercury and appear brighter and bluer than their mature counterparts (Robinson et al., 2008). The albedo differences between fresh and mature material produce the most extreme color variations seen on Mercury’s surface (Robinson et al., 2008). In many cases fresh deposits are not compositionally different from the underlying materials, so they will not be discussed further in this study.

The second minor color unit consists of bright crater-floor deposits (BCFDs) (e.g., Robinson et al., 2008; Blewett et al., 2009). BCFDs are spectral outliers on the surface of Mercury (see Fig. 1), with high reflectance values and relatively blue (shallow) spectral slopes, implying a distinctive composition (Blewett et al., 2010). These units appear to be associated with impact craters, but the origin and composition of the BCFDs are not yet known.

The last of the minor color units is characterized by a relatively red spectrum and an elevated albedo, first noted in Mariner 10 data (Rava and Hapke, 1987; Dzurisin, 1977; Schultz, 1977), and subsequently recognized in MESSENGER data (Robinson et al., 2008; Murchie et al., 2008; Head et al., 2008; Blewett et al., 2009; Kerber et al., 2009). These red units, sometimes referred to as “red spots,” have either primary or secondary associations with craters and basins. Primary associations are those where the red material (RM, see Fig. 1) is a part of the structure of the crater, having been emplaced during the impact process. Lunar dark-halo craters (DHs), sites where impact craters excavate mare basalt from beneath a highland-rich ejecta deposit, are analogous features on the Moon (e.g., Schultz and Spudis, 1979; Hawke and Bell, 1981; Bell and Hawke, 1984; Head and Wilson, 1992; Blewett et al., 1995; Antonenko et al., 1995). Primary occurrences of red spots can be located on the crater floor, rim, wall, or central structures, or in the ejecta deposit. Secondary associations are those where the red unit is found in or around an impact structure but was emplaced as a result of a subsequent process rather than the impact itself. These occurrences are typically associated with rimless depressions located on the floor or in the walls of craters and basins and are thought to be of pyroclastic origin (Robinson et al., 2008; Murchie et al., 2008; Head et al., 2008, 2009a; Blewett et al., 2009; Kerber et al., 2009). We focus on the primary associations in this study.

3. Determining depth of origin from impact craters

There are varying degrees to which an impact crater can provide insight into the local stratigraphy, as illustrated in Fig. 2. If an impact does not penetrate through the surface layer, the ejecta will have the same composition as the surface material (Fig. 2a), with maturity effects dominating any spectral differences between the excavated material and the surrounding terrain. In such cases, the minimum thickness of the surface layer can be derived. If an impact penetrates through the surface layer and excavates material of a different composition, the ejecta will be spectrally distinct from the surrounding terrain (Fig. 2b), as in a lunar dark-halo crater. In this case, the thickness of the surface layer can be constrained and a minimum thickness of the underlying material can be derived. The maximum thickness of the second material will remain unconstrained unless the impact also exposes an additional color unit (e.g., in a central peak, Fig. 2c).

During crater excavation, material from shallower depths is thrown the farthest from the crater center. Closer to the crater rim, these ejecta are covered by the deposition of material from deeper layers (e.g., Stöffler et al., 1975). The resulting overturned flap of ejecta at the crater rim exhibits a stratigraphy inverted from that of the pre-impact subsurface, with the most deeply excavated materials visible at the surface; therefore, spectrally distinct material located on a crater rim should have been derived from depths approaching the maximum excavation depth of the crater.

3.1. Maximum excavation depth of ejecta

Ejecta generated by an impact event can originate from no deeper than the maximum depth of excavation. This depth is approximately equal to one-tenth the transient crater diameter ($D_{tc}$, measured from the pre-impact surface) for vertical impacts and is shallower for oblique ones (Gault et al., 1968; Melosh,
The excavation depth achieved by a vertical impact is therefore a maximum value.

The transient crater represents the maximum extent of the growing cavity at the end of the excavation stage, before any collapse has occurred. Because $D_{te}$ is not observable in post-impact images, it is necessary to relate $D_{te}$ to the observable rim-to-rim diameter ($D_{r}$). Crater size greatly affects such a relationship. Both complex and simple craters are larger in diameter and shallower than their associated transient craters. Due to extensive modification (e.g., rim failure, slumping) the final $D_{r}$ of a complex crater is substantially enlarged relative to the equivalent-sized simple crater. Target-specific conditions also must be considered, including surface gravitational acceleration, average impact speed, and target strength (e.g., Cintala et al., 1977; Pike, 1980; Stewart and Vallant, 2006). These factors affect the diameter at which simple craters transition to complex ones. This transition diameter ($D_{t}$) is an important parameter used in deriving the relationship between the observed rim-to-rim diameter and the transient crater diameter. We use $10.3 \pm 4$ km as $D_{t}$ for Mercury, as determined from Mariner 10 data by Pike (1988). This variation in the value of $D_{t}$ results in a ±4% variation in the calculated transient crater diameter, which is not significant, considering our rule-of-thumb estimate that the maximum depth of excavation is equal to one-tenth of $D_{t}$.

Three analytical methods relating $D_{r}$ to $D_{te}$ were derived from lunar and terrestrial observations (Croft, 1985; Melosh, 1989; Holsapple, 1993). The most comprehensive of these studies (Holsapple, 1993) also incorporated experimental observations and impact scaling rules. All three methods can be written in the form of a power law:

$$D_{te} \cong CD_{r}^{k}$$

The values of constants $C$ and $k$ are given in Table 1. The relationships of $D_{r}$ to both $D_{te}$ and the maximum excavation depth for all three methods are shown in Fig. 3. The Croft and Holsapple relationships incorporate transition diameter into the constant $C$, making these methods more suitable for analyzing Mercury craters as they inherently account for the differences in complex and simple crater collapse and for some planet-specific variables such as gravity and target strength. The Melosh relationship was derived only for lunar and terrestrial craters and requires separate constants for simple and complex craters and for each planetary body.

The maximum excavation depths reported in the following sections are calculated with the Holsapple method for vertical impacts. The equivalent values using the Croft or Melosh methods are provided in Table 2. The relative depths and stratigraphy in this area will remain the same with any of the three methods; only the absolute depth values are affected. Because all three methods were derived in large part from lunar and terrestrial data, modifications may be necessary to translate the relationships to Mercury's higher-impact-speed environment.

There is a large spread in the expected impact speed at Mercury, with mean values typically quoted as being between ~20 and 40 km/s (e.g., Hartmann et al., 1981; Horedt and Neukum, 1984; Schultz, 1988; Cintala, 1992; Marchi et al., 2005; Le Feuvre and Wieczorek, 2008). Despite the range in absolute values, the relative mean impact speed at Mercury is approximately a factor of two higher than that at the Moon. For this study, we take the mean impact speed on Mercury to be 42.5 km/s, compared to 19.4 km/s for the Moon and 20.4 km/s for the Earth, on the basis of recent work (Le Feuvre and Wieczorek, 2008). Low-speed (Barnouin-Jha et al., 2007) and preliminary higher-speed (Barnouin-Jha et al., 2009) experimental investigations indicate that increases in impact velocity may lead to decreases in the depth of the transient crater, and thus the maximum depth of excavation, with an effect of up to ~20% for a velocity difference such as that between the mean impact speeds at the Moon and Mercury. This area warrants further study, but the values calculated for lunar impact speeds still provide maximum values for depth of excavation that would hold in a 40 km/s environment, and a broad range in the speed of projectiles striking Mercury will lead to impacts in the speed range of those occurring on the Moon.

### Table 1

| $D_{te}$ & $D_{r}$ | Source          |
|------------|----------------|-----------------|
| 0.85       | 0.5–0.65       | Melosh (1989) – complex |
| 0.84       | 1              | Melosh (1989) – simple |
| 0.75       | 0.92           | Holsapple (1993) |
| 0.50       | 0.85           | Croft (1985) |

Fig. 3. Rim-to-rim diameter as a function of transient crater diameter calculated using the methods of Holsapple (1993), Melosh (1989), and Croft (1985). The maximum excavation depth is taken to be one-tenth of the transient crater diameter.
The melt zone typically has a circular shape with a maximum depth of origin of peak rings can be determined following the same logic as for central peaks. Peak rings occur away from the crater center, and the maximum melting depth beneath the peak ring location corresponds to the minimum depth of origin of the peak ring.

Three methods are used to calculate the relationship between \( D_{sc} \) and the maximum melting depth for craters on the Moon, Earth, and Mercury. In all cases, either dense basalt or dunite is assumed to compose both the projectile and target, and the impacts are assumed to be vertical. Because these methods determine the volume or depth of melt relative to impactor diameter, we calculate the corresponding \( D_{sc} \) on a given planetary body using the Schmidt and Housen (1987) scaling relation:

\[
D_{sc} = 1.16 \left( \frac{\rho_{p}}{\rho_{t}} \right)^{1/2} \frac{v^{2.78}}{\rho^{0.44}} g^{-0.22},
\]

where \( \rho_{p} \) and \( \rho_{t} \) are the densities of the projectile and target, respectively, \( v \) is the impact velocity, and \( g \) is the surface gravitational acceleration of the body. The relationship between projectile size and transient crater size is the only place the gravitational acceleration of the body appears in these calculations.

The first method follows the calculations of Cintala (1992) and Cintala and Grieve (1998a,b), which were derived analytically. The pressure, \( P \), at a given location is determined by

\[
P \sim P_{\text{max}} \left( \frac{x}{\Delta} \right)^{-2} \cos^{2/3} \theta
\]

This equation is used to calculate the shape of the melt zone, where \( P_{\text{max}} \) is the maximum pressure occurring during the impact, \( x \) is the distance of the pressure zone from the impact point in cylindrical coordinates, \( \rho \) is the initial ratio of target compression to projectile compression, and \( \theta \) is the angle measured from the center line beneath the impactor to the point of interest in the target (Cintala, 1992). The phase changes are determined by calculating the enthalpy along the shock Hugoniot using a Murnaghan equation of state [which relates pressure to density non-linearly but does not incorporate a dependence on internal energy, see Appendix II.4 of Melosh (1989)]. [Cintala, 1992]. Solving this equation for \( x \) when \( \theta = 0^\circ \) for the pressure at which complete melting of the target material occurs gives a measure of the maximum depth of melting.

In order to match the results of Cintala (1992) and Cintala and Grieve (1998a,b), we add a penetration term to the calculated depth to account for the initial penetration of the projectile into the target before the projectile is fully coupled and transfers its energy to the target. This value was calculated through impedance matching following Gault and Heitowit (1963). These calculations are in reasonable agreement with those of other more complex numerical models [see Fig. 3 of Cintala (1992), O’Keefe and Ahrens (1975, 1977), and Austin et al. (1980)].

The second method uses the melt calculations of Pierazzo et al. (1997), derived from numerical hydrocode calculations of vertical impacts. These calculations incorporate an Analytical Equation of State (ANEOS), which derives thermodynamic functions (e.g., pressure) from the Helmholtz free energy (Thompson and Lason, 1972). As in the case of Cintala and Grieve (1998a,b), the equation of state is used to compute the entropy along the shock Hugoniot at which phase changes in the target occur. The volume of melt expected for a given impact is calculated using the dunite-specific fits to the numerical calculations relating the radius and depth of the complete melting region to the impact velocity from Table III of Pierazzo et al. (1997).

Both Cintala and Grieve (1998a, their Fig. 8) and Pierazzo et al. (1997, their Fig. 12) compare their derived impact melt volumes with terrestrial data, indicating good agreement with the observational data. These comparisons are reproduced in Fig. 4. The Cintala and Grieve values tend to overestimate the melt, whereas the Pierazzo et al. values are more consistent with the terrestrial observations; however, the amounts of melt measured in terrestrial craters will tend to be minimum estimates, as erosion over time will erase the melt zone temporarily.

### Table 2
Measured craters on Mercury.

| Region          | Crater name | Diameter (km) | Minimum depth of central peak origin (km) | Maximum depth of excavation (km) |
|-----------------|-------------|---------------|-----------------------------------------|---------------------------------
|                  |             |               |                                        | Holsapple method | Croft method | Melosh method (complex) |
| Rudaki plains    | Calvino     | 68            | 6                                       | 4.4 | 5.1 | 3.4–4.4 |
| R1              | 16          |               | 1.2                                     | 1.5 | 0.8–1.0 |
| Titian basin    | Titian      | 121           | 12                                      | 7.6 | 8.4 | 6.1–7.9 |
| T1              | 4           |               | 0.3                                     | 0.5 | 0.2–0.3 |
| T2              | 9           |               | 0.7                                     | 0.9 | 0.5–0.6 |
| T3              | 7           |               | 0.6                                     | 0.7 | 0.4–0.5 |
| T4              | 8           |               | 0.6                                     | 0.8 | 0.4–0.5 |
| Homer basin     | Dominici    | 20            | –                                       | 1.4 | 1.8 | 1.0–1.3 |
| H1              | 11          |               | 0.8                                     | 1.1 | 0.6–0.7 |
| H2              | 38          |               | 2.6                                     | 3.1 | 1.9–2.5 |
| H3              | 15          |               | 1.1                                     | 1.4 | 0.8–1.0 |
| H4              | 15          |               | 1.1                                     | 1.4 | 0.8–1.0 |
| H5              | 50          | 4             | 3.3                                     | 4.0 | 2.5–3.3 |
| Hemingway region| Hemingway   | 130           | 13                                      | 8.1 | 8.9 | 6.5–8.5 |
| de Grant        | 65          | 6             | 4.3                                     | 4.9 | 3.3–4.2 |
| He1             | 43          |               | 2.9                                     | 3.5 | 2.2–2.8 |
| He2             | 30          |               | 2.1                                     | 2.6 | 1.5–2.0 |
| Caloris basin   | Arget       | 100           | 10                                      | 6.3 | 7.1 | 5.0–6.5 |
| Apollodorus     | 41          | 4             | 2.8                                     | 3.3 | 2.1–2.7 |
| Cunningham      | 37          | 3             | 2.5                                     | 3.1 | 1.9–2.4 |
evidence of portions of the melt (Grieve and Cintala, 1992). Therefore, the Pierazzo et al. values likely underestimate the amount of melt generated during an impact.

Here, we incorporate a third method of calculating the volume and depth of melt produced during an impact. This method is based on the W.A. Watters et al. (2009) “foundering shock” method for calculating shock heating. W.A. Watters et al. take the geometry of the shock pressure field as determined by the decay law proposed by Ahrens and O’Keefe (1977), with the best-fit decay constants to the numerical calculations of Pierazzo et al. (1997), and use the waste heat or shock-Hugoniot method (Gault and Heitowit, 1963) to determine the amount of energy released during decompression of the shock phase along the Hugoniot. W.A. Watters et al. assumed that all waste heat increases the temperature of the target; this premise does not take into account the latent heat of the melted material, which consumes a substantial amount of the waste heat. Therefore, as a modification to this method, the waste heat is assumed here to contribute to an increase in temperature up to the melting point of the target material at a given pressure, after which additional temperature increase occurs only with partial melting. The points at which the temperature is equal to or exceeds the melting temperature of the target material determine the volume and maximum depth of melting. Constant and conductive crustal temperature profiles were examined, and both yielded similar results for the crater sizes of interest for this study.

The melt volumes calculated by this method are also shown in Fig. 4. The amounts of melt generated by this modified W.A. Watters et al. (2009) model for terrestrial impacts are greater than for the results of Pierazzo et al. (1997), but less than those of Cintala and Grieve (1998a). This model appears to provide the best fit to the amount of melt seen at terrestrial craters, only slightly overestimating the amount of terrestrial melt observed.

The relationship between $D_m$ and the maximum melting depth for craters on Mercury is shown in Fig. 5 using all three methods described here for vertical impacts at 42.5 km/s into dunite targets. The assumed shape of the melt volume for the modified W.A. Watters et al. (2009) method is shallower and broader than for the other two methods (where the volume is roughly spherical) due to the nature of the incorporated pre-existing solidus. When these methods are used to calculate melt depths for high velocities (>25 km/s), the modified W.A. Watters et al. method yields the shallowest values for the maximum melting depth. The spread of values across the three methods for Mercury is less than a factor of two and is low enough to provide reasonable confidence in the absolute depth values calculated here for Mercury. This range also reflects the broad range of impact velocities expected on Mercury (Le Feuvre and Wieczorek, 2008). For all calculations below we use the maximum depth of melting calculated with the modified W.A. Watters et al. method.

4. Observations

We use MESSENGER WAC multispectral data to identify occurrences of spectrally distinct material in primary association with impact craters. Material is defined as spectrally distinct if it is found to contrast with the pre-impact surface on the basis of spectral shape, band ratios, and principal component analysis. Spectral contrasts due solely to maturity effects are excluded from this study. The enhanced-color images featured in this paper contain the inverse of the second principal component (-PC2), the first principal component (PC1), and the 430-nm/1000-nm color ratio emphasizes relative slope variations. Excavated RM and LRM stand out conspicuously from the surrounding terrain in these color composites, because they are the spectral end-members in Mercury’s compositional continuum. Therefore, these units form the focus of this study.

Once areas of interest are identified, WAC data are overlaid on NAC mosaics in order to examine the regions at higher spatial resolution. Table 2 lists the location, diameter, maximum excavation depth, and minimum origin depth of central peaks (if applicable)
for each crater discussed in this study. The specific areas of interest that we have examined are described below in Sections 4.1–4.4.

4.1. Second flyby high-resolution color data

The first area that was closely examined is located in the MESSENGER second flyby departure hemisphere, within the region of Mercury that has been imaged at the highest MDIS spatial resolution to date (C24 460 m/pixel for the WAC, C24 120 m/pixel for the NAC). Fig. 6 is a C24 1150-km-wide WAC mosaic centered at 1.3°S, 313.3°E and ranging from the plains just west of Rudaki crater (the Rudaki plains) eastward to Homer basin. The reddest areas in this image represent occurrences of RM, and the darkest blue areas represent LRM. There are several occurrences of excavated, spectrally distinct materials in this area. We focus separately on the western (Rudaki plains, Fig. 7), central (Titian basin, Fig. 8), and eastern (Homer basin, Fig. 9) portions of the image in order to examine each area in greater detail.

4.1.1. Rudaki plains

The Rudaki plains (Fig. 7) have been classified as smooth volcanic IP material that has filled an unnamed, degraded, ~370-km-diameter basin (Denevi et al., 2009). This area contains a prominent example of spectrally distinct, excavated material: the 68-km-diameter complex crater Calvino, located near the center of the Rudaki plains at 3.9°S, 304°E. Calvino is superposed on the IP terrain. The crater rim is formed by RM, whereas LRM makes up a portion of its central peak structure. This region was also described by Denevi et al. (2009).

The multiple stratigraphic units exposed by Calvino constrain the thicknesses of the surface IP layer and the underlying RM and define the minimum depth of origin for the LRM material exposed in the central peaks. Small, post-flooding craters up to 16 km in diameter (e.g., crater R1 in Fig. 7) do not penetrate through the surface IP. From the maximum excavation depth of R1 (the largest non-penetrating crater), the IP layer at the center of the Rudaki plains must be at least 1.2 km thick. There are no post-flooding craters intermediate in size between R1 and Calvino in the central Rudaki plains; therefore, the maximum excavation depth of Calvino, ~4.4 km, provides the only depth constraint for the location of the RM layer. Because Calvino excavates RM, the top of this layer must lie at a depth between 1.2 km (maximum excavation depth of R1) and 4.4 km (maximum excavation depth of Calvino).

![Fig. 6. Color composite of high-resolution WAC images, showing PC2, PC1, and relative visible color (430-nm/1000-nm ratio) in the red (R), green (G), and blue (B) image planes, respectively. The scene is approximately 1150 km across and spans from the Rudaki plains in the west to Homer basin in the east. The image is in equirectangular projection and is centered at 1.3°S, 313.3°E.](image1)

![Fig. 7. Color composite of the Rudaki plains, centered on Calvino crater (68 km in diameter, 3.9°S, 304°E). The Rudaki plains are IP that have filled a degraded, ~370-km-diameter basin. Calvino, superposed on the IP, has excavated RM in its rim and LRM in a portion of its central peaks. The inset provides a higher-resolution view of Calvino crater.](image2)
The dual composition of Calvino’s central peaks implies either that the lateral distribution of LRM is uneven beneath the crater, or that the peaks are derived from a depth range containing the LRM interface. Under the latter scenario, the LRM unit begins at a depth of ~6 km, constraining the thickness of the RM layer beneath Calvino to between ~1 and 5 km. No constraint can be placed on the vertical extent of the LRM in this area.

4.1.2. Titian basin

The impactor that formed Titian basin (121 km in diameter, 3.6°S, 317.9°E, Fig. 8) struck smooth plains (Denevi et al., 2009), which exhibit IP spectral characteristics. The Titian ejecta deposit contains LRM, which conceals the spectral signature of much of the underlying plains unit (Denevi et al., 2009). From the maximum excavation depth of Titian, the ejected LRM originated from no deeper than 7.6 km. The basin is filled with younger smooth plains (Denevi et al., 2009), also of the IP spectral type, and no central peaks are visible.

Two craters (T1 and T2), 4 and 9 km in diameter, superposed on the northern floor of Titian excavated LRM from beneath the younger IP (Denevi et al., 2009). Therefore, the IP infill beneath these craters must be <0.3 km thick (maximum excavation depth of T1). A 7-km-diameter crater in the northeastern portion of Titian (T3) does not expose LRM, suggesting an uneven unfill of IP, which must be >0.6 km beneath T3.

A prominent small, 8-km-diameter crater north of Titian (T4 in Fig. 8) excavated through the LRM ejecta deposit to expose the underlying, older IP unit. This crater is analogous to lunar dark-haloed craters (e.g., Schultz and Spudis, 1979; Bell and Hawke, 1984; Head and Wilson, 1992), which excavate dark basaltic mare material from beneath higher-albedo, highland-rich ejecta deposits. The T4 case exhibits an inverted color relationship relative to the case of lunar DHCs: it exposes relatively high-reflectance, likely volcanic IP from beneath dark LRM ejecta. The maximum excavation depth of this crater limits the thickness of the LRM in this portion of the ejecta deposit to <0.6 km.

4.1.3. Homer basin

Homer basin (Fig. 9) is a ~310-km-diameter basin that has been flooded with IP (Denevi et al., 2009). Homer was classified as a two-ring basin on the basis of Mariner 10 images (e.g., Spudis and Guest, 1988; Pike, 1988). The northern and western portions of the basin’s outer rim appear to have been modified. The area in and around Homer basin contains several occurrences of the red unit in primary association with craters. There is an 11-km-diameter crater, designated H1 in Fig. 9, located near the center of Homer that does not appear to have excavated through the IP. The maximum excavation depth of H1 constrains the thickness of the IP beneath this crater to >0.8 km. There are several possible explanations for this observation, including: (1) the red material...
was covered by >0.8 km of IP; (2) the impact was oblique, making the excavation depth shallower than ~1/10 the crater diameter (Gault et al., 1968); or (3) the red unit does not extend beneath this crater or is present only at greater depth.

The craters in and around Homer basin that expose RM include the following: H2 on the floor of a ~38-km-diameter crater (maximum excavation depth ~2.6 km); H3 and H4 on the floor and in the near-crater ejecta of two ~15-km-diameter craters (maximum excavation depth ~1.1 km); and the ejecta of Dominici, a 20-km-diameter crater (maximum excavation depth ~1.4 km). The still-visible rays of Dominici indicate that it is one of the youngest (Gault et al., 1968); or (3) the red unit does not extend beneath this crater.

Just south of Homer basin in the highly cratered, intermediate terrain is a 50-km-diameter crater, H5, which exhibits a central peak spectrally consistent with the other RM in the scene (Fig. 9). The maximum excavation depth of H5 is 3.3 km, and the minimum depth of origin of this peak is ~4 km; therefore, the top of the RM layer outside of Homer basin must exist between these depths. No RM is seen in the ejecta of H5 or in nearby smaller craters. If this RM represents the same unit as that exposed inside of Homer basin, no large basin-forming impact helped to bring it to the surface, so the RM outside of Homer remained too deep to be tapped by smaller craters.

4.2. Hemingway region

The region surrounding the crater Hemingway (17.5°N, 357.1°E, imaged during MESSENGER's second flyby) is shown in a ~1000-km-wide WAC image (Fig. 11). The spatial resolution of the data in this area (WAC ~ 2400 m/pixel, NAC ~ 580 m/pixel) is approximately five times lower than that for the areas described in Section 4.1. The cyan-colored, linear features running through the area are part of the extensive ray system of the 95-km-diameter crater Hokusai, to the north and east of the scene (57.8°N, 16.8°E). Hokusai was first identified in Earth-based radar images (Slade et al., 1992; Butler et al., 1993; Harmon, 2007).

Four large craters in Fig. 11 are easily identifiable as exposing RM. Hemingway, a 130-km-diameter crater (maximum excavation depth ~8.1 km), has RM exposed on its floor. Hemingway also has LRM material exposed in its central peak, which has a minimum depth of origin of 13 km. De Graff, a 65-km-diameter crater (maximum excavation depth ~4.3 km), has a rim of RM. BCFD material is present inside the crater, and a crater ray is draped around it (Blewett et al., 2010), complicating spectral observations of its central peaks. A 43-km-diameter crater, labeled He1 in Fig. 11 (maximum excavation depth ~2.9 km), and a 30-km-diameter crater, He2 (maximum excavation depth ~2.1 km), also have RM rims.

Because it is incorporated into Homer's peak ring, the RM must have been present in the pre-basin stratigraphy. The formation of Homer would have brought this RM up from depth, exposing it at the surface at least in the area of the peak ring. The basin floor was subsequently infilled by IP, covering the northern portion of the peak ring, and leaving RM exposed only in the southern portion of the peak ring. The RM was then easily excavated by the subsequent H2–H4 and Dominici impacts, which were large enough to penetrate the IP layer. No LRM or other underlying material is observed in the craters on Homer's floor, so no maximum depth can be placed on the extent of the RM.

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Fig. 10. (a) WAC ~ 460 m/pixel color composite and (b) NAC ~ 120-m/pixel image centered on a 15-km-diameter crater (H4 from Fig. 9) containing RM at 1.6°S, 325.1°E. Arrows indicate the presence of a ridge (Homer's peak ring) that acted as an obstruction to the ejecta. Some of this blocked ejecta and material from a slumping wall were deposited onto the eastern floor of the crater.

Fig. 11. Color composite of the ~1000-km-wide region surrounding Hemingway (17.5°N, 357.1°E), which excavated RM and exhibits central peaks composed of LRM. Also containing RM in their rims are de Graft (65 km in diameter), He1 (43 km in diameter), and He2 (30 km in diameter). The incorporation of RM into the rims of craters with a wide range of diameters suggests that the underlying RM layer may be several kilometers thick.
Spectra of the central peaks of craters He1 and He2 cannot be observed due to the limited spatial resolution of the WAC images.

Because these craters with a wide range of diameters have excavated and incorporated RM into their rims, this material must be present over a large depth interval, at minimum ranging from ~2 to 8 km. If this region has a relatively uniform subsurface, the RM layer would be ~6 km thick. Smaller bright-red areas can be seen in Fig. 11; however, the resolution of the data precludes determining whether they are associated with craters or other geological structures, such as pyroclastic vents. It is possible that RM exists closer to the surface than 2 km; higher-resolution images from the MESSENGER mission orbital phase will help to settle this issue. The LRM in Hemingway’s central peak originates at a minimum depth of ~13 km, providing a limit to the maximum depth of the RM layer.

4.3. Caloris basin

At 1550 km in diameter, Caloris basin is the largest well-preserved impact basin on Mercury (Murchie et al., 2008). The red spots located along the periphery of Caloris (Fig. 12) have a secondary association with the basin. The red spots surround rimless depressions and have been interpreted as pyroclastic deposits (Murchie et al., 2008; Head et al., 2008; Kerber et al., 2009; Blewett et al., 2009). The interior of Caloris contains smooth plains that have been classified as HRP (Robinson et al., 2008; T.R. Watters et al., 2009). Several craters within the basin interior (Nawahi, Munch, Sander, and Poe) are embayed and infilled by HRP (see Fig. 12), indicating that they predate the latest emplacement of this plains material (Murchie et al., 2008). The still-visible rims of these craters provide clues to the nature of the material beneath the HRP, which is spectrally consistent with LRM (Robinson et al., 2008). If this LRM represents the original floor of Caloris basin, it was likely brought to the surface from great depth by the large basin-forming impact.

Many impacts are superposed on the smooth plains, and some of them were sufficiently large to excavate through the HRP layer and expose the underlying LRM (Murchie et al., 2008), as indicated in Fig. 12. The largest of these superposed craters is Atget (100 km in diameter, 25.7°N, 166.1°E). This crater exposed substantial amounts of LRM, both in its ejecta and in its central peaks. The maximum excavation depth of ejecta from Atget implies that LRM is present <6.3 km below the surface, and the minimum depth of origin of the central peaks indicates that LRM is present at a depth of 10 km. If these materials were derived from the same LRM layer, the subsurface unit could be at least 3.7 km in thickness.

Apollodorus (30.6°N, 163.0°E), the 41-km-diameter crater central to the Pantheon Fossae graben complex, exposes LRM in its floor, rim, and central peaks (Murchie et al., 2008), though several similarly sized craters (e.g., Cunningham, 37 km in diameter) expose LRM only in their central peaks. The maximum depth of excavation and minimum depth of central peak origin for Apollodorus are 2.8 and 4 km, respectively. The differences in excavated material between similarly sized craters suggest an uneven depth to HRP inside Caloris. The fact that some 37–40-km-diameter craters expose LRM in their central peaks but not in their floors or ejecta limits the thickness of the HRP infill near the center of Caloris to between 2.5 and 4 km. Volcanic flooding of lunar impact basins by mare deposits shows similar trends and thicknesses (e.g., De Hon, 1979; Head, 1982; Head and Wilson, 1992; Budney and Lucey, 1998; Thomson et al., 2009).

4.4. LRM centers

LRM covers at least 15% of Mercury’s surface and typically occurs as diffuse regional deposits or in concentrated centers associated with large craters and basins (Robinson et al., 2008; Blewett et al., 2009; Denevi et al., 2009). Its morphological characteristics and association with large craters and basins are consistent with excavation from depth by impacts (Denevi et al., 2009). LRM centers can be seen in several locations across the planet; however, not all large craters and basins have excavated LRM, indicating a heterogeneous distribution in the crust (Denevi et al., 2009). The large (>80-km diameter) impact structures that are associated with LRM centers have maximum depths of excavation ranging from 5 km to in excess of 22 km (or possibly exceeding 75 km in the case of Caloris, if these scaling relationships are valid for a basin of that size). Many of the LRM-excavating basins are pre-Tolstojan in age, indicating that the time of original emplacement of LRM into the subsurface crust must be >4.0 Ga.

In some areas, LRM is seen in the ejecta of smaller craters (4–40 km in diameter). Many of these occurrences are seen within basins (e.g., Caloris, Titian), where the LRM was originally brought to the surface or relatively shallow subsurface from greater depth by the basin-forming event. Even if this LRM were subsequently buried by volcanism, it thereafter resided sufficiently close to the surface to be excavated by smaller craters.

5. Discussion

We use the depth bounds calculated in Section 4 to construct schematic cross-sections depicting the overall stratigraphy of the five analyzed areas (Fig. 13). We make the simplifying assumption that subsurface structure is laterally uniform over each area considered and that only the layers observed are present. Comparing the cross-sections allows a preliminary assessment of color units and their vertical and horizontal distributions. Dotted lines represent the pre-basin position of units: the pre-Titian subsurface location of the LRM, from the maximum excavation depth of Titian;
and the pre-H5 location of the RM exposed in the central peak, unaffected by the formation of Homer basin.

The smooth plains units analyzed in this study all infill large craters or basins. The three IP units are ≤1 km thick, and the HRP unit that fills the much larger Caloris basin is at least 2.7 km thick at the center of the basin. Because these plains are believed to be volcanic in origin and the crater floors may be topographically uneven, the thickness of volcanic plains across each host crater or basin is likely to be variable. These thicknesses are consistent with those inferred by Head et al. (2009a), which ranged from many hundreds of meters to several kilometers, and are comparable to those of many lunar maria (e.g., De Hon, 1979; Head, 1982; Head and Wilson, 1992; Budney and Lucey, 1998; Thomson et al., 2009).

In three of the areas examined (Rudaki, Homer, and Hemingway), there is evidence of RM originating at depths between 1 and 4 km. The Homer and Rudaki areas are ~850 km apart on the surface; however, it is unknown whether the RM observed in the two areas is related. In the Homer and Hemingway regions, the presence of RM in the rims of craters with a range of diameters suggests that the underlying RM layer may be several kilometers thick. Radar studies of lunar pyroclastic deposits have yielded thicknesses ranging from 5 to 20 m (Zisk et al., 1974; Hawke et al., 2009). If these thicknesses are typical of large pyroclastic deposits, it is unlikely that pyroclastic deposits of several-kilometer thickness were emplaced on Mercury and subsequently buried, despite their spectral similarity to the “red spot” pyroclastic deposits elsewhere on the planet’s surface.

Of the four “red spots” identified by Blewett et al. (2009), two (contained within the craters Moody and Navoi) may have primary associations with impact craters, like the examples of RM detailed in this study. The spectra of these two occurrences of RM are similar to those of the HRP, suggesting that they may represent deposits of HRP-like material (Blewett et al., 2009). The limited spatial resolution of the current images does not permit the determination of whether these deposits represent post-crater flooding or pre-crater volcanic units that were covered or intrusive and subsequently excavated from depth by impacts. The deposits in Moody and Navoi warrant further investigation after the acquisition of higher-resolution images from the MESSENGER mission orbital phase.

The spectra of the Blewett et al. (2009) RM as well as the areas of RM described in this study are shown in Fig. 14, along with an example of HRP from the interior of Caloris basin and examples of pyroclastic deposits. Although there are slight variations among the RM spectra, they all have spectral characteristics (albedo and slope) that are similar to the HRP and distinct from the pyroclastic examples. If these subsurface RM layers do represent buried or...
intrusive volcanic material, they are evidence for large-scale deposits. If they were deposited at the surface, these layers would represent an older generation of smooth plains.

Because the observed RM units occur as small deposits that have been emplaced during subsequent impact events, it is difficult to constrain their original ages. In at least two of the main examples, RM has been excavated from beneath subsequent IP fill, limiting the RM to ages older than those of these plains. We can place additional constraints on the age of the RM in Homer, since its incorporation into the peak ring requires that it predate the basin. Homer is highly degraded and was classified as pre-Tolstojan in age from Mariner 10 mapping (Spudis and Guest, 1988), indicating that the RM in its peak ring, whether intrusive or extrusive, must be >4.0 Ga.

In the Rudaki, Titian, Hemingway, and Caloris areas examined above, LRM represents the deepest material observed. The Caloris LRM in particular is excavated by craters of many sizes, indicating that it is present over a depth range of several kilometers. These observations suggest some possible scenarios: (a) any non-LRM layers intermixed with the subsurface LRM are very thin and are not detectable at the resolutions available; (b) non-LRM layers are present but exist far enough below LRM deposits that they were not tapped by impacts; or (c) non-LRM layers are not present in these stratigraphic columns. If this last option is true, it could mean the LRM represents basement material uncovered by basin-forming impacts. The higher-resolution images acquired during the MESSENGER orbital mission phase will permit the detailed search for spectrally distinct craters in LRM centers and a closer examination of craters exposing LRM in order to further investigate these scenarios.

The derived stratigraphies imply that large parts of Mercury's surface have been built up over time by a series of volcanic flows, in agreement with several recent studies (e.g., Head et al., 2008, 2009b; Murchie et al., 2008; Denevi et al., 2009). In some areas, accumulations of volcanic material reach thicknesses of several kilometers, at least to a depth of ~5 km. The subsurface RM may represent an older generation of HRP-like smooth plains, at least some of which are pre-Tolstojan in age. The craters uncovering RM might, therefore, expose some of the earliest volcanism on Mercury, analogous to lunar cryptomare material, ancient mare basalts that are hidden beneath highland-rich ejecta (e.g., Head and Wilson, 1992; Blewett et al., 1995; Antonenkov et al., 1995). In this way, these craters are analogous to the Moon's dark-halo craters (e.g., Schultz and Spudis, 1979; Hawke and Bell, 1981; Bell and Hawke, 1984), although most lunar DHCs are smaller than these examples on Mercury (Head et al., 2009a). The size difference may be due to the difference in the overlying material (ejecta deposit for the DHCs, volcanic plains for the RM), but the MESSENGER flyby images are not of sufficient resolution or coverage to rule out occurrences on the scale of lunar DHCs. Another analogy to lunar DHCs is the previously mentioned T4 crater near Titian, which uncovers older IP from beneath an LRM ejecta deposit, producing a bright-halo crater. This is a more direct analogy, since the older volcanic material is excavated from beneath ejecta, not later volcanic flows.

Knowledge of Mercury's crustal stratigraphy and volcanic history is important for interpreting several types of geological and geophysical observations that will be made during MESSENGER's orbital phase. On the most basic level, understanding the thicknesses of individual volcanic units (both surface and buried) helps to define the extent and role of volcanism in Mercury's crustal evolution, which in turn relates to the planet's thermal history. Additionally, these thicknesses will be valuable inputs to lithospheric loading models designed to test ideas for the current topography within major basins and for the formation of tectonic features. Older generations of volcanism can be catalogued and examined, in order to search for temporal changes in magma composition (for example, are these ancient volcanic materials all RM in composition, or do they also span the range of compositions seen on the surface of Mercury?).

If, as proposed by Robinson and Lucey (1997) and Denevi et al. (2009), LRM has a relatively high proportion of opaque mineral phases such as Fe- or Ti-bearing oxides, it will be denser than the average Mercury crustal material. Therefore, the LRM centers could be locations of positive gravity anomalies. Likewise, an unexplained positive gravity anomaly, for example one not correlated with an impact basin, as with lunar mascons (e.g., Muller and Sjogren, 1968), might be a candidate location for an unexposed subsurface LRM deposit. If the LRM darkening agent is capable of retaining remanent magnetization, buried LRM bodies, particularly if emplaced over limited timescales, may also produce measurable crustal magnetic anomalies.

6. Conclusions

The craters of Mercury provide critical windows into the subsurface, through which the stratigraphy of the upper crust can be explored. Craters excavating material that is spectrally distinct from the surrounding surface provide a means to bound the depth of origin of the spectrally distinct ejecta and central peak structures. The resulting stratigraphic knowledge provides insights into the origin and evolution of Mercury's crust.

The examination of several small areas on Mercury reveals the complex nature of the local stratigraphy. The surface smooth plains deposits examined range from less than 1 km to ~2.7 km in thickness and are likely volcanic in origin. Examples of RM, some of which may be several kilometers in thickness, were found to be spectrally similar to the HRP, suggesting that they may represent an older generation of volcanic plains that were subsequently covered by intermediate plains. Some of the RM and LRM deposits were excavated by pre-Tolstojan basins, indicating relatively old crustal emplacement ages (>4.0 Ga). Exposures of RM by impacts may be uncovering Mercury's equivalent to lunar cryptomaria, although most of these Mercury instances are excavated from beneath subsequent volcanic flows rather than basin ejecta. All of these observations highlight the importance of volcanism on Mercury and support the origin of large regions of the upper crust by the buildup of volcanic units (e.g., Denevi et al., 2009), at least to a depth of ~5 km. This conclusion points to Mercury as being characterized by a volcanic crustal component that is more substantial in thickness, volume, and areal distribution than the mare basalt phase of crustal formation on the Moon (e.g., Head and Wilson, 1992).

The five areas examined in this study provide a preliminary view of Mercury's crustal stratigraphy. Further refinements are contingent on global high-resolution MDIS coverage of the planet, along with targeted spectra and topography, which will be acquired during MESSENGER's orbital mission phase beginning in 2011. Improvements in coverage and resolution will directly enhance the level of detail to which we can determine Mercury's crustal stratigraphy.

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